Diagnosing Land–Atmosphere Interaction from a Regional Climate Model Simulation over West Africa

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ABSTRACT

Land–atmosphere interaction at climatological time scales in a large area that includes the West African Sahel has been explicitly explored in a regional climate model (RegCM) simulation using a range of diagnostics. First, areas and seasons of strong land–atmosphere interaction were diagnosed from the requirement of a combined significant correlation between soil moisture, evaporation, and the recycling ratio. The northern edge of the West African monsoon area during June–August (JJA) and an area just north of the equator (Central African Republic) during March–May (MAM) were identified. Further analysis in these regions focused on the seasonal cycle of the lifting condensation level (LCL) and the convective triggering potential (CTP), and the sensitivity of CTP and near-surface dewpoint depressions $H_{dew}$ to anomalous soil moisture. From these analyses, it is apparent that atmospheric mechanisms impose a strong constