

Combined statistical and dynamical assessment of simulated vegetation–rainfall interactions in North Africa during the mid-Holocene*

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Abstract

A negative feedback of vegetation cover on subsequent annual precipitation is simulated for the mid-Holocene over North Africa using a fully coupled general circulation model with dynamic vegetation, FOAM-LPJ (Fast Ocean Atmosphere Model-Lund Potsdam Jena Model). By computing a vegetation feedback parameter based on lagged autocovariances, the simulated impact of North African vegetation on precipitation is statistically quantified. The feedback is also dynamically assessed through initial value ensemble experiments, in which North African grass cover is initially reduced and the climatic response analyzed. The statistical and dynamical assessments of the negative vegetation feedback agree in sign and relative magnitude for FOAM-LPJ. The negative feedback on annual precipitation largely results from a competition between bare soil evaporation and plant transpiration, with increases in the former outweighing reductions in the latter given reduced grass cover. This negative feedback weakens and eventually reverses sign over time during a transient simulation from the mid-Holocene to present. A similar, but weaker, negative feedback is identified in Community Climate System Model Version 2 (CCSM2) over North Africa for the mid-Holocene.

Keywords: climate model, evapotranspiration, grassland, mid-Holocene, monsoon, North Africa, precipitation, soil, vegetation feedback, vegetation model

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Introduction

Based on multiple paleo-records, North Africa was characterized by expanded savanna and steppe and abundant lakes during the mid-Holocene (Hoelzmann *et al.*, 1998; Jolly *et al.*, 1998; Prentice *et al.*, 2000), due to an intensified monsoon under increased Northern Hemispheric summer insolation (Kutzbach & Otto-Bliesner, 1982; Kutzbach & Street-Perrott, 1985). This expanded vegetation likely induced feedbacks on climate (Kutzbach *et al.*, 1996; Broström *et al.*, 1998; Doherty *et al.*, 2000; Levis *et al.*, 2004), although the specifics of these feedbacks are poorly understood. Vegetation generally impacts the atmosphere through fluxes of heat, moisture, and momentum, which are, in turn,

driven by variations in surface albedo, evapotranspiration, and surface roughness, respectively.

Charney (1975) introduced the concept of a positive vegetation feedback on rainfall regarding North African desertification, suggesting that a reduction in vegetation produces an increase in surface albedo, leading to increased atmospheric stability, sinking motion, and drying. Charney's theory is based upon the large contrast in surface albedo between vegetated surfaces (0.14) and bare ground (0.35) over North Africa. Using Meteosat satellite measurements, Ba *et al.* (2001) estimates that surface albedo ranges from 0.15–0.2 over the Congo rainforests to 0.3–0.45 over the Sahara Desert. Using MODIS data over North Africa and the Arabian Peninsula, Tsvetsinskaya *et al.* (2006) reported large fluctuations in 300–5000 nm broadband albedos ranging from 0.22 for luvisols (nutrient-rich soil) to 0.43 for dunes and shifting sand, but in general agreement with the Meteosat values.

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Paleo-Modeling Intercomparison Project (PMIP) (Joussaume & Taylor, 1995; Joussaume *et al.*, 1999; Braconnot, 2000) simulations of the mid-Holocene generally failed to produce a sufficient northward advance in North African monsoon rainfall. Hales *et al.* (2006) attribute this deficiency to the lack of critical vegetation and ocean ventilation feedbacks and the use of modern-day surface albedos for the Sahara in many PMIP models. The Saharan albedo during the mid-Holocene was likely much lower than modern values due to wetter soils and the presence of steppe vegetation. Small & Kurc (2001) found that observed surface albedo is lower under wet conditions in semiarid environments (Govaerts & Lattanzio, 2007). Tetzlaff (1983) assessed an average Saharan albedo of 0.32 for the present day. Applying known modern albedos for different surface types to the conditions that presumably existed in the mid-Holocene, Tetzlaff (1983) suggested that the Saharan albedo was close to 0.27 around 5500 years ago. The artificially high surface albedos used in many PMIP experiments resulted in overly dry Saharan conditions and limited the poleward advance of the intertropical convergence zone. Wetter conditions over North Africa during the mid-Holocene would have reduced the albedo contrast between soil and vegetation, bringing the validity of Charney's theory in question for the mid-Holocene, and hinting at a greater role from evapotranspiration.

Numerous paleodata records offer insight into the transient change in North African climate and ecosystems between the mid-Holocene and present. Lake levels declined from 6K to 0K (Hoelzmann *et al.*, 1998, 2004) as rainfall diminished across North Africa. An abrupt increase in terrigenous dust flux over the Atlantic Ocean off North Africa, at Ocean Drilling Program (ODP) core 658C (de Menocal *et al.*, 2000; Adkins *et al.*, 2006), hints at an abrupt collapse in Northwest African vegetation around 6K and a greater dust source. These records suggest a more abrupt collapse in vegetation than rainfall following the African Humid Period (Liu *et al.*, 2007a).

An array of modeling studies has examined the transient change in climate and vegetation since the mid-Holocene and reached widely varying conclusions using simplified climate–ecosystem models. Claussen *et al.* (1999) simulated an abrupt collapse in Northwest African vegetation using a climate–vegetation model of intermediate complexity and attributed the collapse to a strong positive vegetation feedback on rainfall. Other studies failed to produce an abrupt collapse (Renssen *et al.*, 2003; Wang *et al.*, 2005), even with the same vegetation model as Claussen. Renssen *et al.* (2003, 2006) did not simulate an abrupt vegetation change due to high precipitation variability in their atmo-

spheric model. Numerous studies have suggested that positive vegetation feedbacks amplified the response of the monsoon to orbital forcing changes (Street-Perrott *et al.*, 1990; Kutzbach *et al.*, 1996; Claussen, 1997, 1998; Claussen & Gayler, 1997; Texier *et al.*, 1997, 2000; Broström *et al.*, 1998; Pollard *et al.*, 1998; Braconnot *et al.*, 1999; Doherty *et al.*, 2000; Foley *et al.*, 2003). The enhanced vegetation cover over the Western Sahara as simulated by Claussen & Gayler (1997), Brovkin *et al.* (1998), and Renssen *et al.* (2003) is in close proximity to the Atlantic marine core ODP 865C.

Liu *et al.* (2007a) produced the first synchronously coupled transient simulation of the evolution of North African climate and ecosystems for the past 6500 years using a global general circulation atmosphere–ocean–terrestrial ecosystem model, ANL/UW FOAM-LPJ (Argonne National Laboratory/University of Wisconsin-Madison Fast Ocean Atmosphere Model-Lund Potsdam Jena Model). In agreement with paleo-records, the transient simulation, driven by gradual changes in orbital forcing, includes an intensified North African monsoon and poleward expanded vegetation at 6K, a gradual decline in rainfall over time, and an abrupt vegetation collapse around 5.5K (Liu *et al.*, 2007a). The cause of the abrupt vegetation collapse is explored by Liu *et al.* (2006a). Over eastern North Africa ('region A': 18–23°N, 11–36°E), simulated annual precipitation decreases by at least 80%, supporting a southward retreat in grasslands (Fig. 1). The band of grasslands narrows over time. Within the North African zone of 18–22°N, the number of grid cells with a grass cover fraction (fraction of grid cell covered by C3 and C4 grasses) of at least 0.1 reduces from 67 to 39 in the transient simulation. Over region A, annual precipitation gradually declines from 1.2 to 0.1 mm day⁻¹, while grass cover fraction shows an abrupt collapse around 5.5K, decreasing from 0.6 at 6K to near 0 at 3–0K (Fig. 2).

The East African monsoon (Camberlin, 1997; Block & Rajagopalan, 2007), while less distinct and studied as its West African counterpart, is a critical source of rainfall to countries such as Ethiopia, Sudan, and Eritrea. The rainy season in northern summer is known as 'Kiremt' in Ethiopia. The ITCZ over Africa results from convergence between moisture southwesterlies from the Atlantic and dry northeasterlies off the Sahara and Arabian Peninsula. As the land is heated in summer, the ITCZ shifts northward and triggers the East African monsoon. Moisture flux into the East African monsoon region results from the northward shift of the ITCZ (Griffiths, 1972; Gamachu, 1977), key circulation features (Arabian and Sudan thermal lows and South Atlantic and Indian Ocean highs), and moisture advection with southwesterly airflow across the Congo (Block & Rajagopalan, 2007). We, therefore, refer to 'region A'

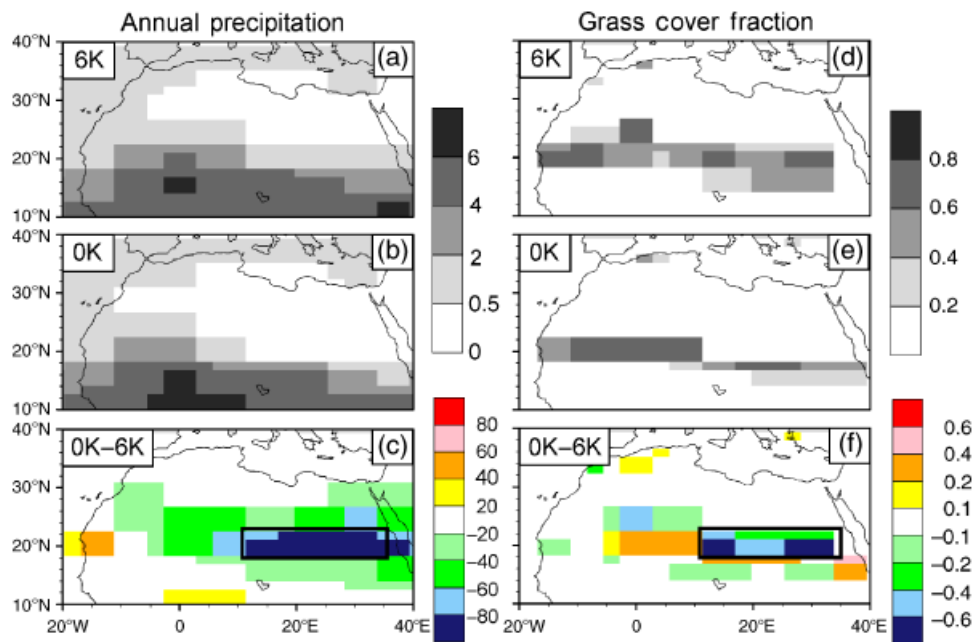


Fig. 1 Annual mean precipitation (mm day^{-1}) over North Africa for (a) 6500–6000K and (b) 500–0K in the transient FOAM-LPJ simulation. (c) Percent difference in annual mean precipitation (0–6K). Mean grass cover fraction for (d) 6500–6000K and (e) 500–0K. (f) Difference in mean grass cover fraction (0–6K). The largest percent reduction in precipitation and absolute reduction in grass cover occurs over eastern North Africa in 'region A' ($18\text{--}23^\circ\text{N}$, $11\text{--}36^\circ\text{E}$), shown in the black box. The grey shade legend applies to (a, b, d, e) and the color legend applies to (c, f).

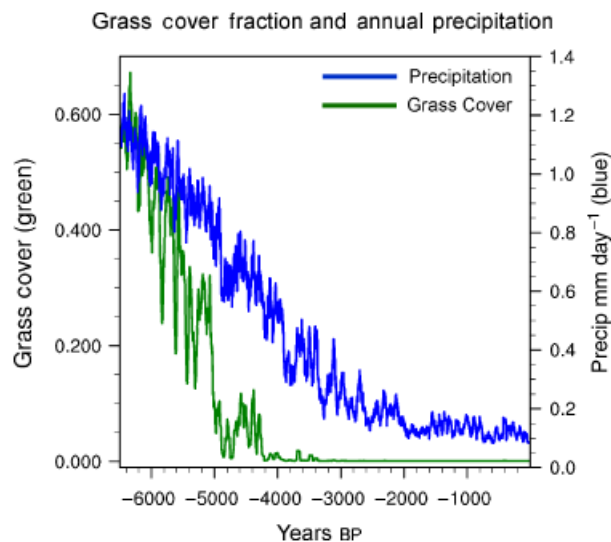


Fig. 2 Time series of annual precipitation (mm day^{-1}) (blue) and grass cover fraction (green), from 6.5 to 0K, for Region A ($18\text{--}23^\circ\text{N}$, $11\text{--}36^\circ\text{E}$) from the transient 6500-year FOAM-LPJ simulation. A 51-year running mean is applied to the data. Despite a gradual decline in rainfall, vegetation abruptly collapses around 5–6K.

as a monsoon region during the mid-Holocene, likely further north than the modern day East African monsoon due to the greater inland push of the monsoon/ITCZ from enhanced solar insolation. FOAM-LPJ suc-

cessfully simulates the East African monsoon, including convergence between the moist southwesterlies from the Atlantic and dry northeasterlies and a distinct wind shift from northerlies during the dry season to southwesterlies during the wet season, with the ITCZ reaching its northernmost position in August–September. FOAM-LPJ produces an intensified East African monsoon during the mid-Holocene (Liu *et al.*, 2007a), with enhanced southwesterlies during the wet season.

FOAM-LPJ lacks a strong positive vegetation feedback on annual North African rainfall at 6K and yet produces a vegetation collapse (Liu *et al.*, 2006a, 2007a; Wang *et al.*, 2007). Using a conceptual climate–ecosystem model, Liu *et al.* (2006a) showed that an abrupt ('stable') collapse is possible, as with FOAM-LPJ, in the absence of a strong positive vegetation feedback given strong low-frequency climate variability. Previously, Wang (2004), using a conceptual model, demonstrated that biosphere–atmosphere interactions can result in multiple equilibrium states and moderate precipitation variability can support an abrupt collapse of vegetation. FOAM-LPJ produces a weakly positive vegetation feedback on rainfall at the monthly time scale, based on monthly fraction of photosynthetically active radiation (FPAR) vs. monthly precipitation (Wang *et al.*, 2007). However, when vegetation (or grass) cover fraction is compared with subsequent annual precipitation, a

negative vegetation feedback is identified (Wang *et al.*, 2007). An increase in precipitation enhances vegetation cover, while an increase in vegetation cover leads to diminished precipitation, thereby comprising a negative vegetation feedback. This is consistent with the observational study by Wang *et al.* (2006), using Granger causality, which concluded that vegetation feedbacks vary in sign at different time scales. We will focus on the simulated negative feedback over North Africa at 6K.

Several recent studies have explored the potential for negative feedbacks on rainfall. Liu *et al.* (2006b) and Notaro *et al.* (2006) statistically quantified observed local vegetation feedbacks using satellite-derived FPAR and monthly climate data. While they found that the boreal forests impose a strong positive feedback on temperature, they determined that vegetation feedbacks on precipitation are relatively weak and vary in sign regionally. These observational studies suggest that vegetation feedbacks on precipitation are more complex than previously thought. It is possible that these feedbacks are primarily nonlocal, with evaporated water transported by the atmospheric circulation, where it precipitates over another region. Using Granger causality to examine observed vegetation–climate interactions over North American grasslands, Wang *et al.* (2006) found that anomalously high normalized difference vegetation index (NDVI) early in the growing season rapidly depletes soil water, leading to an anomalous reduction in summertime rainfall.

In the first mid-Holocene simulation using a dynamic global vegetation model (DGVM) synchronously coupled to an atmospheric general circulation model (AGCM) with a slab ocean, Doherty *et al.* (2000) found positive vegetation feedbacks on rainfall over western North Africa and negative feedbacks over central/eastern North Africa. Using CCM3-LSM (land surface model), Hoffmann & Jackson (2000) converted modern-day North African tropical savannas to grasslands and failed to simulate a significant precipitation decline, while producing a net increase in moisture convergence. In a mid-Holocene simulation using NCAR's Community Climate System Model Version 2 (CCSM2), Levis *et al.* (2004) simulated a local enhancement of precipitation due to the northward advance in African grasslands, but a regional suppression of precipitation due to the insufficient decrease in albedo with expanding vegetation. They concluded that albedo, related to soil characteristics, is the key variable in producing a positive feedback. Ensemble simulations of NCAR's Community Atmosphere Model–Community Land Model (CAM–CLM) performed by Kim & Wang (2007) revealed that, during the summer following an imposed springtime soil moisture anomaly over North

America, the vegetation feedback is negative, suppressing the response of precipitation to the soil moisture anomaly. During the wet spring, vegetation is enhanced, leading to greater transpiration and soil water loss; therefore, the vegetation feedback weakens the impact of the initial springtime soil moisture anomaly.

Vegetation feedbacks are difficult to study, due to their complex nature. Previous feedback studies have either focused purely on model simulations (Bonan *et al.*, 1992; Snyder *et al.*, 2004; Wang *et al.*, 2004; Kim & Wang, 2007; Notaro & Liu, 2007) or statistical analyses (Kaufmann *et al.*, 2003; Liu *et al.*, 2006b; Los *et al.*, 2006; Notaro *et al.*, 2006; Wang *et al.*, 2006). Through explicit experiments, the modeling approach leads to a physically based understanding of feedbacks. However, since it is based on producing sets of model experiments, it cannot be directly applied to assess feedbacks in a single control simulation or the observational record, which has only one realization. The statistical approach applies to a single realization, either a model simulation or observational record. Each statistical method has intrinsic limitations, and therefore it is unclear if the statistical method is accurately assessing the feedback; also, the statistical approach offers limited insight into physical mechanisms.

Here, we take a comprehensive approach with a combined statistical–dynamical assessment in a climate–terrestrial ecosystem model. First, we apply a statistical method to assess the simulated negative vegetation feedback on annual North African rainfall in 6K control experiments from two fully coupled GCMs, FOAM-LPJ and CCSM2. Next, we compare this statistical assessment with a dynamical assessment using explicit model simulations. Consistency between the two independent methods provides a strong indication of an accurate feedback assessment. Furthermore, the dynamical approach offers insight into the mechanisms behind the assessed feedbacks. This approach has been applied successfully in recent studies of ocean–atmosphere feedback (Liu & Wu, 2004; Liu *et al.*, 2007b).

Materials and methods

Statistical method

We apply a statistical approach to quantify vegetation feedbacks, based on the theory of Frankignoul & Hasselmann (1977) used to examine ocean–atmosphere interactions. This methodology was previously used to study sea-surface temperature (SST) feedbacks on air–sea heat flux and atmospheric circulation (Frankignoul *et al.*, 1998; Czaja & Frankignoul, 2002; Frankignoul & Kestenare, 2002; Liu & Wu, 2004). Later, Notaro *et al.* (2006) and Liu *et al.* (2006b)

applied this approach to quantify observed vegetation feedbacks, over the United States and globally, respectively. Similar to SSTs, FPAR exhibits a significant multi-month memory and can impose feedbacks on the atmosphere, whose natural internal time scale is much shorter.

The atmospheric variable can be divided into two components:

$$A(t + dt_a) = \lambda_A V(t) + N(t + dt_a),$$

where $A(t)$ is the atmospheric variable (e.g. precipitation) and $V(t)$ is a vegetation variable at time t (e.g. vegetation cover), λ_A is the feedback parameter, dt_a is the atmospheric response time (≈ 1 week), and $N(t)$ is the climate noise generated internally by atmospheric processes that are independent of vegetation variability. The atmospheric variable is determined by $\lambda_A V(t)$, which is the feedback response to a vegetation change, and $N(t + dt_a)$, which is atmospheric noise.

The noise term is eliminated by multiplying both sides of the equation by $V(t - \tau)$ and taking the covariance. τ is defined as the lead-time (vegetation leads the atmospheric variable by time τ), longer than the decorrelation time of the atmospheric noise forcing.

$$\langle V(t - \tau)A(t + dt_a) \rangle = \lambda_A \langle V(t - \tau)V(t) \rangle + \langle V(t - \tau)N(t + dt_a) \rangle.$$

Assuming that earlier vegetation does not impact later noise, and noise cannot impact earlier vegetation, the respective covariance (last term) is approximately equal to zero. Because the atmospheric response time is short (typically < 1 week) and the datasets to be analyzed are either monthly or annual, we can neglect dt_a . Therefore, when $\tau \gg dt_a$, the equation for the vegetation feedback parameter becomes

$$\lambda = \frac{\langle V(t - \tau)A(t) \rangle}{\langle V(t - \tau)V(t) \rangle}.$$

The feedback parameter is estimated as the ratio of the lagged covariance between A and V to the lagged covariance of V . Following Frankignoul *et al.* (1998), it is computed as the average from the first few time lags (e.g. $\tau = 1$ –4 years). The feedback parameter quantifies the instantaneous, local response of the atmosphere to changes in vegetation. To estimate the statistical significance of feedback parameters, the Monte Carlo bootstrap approach is applied, in which 500 individual λ 's are computed from shuffled series (Czaja & Frankignoul, 2002). The significance is determined by the percentage of these λ 's that are smaller in magnitude than the actual computed feedback parameter. Unlike Friedlingstein *et al.* (2006), who quantifies two-way feedbacks in the form of a gain factor, we apply a

statistical methodology, based on Frankignoul *et al.* (1998), that quantifies the one-way sensitivity of precipitation to changes in vegetation cover.

Vegetation feedbacks are statistically assessed in three FOAM-LPJ experiments, with $[\text{CO}_2] = 280$ ppmv. The first run is a 400-year mid-Holocene control experiment, using 6K orbital forcings. The second run is a 400-year preindustrial (0K) control experiment, using 0K orbital forcings. The third simulation is a 6500-year transient simulation, driven by changes in orbital forcing from the mid-Holocene (6.5K) to present (0K) (Liu *et al.*, 2007a). North African feedbacks are also assessed in 6K and 0K simulations of NCAR's CCSM2 (Kiehl & Gent, 2004; Levis *et al.*, 2004), both 200 years in duration and using a horizontal resolution of T31. LPJ provides the vegetation dynamics to FOAM, while CCSM2 uses LPJ for annual vegetation processes and CLM for sub-annual processes. CCSM2 uses 10 soil layers, as opposed to two in FOAM-LPJ. Gallimore *et al.* (2007) presented a detailed comparison of FOAM-LPJ and CCSM2.

Dynamical method

Experiments are performed using the fully coupled global atmosphere–ocean–land model, FOAM-LPJ, with dynamic vegetation (Gallimore *et al.*, 2005; Notaro *et al.*, 2005; Liu *et al.*, 2007a). The coupled atmospheric–oceanic component is FOAM version 1.5 (Jacob, 1997). The atmospheric component, PCCM3-UW (Drake *et al.*, 1995; Jacob, 1997), uses a horizontal resolution of R15 and 18 levels in the vertical, while the z-coordinate oceanic component, OM3, uses a horizontal resolution of 1.4° latitude $\times$ 2.8° longitude and 32 levels in the vertical. FOAM produces a steady long-term climate without using flux adjustment and captures most of the major observed climate features as in state-of-the-art climate models (Jacob, 1997; Liu *et al.*, 2003). It produces reasonable climate variability, including El Niño–Southern Oscillation (Liu *et al.*, 2000; Liu & Wu, 2004), Pacific decadal variability (Wu & Liu, 2003; Wu *et al.*, 2003), and tropical Atlantic variability (Wu & Liu, 2002).

FOAM is synchronously coupled to a modified version of LPJ-DGVM (Sitch, 2000; Cramer *et al.*, 2001; McGuire *et al.*, 2001; Sitch *et al.*, 2003). The land grid has the same horizontal resolution as the ocean grid, 1.4° latitude $\times$ 2.8° longitude. Nine plant functional types (PFTs) are simulated, including seven trees and two grasses. Vegetation cover fraction in LPJ refers to the total fractional coverage of these nine PFTs in a grid cell, or the sum of forest cover and grass cover fractions. The leaf area index (LAI) of a PFT individual, or lai_ind,

is computed by

$$\text{lai_ind} = \text{lm_ind} \times \text{sla}/\text{crownarea}$$

where lm_ind is the individual's leaf carbon mass (gC), sla is a constant PFT-specific leaf area, and crownarea is the individual's crown area (m^2). The individual's LAI is used to determine the foliar projective cover of the PFT individual, or fpc_ind :

$$\text{fpc_ind} = 1 - \exp(-0.5 \times \text{lai_ind}).$$

The formula for the foliar projective cover of a entire grid cell, or fpc_grid , is

$$\text{fpc_grid} = (\text{crownarea} \times \text{nind}) \times \text{fpc_ind}$$

where nind is the individual density (per m^2) (Sitch *et al.*, 2003).

No relaxation of climate forcing towards observations is applied to adjust the simulated vegetation in FOAM-LPJ. Simulated vegetation patterns are in reasonable agreement with the expected potential natural vegetation distribution (Gallimore *et al.*, 2005; Notaro *et al.*, 2005).

Simulated vegetation processes in LPJ include competition for resources, biomass allocation, establishment, mortality, soil and litter biogeochemistry, natural fire disturbances, and successional vegetation changes (Gallimore *et al.*, 2005; Notaro *et al.*, 2005). Vegetation processes within FOAM-LPJ occur along a spectrum of time scales, including sub-daily (e.g. evapotranspiration, albedo, soil temperature), daily (e.g. leaf phenology, soil water, canopy conductance, photosynthesis, respiration), and annual (e.g. establishment, reproduction, mortality, fire, competition) (Fig. 1 of Notaro *et al.*, 2005 and Gallimore *et al.*, 2005). The fractional cover of each PFT is updated annually based on establishment and mortality (Sitch *et al.*, 2003). Albedo within a grid cell is assessed as a weighted average of the albedo of the bare soil and vegetated surface. The bare soil albedo is a function of the upper soil water content, ranging roughly from 0.08 under saturated conditions to 0.27 under dry conditions. The vegetation albedo considers the albedo of both the live leaf cover and the litter/stems.

Based on the BIOME3 model (Haxeltine & Prentice, 1996), the soil hydrology in LPJ is represented semiempirically by two soil layers, with depths of 0.5 and 1.0 m for the top and bottom soil layers, respectively. Soil water is updated daily based on rainfall, evapotranspiration, percolation, and snowmelt. Daily percolation from the upper soil layer to the lower soil layer is determined based on the empirical formula provided by Neilson (1995), in which percolation is a function of the upper layer's volumetric water content squared. Surface runoff and drainage result when water content

exceeds field capacity in the upper and lower soil layers, respectively. The different rooting depths among PFTs, including a greater fraction of roots in the lower soil layer for woody PFTs than herbaceous plants, allow for competition for water resources. Raingreen PFTs maintain full leaf cover as long as the daily water stress factor (based on total soil water content) remains above the threshold of 0.35, while leaf senescence results when the water stress factor drops below 0.35. Sitch *et al.* (2003) demonstrated that LPJ-simulated vegetation patterns, runoff, carbon pools and fluxes, and the seasonal cycle of soil moisture compare favorably with observations.

A 400-year 6K (mid-Holocene) control simulation is created with FOAM-LPJ, based on 6K orbital forcings (precession, tilt, eccentricity) and $[\text{CO}_2] = 280$ ppmv. Restart files are saved every four years throughout the simulation. The area of interest extends across 18–23°N, 11–36°E (region A), over which grasslands expanded due to an intensified monsoon during the mid-Holocene in FOAM-LPJ. Using the restart files from the control simulation and applying four types of anomalies to region A alone, four sets of 100-member ensembles are created, consisting of 4-year simulations (Table 1).

In the first ensemble set (F0.2), an initial reduction in grass cover fraction of 0.2 is applied to each of the control experiment's 100 restart files and each independent experiment is run for 4 years. Grass cover fraction is updated annually in LPJ and consists of the fraction of a grid cell covered by both C3 and C4 grasses. The standard deviation in annual grass cover fraction over region A is 0.2, so the imposed grass cover anomaly in F0.2 represents the region's typical natural variability. The grass cover anomaly is applied initially, but the grass is allowed to grow back throughout each 4-year simulation. In the second set (F0.2fix), the grass cover fraction is initially reduced by 0.2 and then grass coverage is held fixed throughout the 4-year simulations, so vegetation can not grow back. In the third set (F0.2B), the grass cover fraction is initially reduced by 0.2 and a 20% reduction in grass biomass is applied initially,

Table 1 Summary of the four ensemble sets

	F0.2	F0.2fix	F0.2B	F0.2S
Grass cover fraction reduced by 0.2	X	X	X	X
Initial or fixed grass cover anomaly?	Initial	Fixed	Initial	Initial
Initial biomass reduction			X	
Initial lower soil water fraction reduction				X

including grass leaf biomass and grass root biomass. Finally, in the fourth ensemble set (F0.2S), the grass fraction is initially reduced by 0.2 and the lower soil water fraction is likewise initially reduced. The magnitude of the soil water reduction is determined by linear regression between lower soil water fraction and grass cover fraction in the 6K control simulation; the lower soil water fraction is initially reduced by 0.12 to be consistent with the 0.2 reduction in grass cover fraction. FOAM-LPJ uses two soil layers, an upper 0.5-m-thick layer and a lower 1.0-m thick layer; 90% of the grass roots occupy the top layer and 10% occupy the lower layer. The climate of the ensemble members are compared with the 6K control to assess the climatic response to imposed land surface anomalies.

While the statistical method is applied to both FOAM-LPJ and CCSM2 simulations, the dynamical method is only applied to FOAM-LPJ.

Results and discussion

Statistical assessment: FOAM-LPJ

We will focus on 'region A' of North Africa, which exhibits an abrupt vegetation collapse and gradual rainfall decline in the transient 6.5–0K simulation of FOAM-LPJ. In the 6K control simulation of FOAM-LPJ, region A receives a mean rainfall of 1.1 mm day^{-1} (40 cm yr^{-1}), supporting a vegetation cover fraction of 0.55 (Fig. 1). The majority of this vegetation cover, 0.44, is grass. Monsoon rainfalls during August–September comprise 77% of the annual precipitation, while the dry season of November–April averages $<1 \text{ cm month}^{-1}$. Annual mean FPAR for region A is just 0.18, with only August–October exceeding 0.2 and peaking at 0.47 in September. Increased summer insolation during the mid-Holocene supports a stronger monsoon and the poleward expansion of North African vegetation into region A. The slow change in orbital forcing leads to a gradual decline in rainfall over region A (80–90% reduction by 0K) and an abrupt vegetation collapse between 6 and 5K, leading to minimal cover by 4K.

The sign of vegetation's forcing on region A's rainfall varies depending on time scale. Figure 3 shows the autocorrelation of, and lead-lag correlations between, monthly anomalies of FPAR and precipitation over region A in the 6K control experiment. FPAR, which is the FPAR absorbed by the plant canopy or a measure of vegetation activity, is computed using the following formula (Sitch, 2000):

$$\text{FPAR} = \sum_{i=1}^9 \text{FPC}_i \times \text{DPHEN}_i$$

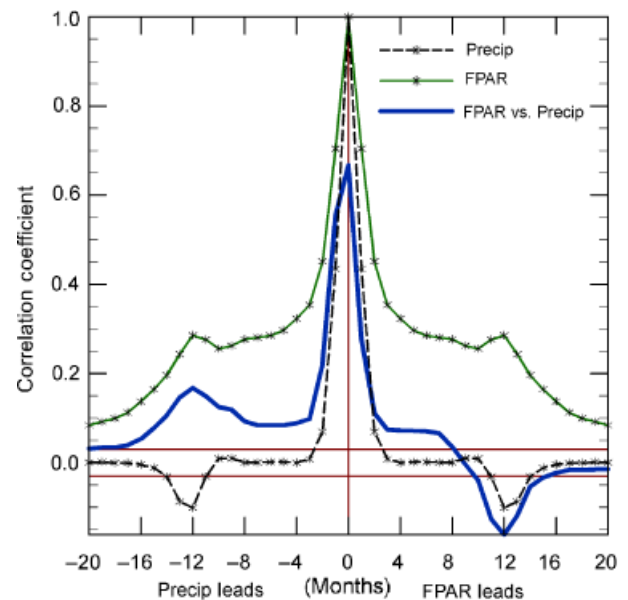


Fig. 3 Lead/lag correlations of monthly fraction of photosynthetically active radiation (FPAR) anomalies and monthly precipitation anomalies, over region A ($18\text{--}23^\circ\text{N}$, $11\text{--}36^\circ\text{E}$) in the 400-year FOAM-LPJ control simulation for 6K. The autocorrelation of fraction of photosynthetically active radiation (FPAR) is shown in the thin green line with asterisks. The autocorrelation of precipitation is shown in the dashed black line. The blue line shows a lead/lag correlation between monthly anomalies of FPAR and precipitation, with FPAR leading on the right side and precipitation leading on the left side. The horizontal red line indicates statistically significant correlations at $P < 0.1$. The memory for precipitation is 2 months and for FPAR is >20 months. Precipitation positively forces FPAR, while FPAR imposes a positive forcing on precipitation on the time scale of several months lead but a negative forcing around 1-year lead time.

where FPC_i is the annual fractional vegetation cover and DPHEN_i is the monthly fractional leaf cover (phenology) of the i th PFT. The exponential relationship between FPAR and LAI is given by

$$\text{FPAR} = 1 - \exp(-0.5 \times \text{LAI}).$$

Observed FPAR is derived from NDVI, but in the case of vegetation models, it is typically computed as a function of vegetation cover and leaf cover, or of total leaf area, thereby measuring the amount of leaf area available for photosynthesis. In FOAM-LPJ, FPAR over region A exhibits a significant ($P < 0.1$) memory (autocorrelation) exceeding 20 months. The peak in FPAR autocorrelation at 12 months reflects the annual time scale of updates to grass cover fraction in LPJ; a wet year results in an increase in vegetation cover when updated at the end of the year, resulting in an elevated vegetation cover throughout the subsequent year that

can enhance FPAR (since vegetation cover fraction is one of the component of FPAR). Unlike vegetation, rainfall memory is limited to 2 months in the 6K FOAM-LPJ control simulation. The atmosphere is largely stochastic with a memory of less than a week, so this 2-month memory is likely explained by coupling with soil water, vegetation, and SSTs (not shown). Precipitation during the preceding 16 months positively forces vegetation, with the strongest rainfall forcing occurring instantaneously or in the preceding month. FPAR is positively correlated with subsequent rainfall during the following 1–8 months, suggesting a positive feedback on shorter, subannual time scales. This correlation becomes negative when FPAR leads rainfall by 12 months, revealing a negative feedback on the annual time scale. This negative feedback will be explored both statistically, through the vegetation feedback parameter, and dynamically, using ensemble experiments.

Correlations between grass cover fraction and subsequent annual precipitation in the 6K control experiment reveal that region A is characterized by the strongest negative correlation ($P < 0.1$) (Fig. 4). These results are similar if total vegetation cover, rather than grass cover, is correlated with subsequent annual precipitation, since region A is largely grassland. The correlation between region A's vegetation cover and subsequent precipitation is significantly negative ($P < 0.1$) only for August–October rainfall and peaks at -0.36 for September rainfall, indicating a significant negative feedback on monsoonal rainfall. Lead-lag correlations and autocorrelations specific to region A are presented in Fig. 5. Grass cover exhibits a memory (significant autocorrelation) of at least 4 years ($P < 0.1$), while annual precipitation shows no significant memory. Instanta-

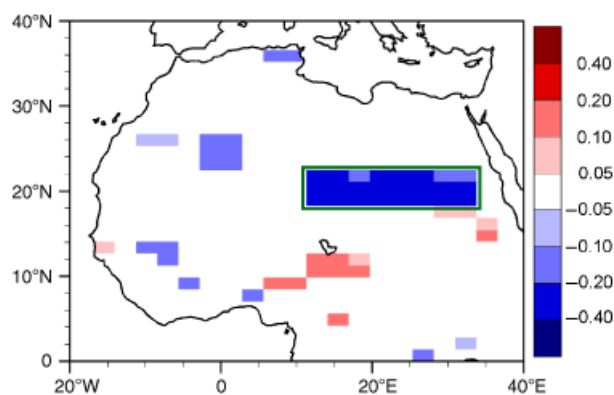


Fig. 4 Correlation between grass cover fraction and subsequent annual precipitation, from the 400-year FOAM-LPJ control simulation of 6K. Only those correlations with $P < 0.1$ are shown. The negative correlations with grass cover leading precipitation over region A (shown in the green box) suggest a negative vegetation feedback during the mid-Holocene.

neous rainfall, along with rainfall during the preceding 4 years, is an important forcing on North African grass cover. When grass cover fraction leads subsequent annual precipitation in region A, a negative correlation of -0.27 ($P < 0.1$) is achieved, indicating a significant negative feedback on annual rainfall.

Figure 6 shows a map of statistical feedback parameters for grass cover fraction and subsequent annual precipitation in the 6K control simulation (Wang *et al.*, 2007). The region of significant negative feedback, based on 500 Monte Carlo bootstrap tests ($P < 0.1$), stretches across North Africa around 20°N , particularly over region A. The strength of the negative feedback generally ranges from -2 to $-10 \text{ cm yr}^{-1}/0.2 \text{ frac}$, meaning that an increase in grass cover fraction by 0.2 typically results in a reduction in subsequent annual precipitation of 2–10 cm ($< 25\%$ of the annual mean), indicating a weak but significant negative vegetation feedback. The mean statistical feedback strength for region A, based on grass cover fraction and subsequent annual precipitation, is $-9.2 \text{ cm yr}^{-1}/0.2 \text{ frac}$, suggesting a reduction in annual rainfall by 23% due to an

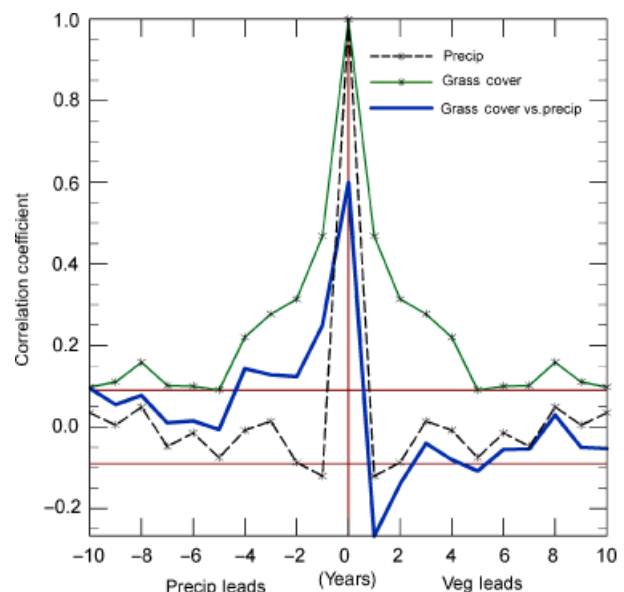


Fig. 5 Lead/lag correlations of grass cover fraction and annual precipitation over region A ($18\text{--}23^\circ\text{N}$, $11\text{--}36^\circ\text{E}$) in the 400-year FOAM-LPJ control simulation for 6K. The autocorrelation of grass cover fraction is shown in the thin green line with asterisks. The autocorrelation of precipitation is shown in the dashed black line. The blue line shows a lead/lag correlation between annual grass cover fraction and annual precipitation, with vegetation leading on the right side and precipitation leading on the left side. The horizontal red line indicates statistically significant correlations at $P < 0.1$. Annual precipitation shows no significant memory, while grass cover shows a memory of at least 4 years. Precipitation positively forces grass cover, while grass cover imposes a negative forcing on subsequent precipitation.

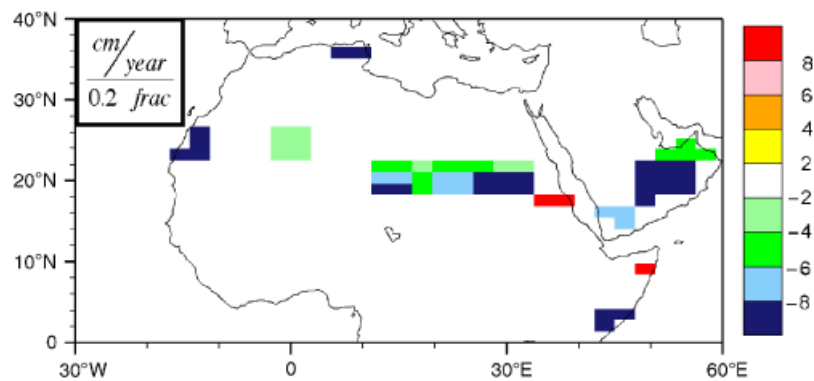


Fig. 6 Statistical vegetation feedback parameter for the 400-year FOAM-LPJ control simulation of 6K, based on grass cover fraction and annual precipitation. The units of the feedback parameter are $\text{cm yr}^{-1}/0.2 \text{ frac}$, quantifying the change in precipitation resulting from an increase in grass cover fraction by 0.2. Only those values that achieve $P < 0.1$ from 500 Monte Carlo tests are shown. Region A shows a significant negative vegetation feedback, on the annual time scale.

increase in grass cover fraction by 0.2. The feedback parameters for region A, computed at lags of 1–4 years, are -13.8 , -10.8 , -3.6 , and $-8.8 \text{ cm yr}^{-1}/0.2 \text{ frac}$, resulting in a mean feedback of $-9.2 \text{ cm yr}^{-1}/0.2 \text{ frac}$.

In the 400-year 6K control simulation of FOAM-LPJ, there are 80 cases in which the grass cover fraction over region A shows a collapse of at least 0.2 between subsequent years. Compositing these cases reveals an anomaly of $+6 \text{ cm yr}^{-1}$ in annual rainfall following the grass cover reduction, as further evidence of the negative feedback on annual precipitation. The increase in rainfall is mainly limited to August–October, peaking at $+4 \text{ cm}$ in September, while there is evidence of a slight drying in May–June. The negative feedback of grass cover on annual precipitation is limited to the rainy season.

Across North Africa, the negative feedback on annual precipitation weakens over time and eventually reverses sign in the transient 6500-year simulation of FOAM-LPJ (Fig. 7). During the mid-Holocene, before the vegetation collapse, the feedback strength over $13\text{--}23^\circ\text{N}$, $11\text{--}36^\circ\text{E}$ (includes region A and land to the south, into which the grasslands migrate) averages -10 to $-20 \text{ cm yr}^{-1}/0.2 \text{ frac}$ (Fig. 7a). By 4K, the feedback strength is close to zero and later becomes positive, although the amount of grass cover is limited at that time. Snapshot control experiments for 5.5K and 5K are performed (not shown) to further investigate the time-evolving nature of vegetation feedbacks. As the grasslands shift southward, the area of negative feedback expands but weakens in the 5.5K control and later lacks a coherent signal in the 5K control.

The number of grid cells over North Africa ($0\text{--}39^\circ\text{N}$, $30^\circ\text{W}\text{--}58^\circ\text{E}$) with a significant ($P < 0.1$) negative correlation between grass cover fraction and subsequent annual precipitation in the transient simulation is tallied

each year, based on a correlation across the surrounding 100-year window (Fig. 7b). The data was detrended with a low-pass filter. The number of grid cells diminishes from approximately 26 at 6.5K to 10 at 0K, as the area of negative feedback contracts. Likewise, the spatial extent of North African grasslands narrows with time. By 0K, there is no distinct pattern of vegetation feedback across North Africa. The negative vegetation feedback on annual precipitation is unique to North Africa and to the mid-Holocene based on this 6500-year transient simulation (Fig. 7).

Dynamical assessment: FOAM-LPJ

In order to evaluate the statistical feedback assessment, we perform initial value ensemble experiments (Liu & Wu, 2004) to quantify the climate system response to fluctuations in North African grass cover during the mid-Holocene. In ensemble F0.2, an initial reduction in grass cover fraction by 0.2 is imposed over region A compared with the 6K control simulation. This grass cover anomaly weakens over time and is mostly gone by the end of the fourth year of the ensemble experiments (Fig. 8). Figure 5 reveals a strong memory in grass cover out to 4 years in region A; the grass recovery time in F0.2 is consistent with this memory. The climatic response to this vegetation anomaly is greatest in the first year of the ensembles and weakens in subsequent years.

The actual mean reduction in grass cover imposed in these ensemble experiments is 0.17 (Fig. 8), because in some cases, the initial grass cover fraction is < 0.2 . Therefore, dynamical assessments are multiplied by $0.2/0.17 = 1.18$ to scale the response to a 0.2 reduction in grass cover. This permits a more straightforward comparison with the statistical assessments of feedback.

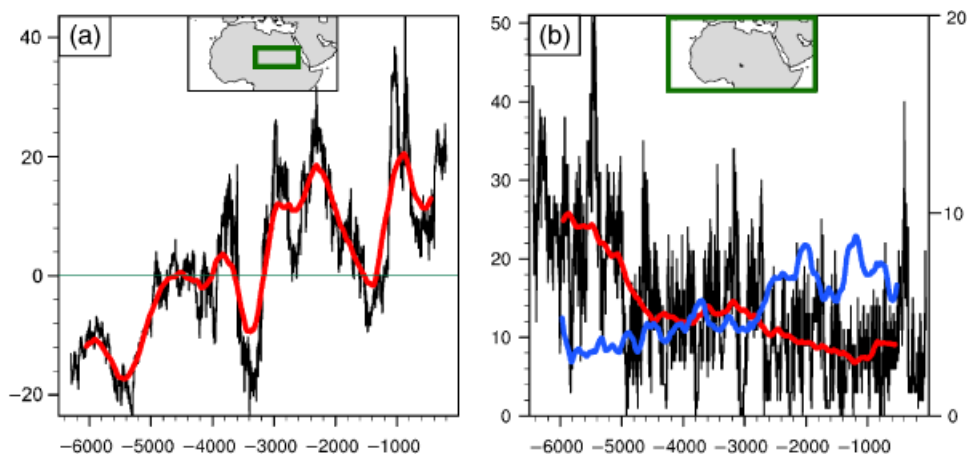


Fig. 7 (a) Statistical vegetation feedback parameter for 13–23°N, 11–36°E from the 6500-year transient simulation of FOAM-LPJ. The parameter quantifies the impact of grass cover on annual precipitation, with the units of $\text{cm yr}^{-1}/0.2 \text{ frac}$. The feedback is computed each year based on the surrounding 400 years of detrended grass cover fractions and detrended annual precipitation. The negative vegetation feedback during the mid-Holocene weakens over time and eventually becomes positive. The area used here includes both Region A and the area south of that region since grasslands shift southward over time in the simulation. (b) Number of grid cells across North Africa (0–39°N, 30°W–58°E) with a significant ($P < 0.1$) negative correlation between grass cover fraction and subsequent annual precipitation in the 6500-year transient simulation of FOAM-LPJ (shown in the black line). Each year's correlation is based on the surrounding 100-year window. The red lines in (a) and (b) are the running mean. The area of negative vegetation feedback diminishes over time. The blue line in (b) shows the number of grid cells with a significant positive correlation, using the y -axis on the right side of the plot. The area of positive feedback increases over time.

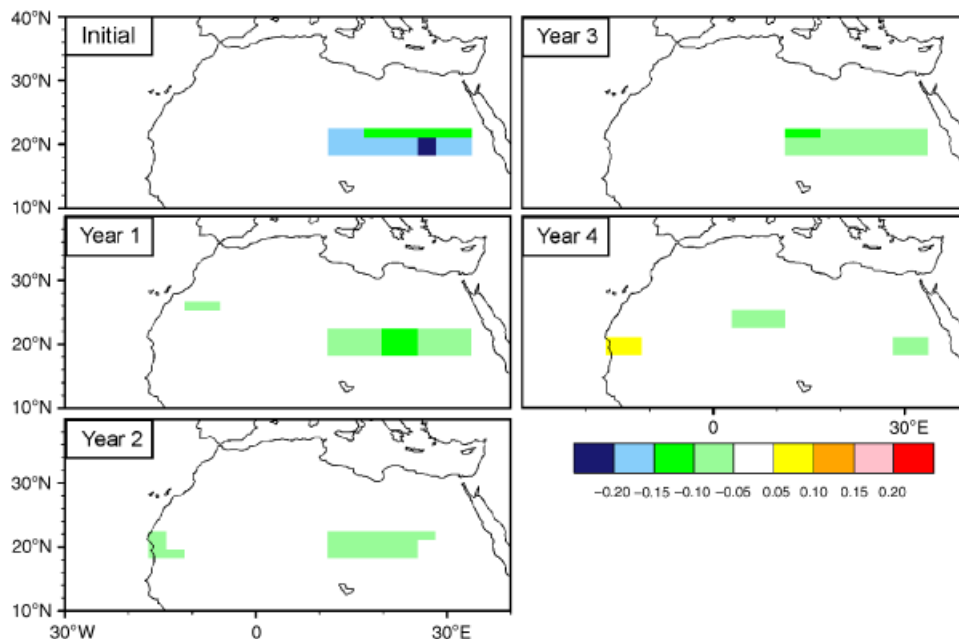


Fig. 8 Grass cover fraction anomaly, computed as the mean grass cover from the F0.2 ensemble experiments minus the mean grass cover from the 400-year 6K control experiment. Initial anomalies and also anomalies at the end of years 1, 2, 3, and 4 of the ensemble experiments are shown. The initial reduction of about 0.2 weakens over time as vegetation is allowed to grow back.

Based on the 100 ensemble members of F0.2, a reduction in region A's grass cover fraction by 0.2 results in a mean increase in subsequent annual precipitation of 9.7 cm, giving a dynamical feedback assessment of

$-9.7 \text{ cm yr}^{-1}/0.2 \text{ frac}$. The dynamical assessment compares favorably, in sign and magnitude, with the statistical assessment of $-9.2 \text{ cm yr}^{-1}/0.2 \text{ frac}$. In both cases, the negative feedback is weak but statistically significant.

In comparing year 1's annual precipitation in the F0.2 ensemble to the 6K control simulation, there is substantial noise among the response (Fig. 9). The negative feedback response in year 1 is most distinct in September, with a mean increase in monthly precipitation of 5.0 cm. The signal-to-noise ratios (SNRs) are quite low and peak in September ($\text{SNR} \approx 0.4$), during the monsoon season. Compared with the control simulation, September precipitation in year 1 increases in 58 ensemble members. The magnitude of the positive rainfall anomalies often exceeds those of the negative anomalies (Fig. 9). Specifically, 32 ensemble members of F0.2 show a positive anomaly in excess of $+1\sigma$ (standard deviation) of the control mean compared with 7 with a negative anomaly exceeding -1σ . A large number of ensemble members is clearly needed to study North African vegetation feedbacks due to substantial rainfall variability. A composite (Fig. 9c) of the 80 cases of grass cover decline in the 6K control likewise reveals a negative vegetation feedback, with a peak response in September.

Ensemble F0.2. The initial reduction in grass cover for ensemble F0.2 leads to a significant increase of $5.0 \text{ cm month}^{-1}$ in September precipitation for year 1; this response weakens substantially in subsequent years (Fig. 10a). The reduction in LAI, corresponding to the initial imposed anomaly in grass cover, is only significant during June–October of year 1 (Fig. 10e). The negative precipitation feedback (Fig. 10a) serves to reduce the negative vegetation anomaly by contributing to its regrowth. The response of the upper soil water fraction to the initial grass cover reduction closely matches that of precipitation, with a significant increase in upper soil water fraction by 0.11 in September of year 1 (Fig. 10i). The lower soil water fraction shows a significant accumulation during the

first 2 years (reaching an anomaly of $+0.07$), with its long memory evident through the persistent response (Fig. 11a); reduced grass cover corresponds to less available roots to extract the deep soil water.

The negative vegetation feedback on annual precipitation is largely due to a competition between transpiration and bare soil evaporation. During year 1, ensemble F0.2 shows a reduction in transpiration, peaking at $-1.0 \text{ cm month}^{-1}$ during August of year 1 (Fig. 11e), due to less abundant grass available to tap soil water. However, the initial reduction in grass cover leads to an increased role of bare soil evaporation. There is a significant increase in July–October bare soil evaporation in year 1 of ensemble F0.2, compared with the control simulation, peaking at $+3.5 \text{ cm month}^{-1}$ in September. The intensified monsoon

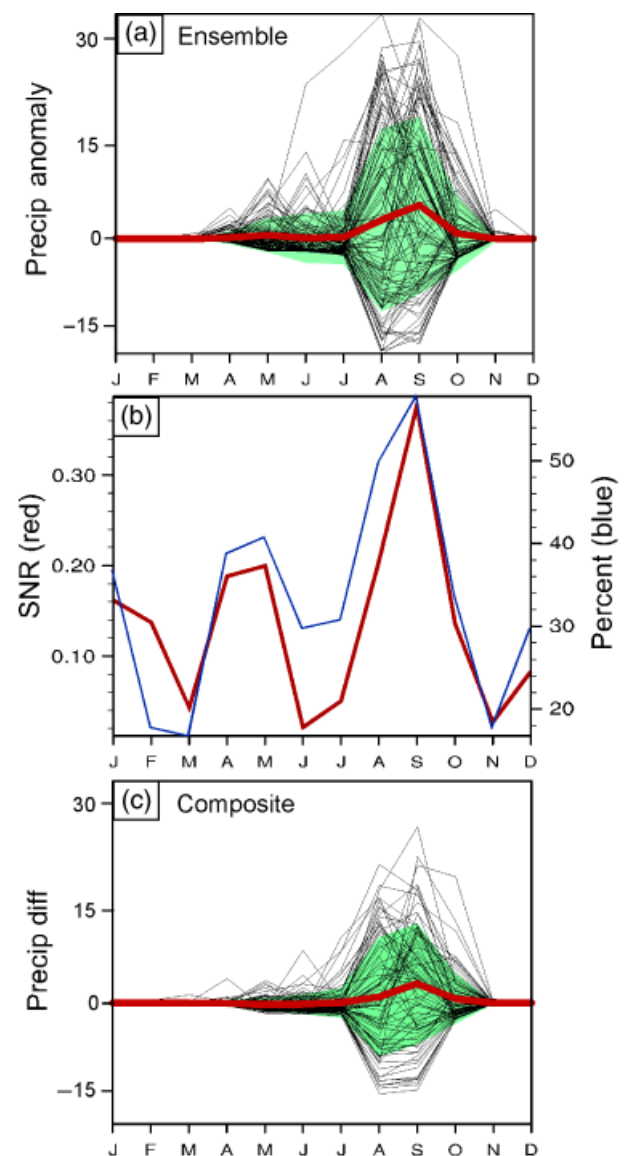


Fig. 9 (a) Difference in monthly mean precipitation (cm month^{-1}) between year 1 of 100 ensemble experiments (F0.2) and the 6K control experiment of FOAM-LPJ (ensemble-control). The red line indicates the mean difference. The largest increase in monthly precipitation, due to a reduction in grass cover, occurs during September. Green shading indicates ± 1 standard deviation among the ensemble members. (b) Signal-to-noise ratios (SNR) of the differences in precipitation from (a) are shown in red. The percentage of ensemble members with an increase in precipitation due to reduced grass cover is shown in blue. September shows the largest SNR and percent of cases with increased precipitation. (c) Composite ($N = 80$) of monthly precipitation anomalies (cm month^{-1}) following decreases in grass cover fraction between subsequent years of at least 0.2 in the 6K FOAM-LPJ control simulation. The green shading and red line have the same meaning as in (a).

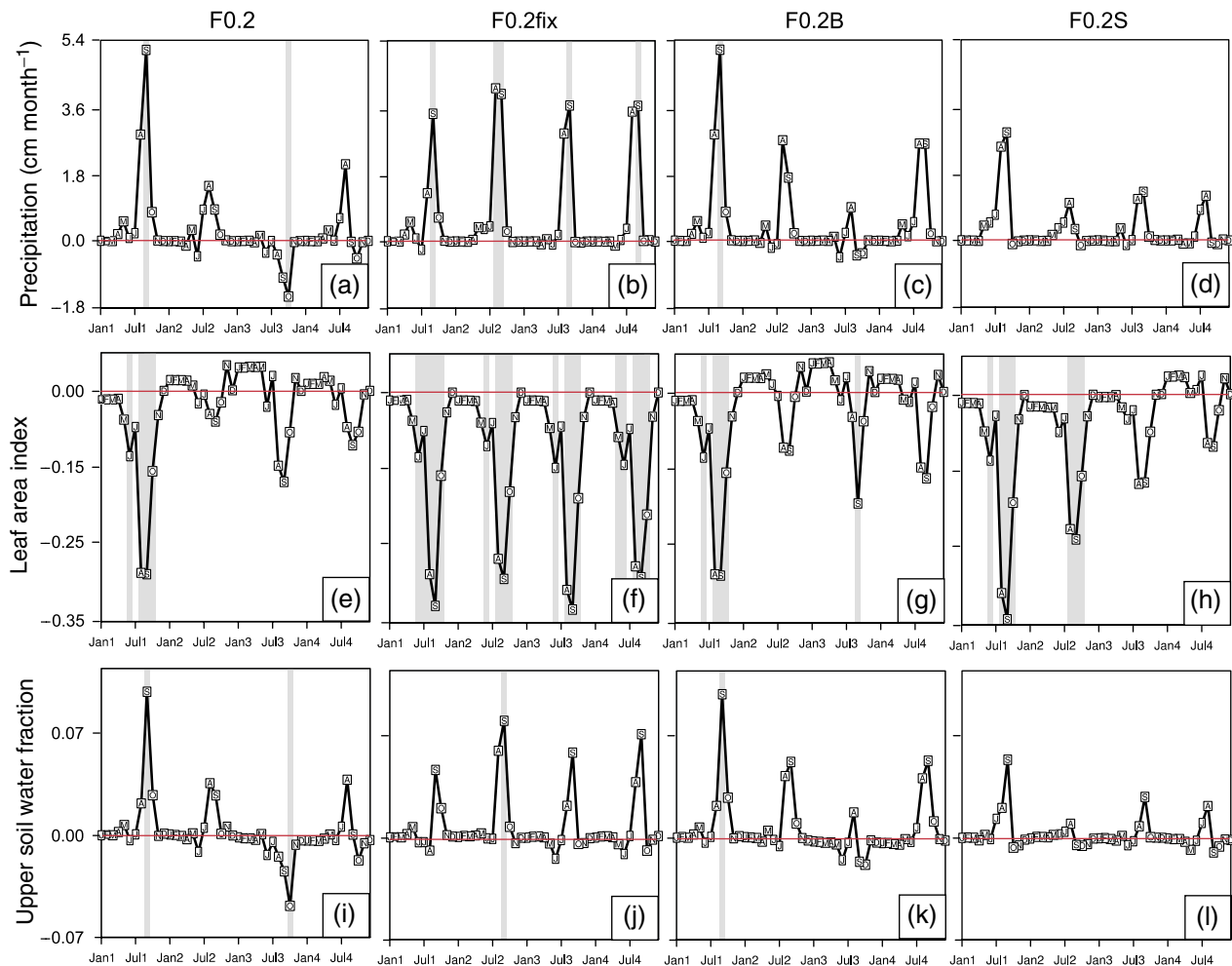


Fig. 10 Difference in (a–d) precipitation (cm month^{-1}), (e–h) leaf area index, and (i–l) upper soil water fraction between ensemble experiments and the 6K control experiment (ensemble-control). Differences are presented for ensembles (a, e, i) F0.2, (b, f, j) F0.2fix, (c, g, k) F0.2B, and (d, h, l) F0.2S. The x-axis ranges from January of ensemble year 1 to December of ensemble year 4. The horizontal red lines represent zero difference between ensembles and control. Vertical gray shading indicates statistically significant ($P < 0.1$) differences, based on t -tests.

during the mid-Holocene supports relatively high soil water content and therefore substantial bare soil evaporation from the top soil layer. Ten percent of grass roots access the lower soil water. Diminished transpiration allows lower soil water to accumulate (Fig. 11a, e). Because the increase in bare soil evaporation outweighs the decrease in transpiration, there is a net increase in evaporation (+3.0 cm in September of year 1), leading to greater atmospheric moisture and rainfall. Climatologically, for region A, annual transpiration and bare soil evaporation are equal in magnitude, yet the response to a change in grass cover is not equivalent between the two variables, revealing that their sensitivity is unique from their

climatology. While it is evident that local moisture recycling is a critical component of the negative feedback over North Africa, imbalances between changes in precipitation and evaporation in F0.2 are explained by changes in runoff and moisture flux convergence (Table 2).

The negative vegetation feedback simulated by FOAM-LPJ for North Africa during the mid-Holocene is evidently a hydrological feedback and not the result of changing surface albedo. The initial 0.2 reduction in grass cover fraction in F0.2 lowers the surface albedo by 0.01 in September of the first year (Table 2) by wetting and darkening the soil. A substantial increase in surface albedo is not produced. A decrease in sensible heating

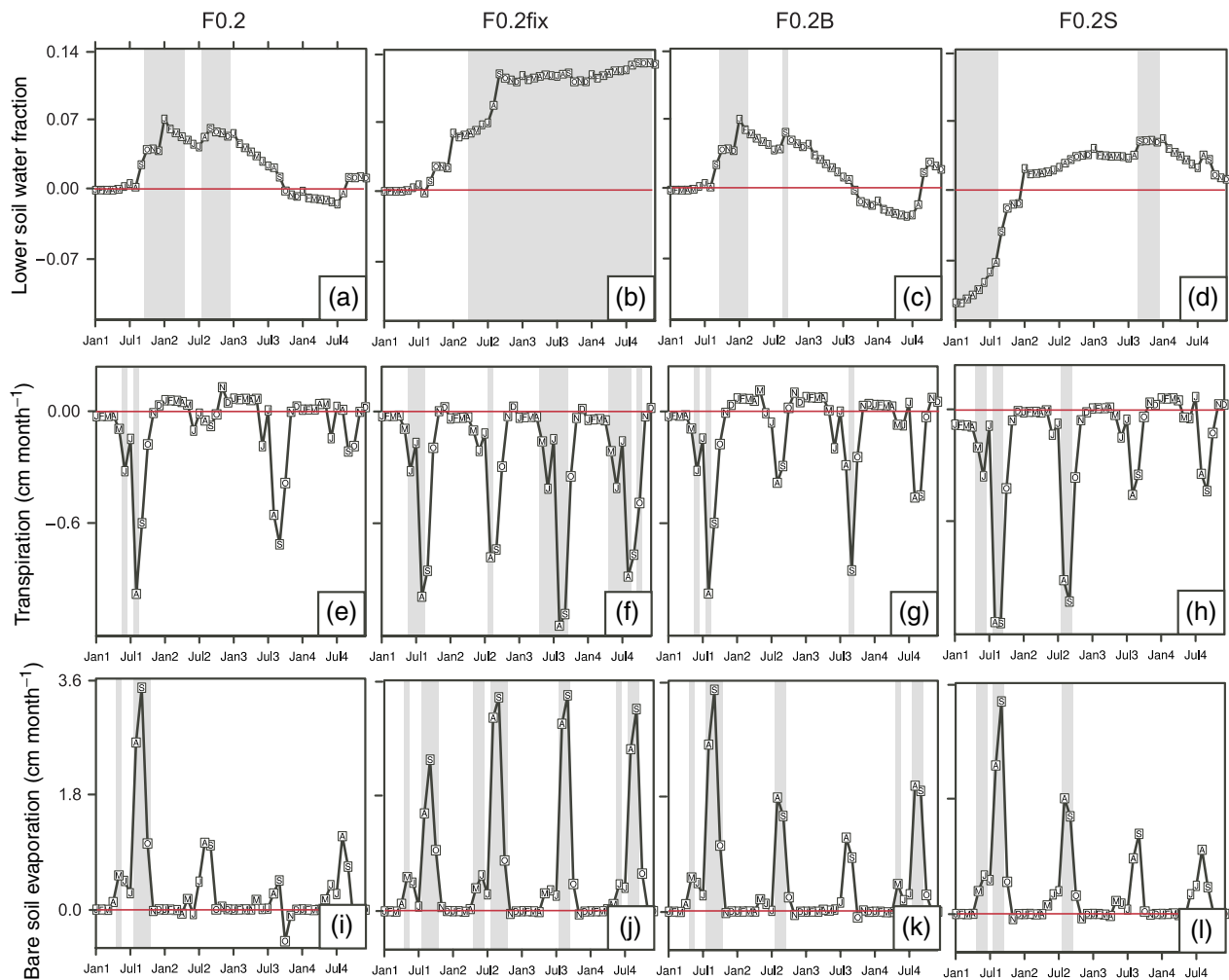


Fig. 11 Difference in (a–d) lower soil water fraction, (e–h) transpiration (cm month⁻¹), and (i–l) bare soil evaporation (cm month⁻¹) between ensemble experiments and the 6K control experiment (ensemble-control). Differences are presented for ensembles (a, e, i) F0.2, (b, f, j) F0.2fix, (c, g, k) F0.2B, and (d, h, l) F0.2S. The x-axis ranges from January of ensemble year 1 to December of ensemble year 4. The horizontal red lines represent zero difference between ensembles and control. Vertical gray shading indicates statistically significant ($P < 0.1$) differences.

and increase in latent heat accompanies a slight cooling at the surface and reduction in boundary layer height during year 1 (Table 2). Increased specific humidity supports a small enhancement in total cloud amount.

Ensemble F0.2fix. In ensemble F0.2fix, the initial grass cover fraction is reduced by 0.2 and then held fixed throughout the 4-year simulations. Consequently, there is a persistent negative LAI anomaly between -0.25 and -0.35 during the wet season for all 4 years, compared with the control simulation (Fig. 10f). Wet season precipitation is significantly increased during all 4 years, related to the persistent negative feedback (Fig. 10b). While the upper soil water fraction increases during the wet season of all 4 years, the change is only significant

in the second year (Fig. 10j). A persistent reduction in transpiration and even greater increase in bare soil evaporation throughout the 4 years (Fig. 11f, j) support the negative feedback on annual rainfall by increasing net evapotranspiration. Reduced transpiration leads to a steady increase in lower soil water fraction until it approaches a new equilibrium state with an anomaly of $+0.10$ to $+0.14$ in years 3–4 (Fig. 11b).

Ensemble F0.2B. In ensemble F0.2B, we impose an initial reduction in grass cover fraction by 0.2 and an initial 20% reduction in grass biomass, namely root and leaf carbon. This is more physically realistic than ensemble F0.2 because a reduction in vegetation cover requires a decrease in biomass. A reduction in biomass, due to its

substantial memory, should prolong the negative vegetation anomaly and the feedback response, particularly in the second year of the ensemble. There is minimal difference between F0.2 and F0.2B in the first year, since vegetation cover and biomass are not updated in the Northern Hemisphere until the end of each simulation year. We focus on year 2 in comparing

Table 2 Difference in 15 variables between ensemble sets and 6K control simulation (ensemble-control)

	Units	Climo	F0.2	F0.2S
Precipitation	cm	14.1	+ 5.0	+ 2.8
Evaporation	cm	10.5	+ 3.0	+ 2.2
Bare soil evaporation	cm	6.3	+ 3.5	+ 3.4
Transpiration	cm	4.3	−0.6	−1.2
Runoff	cm	3.7	+ 0.8	−0.3
Upper soil water fraction		0.49	+ 0.11	+ 0.06
Lower soil water fraction		0.67	+ 0.02	−0.04
Sensible heat flux	W m ^{−2}	57	−19	−14
Latent heat flux	W m ^{−2}	102	+ 28	+ 21
Surface specific humidity	g kg ^{−1}	13	+ 1.7	+ 1.1
Surface air temperature	°C	32	−0.8	−0.6
Surface albedo		0.20	−0.012	−0.008
Boundary layer height	m	582	−90	−61
Leaf area index		0.94	−0.28	−0.34
Total cloud cover fraction		0.69	+ 0.06	+ 0.02

Results are presented for September of the first year of the ensembles. Shading indicates $P < 0.1$. 'Climo' refers to the September climatological mean from the 6K FOAM-LPJ control simulation. Results are not shown for F0.2fix and F0.2B since significant differences from F0.2 are not expected in the first year.

F0.2 and F0.2B. Of the 15 variables in Table 3, the increase in lower soil water fraction by +0.06 is the only significant anomaly in September of year 2, for the F0.2 ensemble. For all variables, the response in F0.2B is either greater or equivalent to that of F0.2 in September of year 2, with the increase in bare soil evaporation and lower soil water fraction and decrease in surface albedo achieving statistical significance for F0.2B. The reduction in LAI is nearly twice as large in F0.2B than in F0.2, leading to a greater reduction in transpiration and increase in bare soil evaporation, thereby further increasing evaporation and precipitation. Precipitation in September of year 2 increases by +1.7 cm in F0.2B compared with +0.7 in F0.2, suggesting an enhanced negative feedback on rainfall.

Ensemble F0.2S. Both initial grass cover fraction and lower soil water fraction are reduced over region A in ensemble F0.2S, compared with the control simulation. In the control simulation, periods of low vegetation cover typically coincide with periods of depleted soil water, so ensemble F0.2S reveals the impact of both anomalies; the F0.2 ensemble is artificial in a sense by reducing grass cover but not altering the soil water initially. While rainfall increases throughout the 4 years of F0.2S compared to the control experiment, this increase never achieves statistical significance (Fig. 10d), unlike in F0.2, which shows a significant increase in September of year 1. The initial reduction in lower soil water fraction weakens the negative vegetation feedback on rainfall. The positive anomaly in rainfall in September of year 1 is +5.0 cm in F0.2 but

Table 3 Difference in fifteen variables between ensemble sets and 6K control simulation (ensemble-control)

	Units	Climo	F0.2	F0.2fix	F0.2B	F0.2S
Precipitation	cm	14.1	+ 0.7	+ 4.0	+ 1.7	+ 0.2
Evaporation	cm	10.5	+ 0.9	+ 2.7	+ 1.2	+ 0.5
Bare soil evaporation	cm	6.3	+ 1.1	+ 3.4	+ 1.5	+ 1.5
Transpiration	cm	4.3	−0.1	−0.7	−0.4	−1.1
Runoff	cm	3.7	+ 0.1	+ 1.2	+ 0.6	−0.3
Upper soil water fraction		0.49	+ 0.02	+ 0.08	+ 0.05	−0.01
Lower soil water fraction		0.67	+ 0.06	+ 0.12	+ 0.06	+ 0.04
Sensible heat flux	W m ^{−2}	57	−6	−18	−7	−2
Latent heat flux	W m ^{−2}	102	+ 9	+ 26	+ 12	+ 5
Surface specific humidity	g kg ^{−1}	13	+ 0.4	+ 1.3	+ 0.5	+ 0.1
Surface air temperature	°C	32	−0.2	−0.7	−0.4	−0.2
Surface albedo		0.20	−0.005	−0.011	−0.008	−0.002
Boundary layer height	m	582	−22	−81	−27	−4
Leaf area index		0.94	−0.05	−0.30	−0.09	−0.22
Total cloud cover fraction		0.69	+ 0.01	+ 0.04	+ 0.01	−0.01

Results are presented for September of the second year of the ensembles. Shading indicates $P < 0.1$. 'Climo' refers to the September climatological mean from the 6K FOAM-LPJ control simulation.

Table 4 Statistical and dynamical response in annual variables to a decrease in North African grass cover fraction by 0.2

	Units	Memory (years) (FOAM)	Statistical (FOAM)	Dynamical (FOAM)	Memory (years) (CCSM)	Statistical (CCSM)
Precipitation	cm yr ⁻¹	0	+ 9.2	+ 9.7	0	+ 3.3
Evaporation	cm yr ⁻¹	0	+ 4.7	+ 6.1	0	+ 0.7
Bare soil evaporation	cm yr ⁻¹	0	+ 10.2	+ 8.6	4	+ 8.7
Transpiration	cm yr ⁻¹	3	−5.5	−2.5	5	−8.0
Upper soil water fraction		0	+ 0.013	+ 0.014	3	+ 0.008
Lower soil water fraction*		1	+ 0.098	+ 0.013	5	+ 0.014
Sensible heat flux	W m ⁻²	0	−2.8	−3.5	5	+ 2.6
Latent heat flux	W m ⁻²	0	+ 3.6	+ 5.0	0	+ 0.6
Surface specific humidity	g kg ⁻¹	0	+ 0.19	+ 0.27	0	+ 0.08
Surface air temperature	°C	0	−0.15	−0.12	1	+ 0.05
Surface albedo*		2	+ 0.001	−0.001	10	+ 0.005
Boundary layer height	m	0	−13.8	−12.6	x	x

The impact of a decrease in North African grass cover fraction by 0.2, over region A (18–23°N, 11–36°E), is statistically (vegetation feedback parameter) and dynamically (F0.2 ensemble experiments) assessed for 12 variables. The memory of these 12 variables, in years, is presented, based on autocorrelations ($P < 0.1$) (number of lag years with significant positive autocorrelation). Asterisks indicate variables with the largest percent difference between the statistical and dynamical assessments; these variables have a memory of 1–2 years. Statistical feedback parameters are computed based on a mean of the feedback strength at time lags of 1–4 years. Since grass cover has a significant memory of at least 4 years, the maximum lag used is 4 years; beyond that lag, feedback estimates are unreliable. In the last two columns, results for CCSM2 are presented for 9–21°N, 15–38°E, including the variables' memory and the statistical estimation of grass cover feedback on annual precipitation. For CCSM2, the 'transpiration' term of evaporation consists of both canopy transpiration and canopy evaporation. For the purpose of comparing with FOAM-LPJ, which has an upper soil layer depth of 0.5 m and a lower soil layer depth of 1.0 m, we define the upper soil as the top six layers and the lower soil as the next two layers in CCSM2, which has 10 soil layers. The memory of grass cover fraction is 4 years for FOAM-LPJ and 7 years for CCSM2.

CCSM2, Community Climate System Model Version 2.

only + 2.8 cm in F0.2S (Table 2). The reduction in LAI is more substantial in years 1–2 of ensemble F0.2S than F0.2 (Fig. 10h), with drier soils leading to a further reduction in grass cover beyond the initial 0.2 reduction. Upper soil water fraction anomalies fail to achieve significance in F0.2S (Fig. 10l). The initial reduction in lower soil water fraction, with a mean anomaly of −0.12 at the start of the ensemble simulations, quickly vanishes after the wet season of year 1 and becomes a positive anomaly for years 2–4, although not statistically significant as in F0.2 (Fig. 11d). The greater negative LAI anomaly in F0.2S than F0.2 results in a larger reduction in transpiration in the first 2 years. This, in combination with lower soil water content in F0.2S, leads to a smaller increase in evaporation and atmospheric moisture and, therefore, a smaller enhancement of precipitation compared to F0.2 (Table 2). Because feedbacks from vegetation and soil water act in opposite directions here, on the annual time scale, the net effect in F0.2S is a weaker land surface feedback on rainfall than in F0.2. Comparisons between ensemble F0.2 and F0.2S reveal the importance of local moisture recycling.

Statistical vs. dynamical assessment

Table 4 compares the statistical estimation of 12 variables' response to a decrease in region A's grass cover fraction by 0.2, based on the 6K control simulation, with dynamical responses from the F0.2 ensemble set. There is good agreement between the statistical assessment for precipitation (+ 9.2 cm yr⁻¹) and the dynamical assessment (+ 9.7 cm yr⁻¹), both revealing a negative vegetation feedback on annual rainfall. Most variables show fairly good agreement. Stronger fluxes from bare soil evaporation and transpiration with the statistical method cancel and result in similar net evaporative fluxes as the dynamical method (+ 4.7 cm yr⁻¹ for statistical vs. + 6.1 for dynamical). The largest percent differences in feedback strength between the statistical and dynamical methods are found for lower soil water fraction and surface albedo. These two annual variables, other than transpiration, are the only ones that exhibit significant memory, based on autocorrelations, equivalent to 1 and 2 years, respectively.

The statistical feedback parameter is typically most reliable when there is a large difference in memory

between the slow forcing variable and the fast responding variable. For instance, rainfall intrinsically has a memory of <1 week, compared with vegetation attributes that often have a memory of several months to years. If the autocorrelation of vegetation is low, then the denominator in the feedback equation is small, resulting in a large sampling error (Liu *et al.*, 2006b) and an artificially large feedback parameter. In Table 4, the feedback parameters are computed as a mean feedback for lags of 1–4 years. The statistical feedback response in subsequent annual precipitation to a reduction in grass cover by 0.2 is estimated at +13.8, +10.7, +3.5, and +8.8 cm yr⁻¹ at lags of 1–4 years, respectively, resulting in a mean response of +9.2 cm yr⁻¹. If the time series of vegetation and precipitation were infinite, then the estimated feedback would be identical at all time lags since it represents an instantaneous feedback. Since the data is limited to 400 years and region A's rainfall is quite variable, there are variations in feedback estimates at different time lags, making it necessary to average the feedback across multiple lags to achieve a reliable estimate.

Comparison of vegetation feedbacks in two GCMs

Similar to FOAM-LPJ, CCSM2 simulates a poleward expansion of North African grasslands at 6K, compared with 0K (Gallimore *et al.*, 2007). While FOAM-LPJ captures the August–September monsoon season over North Africa, the simulated monsoon rain belt pushes about 5–6° further north than observed at 0K due to a wet bias. Both models' 6K simulations show a region of negative feedback between vegetation cover and subsequent annual precipitation across eastern North Africa. This area consists of 18–23°N, 11–36°E in FOAM-LPJ and 9–21°N, 15–38°E in CCSM2 and is primarily grassland. The negative feedback in CCSM2 is weaker than in FOAM-LPJ. The statistical feedback estimates for these regions are -3.4 cm yr⁻¹/0.2 frac in CCSM2 and -9.2 cm yr⁻¹/0.2 frac in FOAM-LPJ, based on grass cover and subsequent precipitation; both achieve statistical significance ($P < 0.1$) based on Monte Carlo testing (Wang *et al.*, 2007).

Figure 12 displays lead-lag correlations between grass cover and annual precipitation from the 6K control experiment of CCSM2. While both models show no significant memory in annual precipitation, the memory of grass cover fraction is larger in CCSM2, reaching 7 years. Both models show a significant negative correlation between grass cover and subsequent annual precipitation: -0.27 ($N = 400$) in FOAM-LPJ and -0.19 ($N = 150$) in CCSM2. The forcing by precipitation is substantially stronger in FOAM-LPJ.

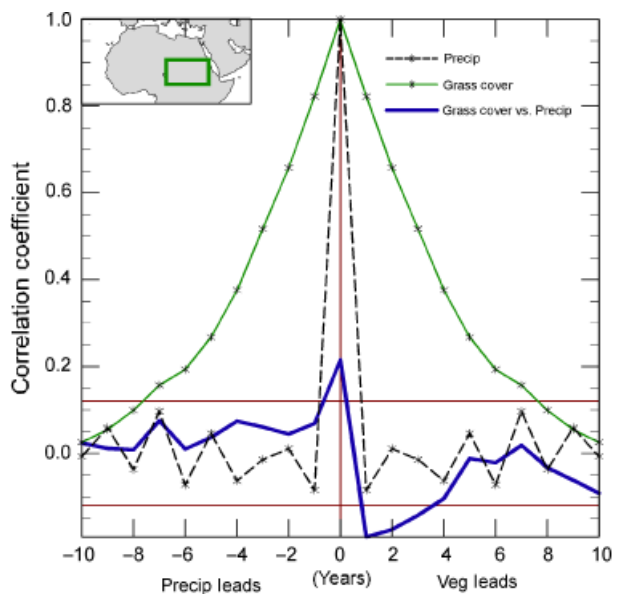


Fig. 12 Lead/lag correlations of grass cover fraction and annual precipitation over 9–21°N, 15–38°E in 150 years of the 6K control simulation of Community Climate System Model Version 2. The autocorrelation of grass cover fraction is shown in the thin green line with asterisks. The autocorrelation of precipitation is shown in the dashed black line. The blue line shows a lead/lag correlation between annual grass cover fraction and annual precipitation, with vegetation leading on the right side and precipitation leading on the left side. The horizontal brown line indicates statistically significant correlations at $P < 0.1$. Annual precipitation shows no significant memory, while grass cover shows a memory of at least 7 years. Precipitation positively forces grass cover, while grass cover imposes a negative forcing on subsequent precipitation.

Table 4 compares statistical estimates of grass cover feedbacks on several variables in the 6K control experiments of FOAM-LPJ and CCSM2, in addition to comparing the memories of these variables, based on autocorrelation. North Africa is characterized by a much more substantial memory in bare soil evaporation, soil water fraction, sensible heat flux, and surface albedo in CCSM2 than FOAM-LPJ, related to the former model's longer vegetation memory. This longer vegetation memory might be partly related to its weaker response to precipitation fluctuations, with a large canopy interception preventing much rainwater from entering the soil. Statistical feedback estimates generally agree favorably between the models. In both cases, a stronger response in bare ground evaporation than transpiration appears to result in an increase in net evaporation due to reduced grass cover, thereby producing a negative feedback. The negative feedback is responsible for a wetter soil and increased atmospheric humidity in both models. The models disagree in terms

of the sign of the response in sensible heat flux, surface albedo, and temperature, although it is evident that the albedo and related air temperature response is minimal in both models at 6K. There is not a strong contrast in albedo between grass cover and wet soil, thereby minimizing the role of vegetation-induced albedo feedbacks and emphasizing hydrological feedbacks. Following a north–south transect through central North Africa, summertime surface albedos at 6K range from 0.14 (0.11) in the Congo to 0.26 (0.27) over the Sahara in FOAM-LPJ (CCSM2). These Saharan albedos are substantially lower than the modern-day satellite estimates of 0.3–0.45 by Ba *et al.* (2001), largely since the 6K soils were wetter, but closely match the estimate of 0.27 suggested by Tetzlaff (1983) for 5500 years ago.

FOAM-LPJ and CCSM2 differ in their components of evapotranspiration. In FOAM-LPJ, evaporation consists of bare soil evaporation and transpiration. On the annual average, both terms are equal in magnitude (18 cm yr^{-1}) in the mid-Holocene. The greater feedback response of bare ground evaporation than transpiration to a change in vegetation cover is responsible for the negative vegetation feedback on annual precipitation. In CCSM2, evapotranspiration consists of bare soil evaporation, plant transpiration, and canopy evaporation. The additional third term involves the evaporation of water previously intercepted by the canopy and is nearly three times as large as canopy transpiration over $9\text{--}21^\circ\text{N}$, $15\text{--}38^\circ\text{E}$. Specifically, the three terms have magnitudes of 22, 11, and 32 cm yr^{-1} , respectively, at 6K. CCSM2 over-simulates rainfall interception by grass canopies (Gallimore *et al.*, 2007). Combining the canopy evaporation and canopy transpiration terms results in an evaporative component that is twice as large as the bare soil evaporation term, significantly different than in FOAM-LPJ. Similar to FOAM-LPJ, we speculate that the negative vegetation feedback results from a stronger response in bare soil evaporation than canopy evaporation + transpiration to a change in vegetation cover. Around 0K, transpiration fluxes over North Africa easily outweigh bare soil evaporation, thereby eliminating the negative feedback mechanism in FOAM-LPJ (not shown).

North African ($0\text{--}40^\circ\text{N}$, $20^\circ\text{W}\text{--}40^\circ\text{E}$) annual mean evapotranspiration, as simulated by FOAM-LPJ for 0K, agrees well with estimates from a simple water budget model by Mintz & Walker (1993): 53 cm yr^{-1} in FOAM-LPJ vs. 49 cm in the water budget model. This agreement helps give credibility to the findings of this paper. One limitation of FOAM-LPJ and CCSM2 is that they only update grass cover fraction once per year, so in the case of a dry, low vegetated year followed by a wet year, the grass cover will not be permitted to expand until the end of the wet year.

Vegetation feedbacks on different time scales

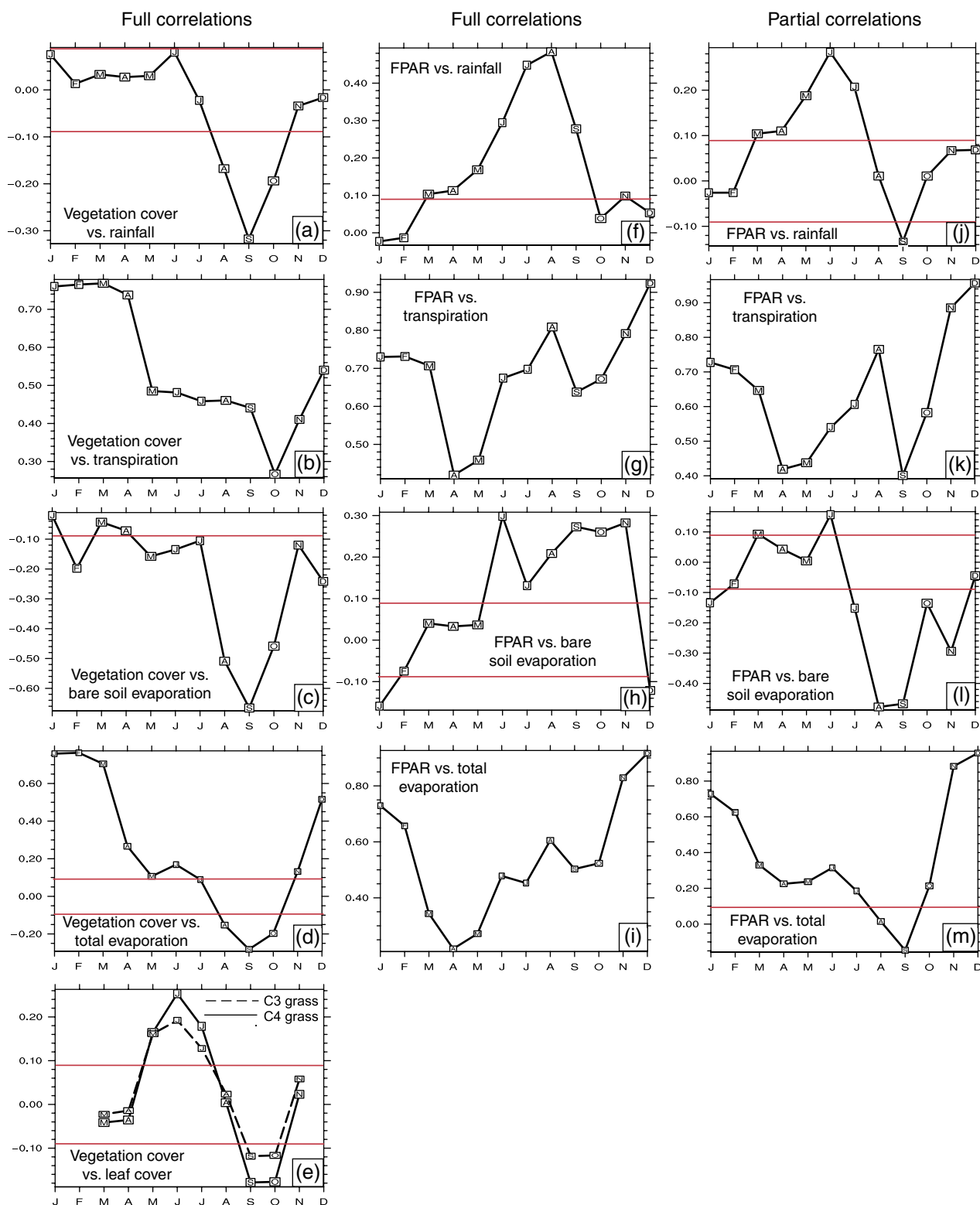
The 6K control experiment of FOAM-LPJ is characterized by a positive vegetation feedback on rainfall at the monthly time scale and a negative vegetation feedback at the annual time scale over North Africa (Wang *et al.*, 2007). Figure 13 applies both correlations and partial correlations to explain the existence of a negative vegetation feedback on the annual time scale despite the existence of a positive feedback on the monthly scale. In the case of the partial correlations, the upper soil water fraction is held ‘fixed’ in order to isolate the vegetation feedback signal from the soil moisture feedback signal. The equation for the partial correlation between variables v_1 and v_2 , while holding v_3 ‘fixed’, is

$$r(v_1, v_2)_{v_3} = \frac{r(v_1, v_2) - r(v_1, v_3) \times r(v_2, v_3)}{\sqrt{\{1 - r(v_1, v_3)^2\} \times \{1 - r(v_2, v_3)^2\}}}.$$

The correlation between annual vegetation cover and subsequent monthly rainfall is weakly positive during the first half of the year and then significantly negative during August–October (Fig. 13a). Since August–October is the monsoon season and comprises the major contributor to the annual precipitation, and the correlations are only significant during these months, vegetation has a net negative feedback on annual precipitation. While vegetation cover fraction is positively correlated with transpiration during all subsequent months (Fig. 13b), it has a strong negative correlation with subsequent August–October bare soil evaporation (Fig. 13c). This negative correlation with August–October bare soil evaporation is critical since the vast majority of region A’s bare soil evaporation occurs during August–September, when the upper soil water is replenished.

During the first half of the year, the upper soil water fraction over region A is typically less than 0.1 and transpiration exceeds bare soil evaporation (although both are quite low). The role of bare soil evaporation is mostly limited to August–October, when upper soil water fractions are substantially higher (peak at 0.48 in September). The lower soil water fraction has a weaker annual cycle and typically ranges from 0.5 to 0.7. Consequently, there is water available year-round in the deep soil for transpiration by deep roots. Therefore, high vegetation cover increases atmospheric moisture year-round through transpiration exceeding bare soil evaporation while reducing it during the wet season (August–October) through bare soil evaporation exceeding transpiration.

A high grass cover fraction supports an increase in transpiration (Fig. 13b), and therefore rainfall, during the dry season. This results in anomalously high leaf cover before the wet season (Fig. 13e). During the dry



season period of March–July, enhanced leaf coverage (positive FPAR anomaly) supports increased rainfall through transpiration from the deep soil water reservoir (Fig. 13j). The positive vegetation feedback acts during most months, outside of the wet season. However, total rainfall is quite low during this period, so the positive feedback on the monthly time scale does not produce a large net effect. Throughout the year, high FPAR anomalies are associated with enhanced transpiration (Fig. 13k). While the anomalously high leaf cover during the first half of the year supports increased rainfall through transpiration, it hinders bare soil evaporation during the wet season (by reducing the amount of exposed bare soil), as indicated by the strong negative partial correlation between monthly FPAR and the next month's bare soil evaporation during August–September (Fig. 13l). Because bare soil evaporation peaks during the wet season and exceeds transpiration then, the result is a net reduction in wet season rainfall, and, consequentially, annual rainfall. To summarize, in the 6K FOAM-LPJ simulation, enhanced vegetation cover over North Africa results in increased transpiration, rainfall, and leaf cover during the dry season but reduced bare soil evaporation, rainfall, and leaf cover later during the wet season, producing a net reduction in annual rainfall. The negative feedback mechanism is no longer valid at 0K due to dry upper soil, limited bare soil evaporation, and greater vegetation–soil albedo contrasts.

Conclusions

A combined statistical and dynamical assessment of North African vegetation feedbacks is performed for the mid-Holocene. Based on 6K simulations using two coupled models, FOAM-LPJ and CCSM2, expanded North African grasslands imposed a negative feedback on subsequent annual rainfall. The negative feedback is statistically quantified in the coupled experiments, using vegetation feedback parameters based on lagged covariance ratios. The negative vegetation feedback, based on grass cover fraction and subsequent annual rainfall, is stronger in FOAM-LPJ than CCSM2, although both achieve statistical significance.

Initial value ensemble experiments are performed with FOAM-LPJ to dynamically assess the vegetation feedback on annual rainfall. In ensemble set F0.2, grass cover fraction is initially reduced over region A (area of expanded grass cover during mid-Holocene) by 0.2 and the climate response is analyzed. The statistical and dynamical feedback assessments closely agree in sign and magnitude. The negative feedback largely results from a competition between transpiration and bare soil evaporation. The positive vegetation feedback acts in the dry season through transpiration, while the monsoon season is characterized by a negative feedback, through bare soil evaporation. The 6.5–0K transient simulation reveals a weakening of the negative feedback and eventually a sign reversal, with drier conditions closer to modern times limiting the role of bare soil evaporation. Both FOAM-LPJ and CCSM2 suggest that hydrological feedbacks dominate over albedo feedbacks in regards to North African vegetation during the mid-Holocene, in contrast to Charney's albedo theory. The contrast in surface albedo between vegetated soil and bare, moist soil is quite low, thereby minimizing the role of albedo feedbacks over North Africa during the mid-Holocene, which are typically associated with strong positive feedbacks.

The negative vegetation feedback over North Africa during the mid-Holocene serves to protect the grasslands from external forcings (Cook *et al.*, 2006). The orbitally-induced drying trend from mid-Holocene to present reduces the vegetation cover, but the negative vegetation feedback opposes this change by increasing the rainfall as vegetation declines. As the negative vegetation feedback weakens over time, this ecological buffer vanishes and sets the stage for the simulated vegetation to collapse around 5.5K, in agreement with observational estimates (de Menocal *et al.*, 2000).

While the negative vegetation feedback for the mid-Holocene over North Africa is a robust signal in FOAM-LPJ and CCSM2, it is seemingly impossible to validate the findings against observations. Little is known about the role of observed modern-day vegetation feedbacks (Liu *et al.*, 2006b), and paleodata from the mid-Holocene cannot offer much insight into vegetation feedbacks during the past. Further measurements that partition evaporation into bare soil evaporation and

Fig. 13 Correlation between total vegetation cover fraction in region A with subsequent (a) rainfall, (b) bare soil evaporation, (c) transpiration, (d) total evaporation, and (e) leaf cover fraction, from the 6K control simulation of FOAM-LPJ ($N = 400$ years). Vegetation cover fraction is updated once per year in FOAM-LPJ, in December for the Northern Hemisphere. Correlation between monthly fraction of photosynthetically active radiation and the following month's (f) rainfall, (g) bare soil evaporation, (h) transpiration, and (i) total evaporation. (j–m) Same as (f–i), but partial correlations, with upper soil water fraction held 'fixed.' This crudely isolates vegetation feedbacks from soil moisture feedbacks. The horizontal red lines represent $P < 0.1$ significance.

transpiration are needed over North Africa to tune the model for modern times. FOAM's overactive hydrological cycle likely results in an exaggerated bare soil evaporation term, although to what extent is not known.

An additional transient experiment from 6.5K to 4K was performed to test the robustness of the negative vegetation feedback and abrupt collapse of North African vegetation and to determine if these features are the result of the model's parameterization of bare soil evaporation. In this case, the bare soil evaporation term was multiplied by the upper soil water fraction, causing the total bare soil evaporation to drop off rapidly during times of low soil wetness. This reflects the idea that bare soil evaporation mainly originates from a relatively thin soil layer near the surface (J. Norman, personal communication). The new simulation is also characterized by the negative vegetation feedback on North African precipitation and abrupt vegetation collapse, illustrating that these features are not simply the result of the parameterization of bare soil evaporation.

Further improvements are needed among DGVMs in simulating grass cover growth and senescence. Both FOAM-LPJ and CCSM2 update the grass cover fraction once per year, in December for the Northern Hemisphere, while leaf cover fraction varies daily. This is acceptable for forest cover fraction, because trees grow slowly, but not ideal for grass cover. Consider the case of a dry year followed by a wet year over a semi-arid grassland. When the grass cover fraction is updated at the end of the dry year, it is substantially reduced. Then, in the following rainy year, soils become wet but grass cover is limited until the update at the end of that year, thus leading to a season-long wet, low vegetated soil. While leaf cover fraction can vary, the total leaf area is limited by the low grass cover fraction. Such situations of wet, mostly unvegetated soil are not realistic, and thus further improvement in the simulation of subannual grass cover variations is needed.

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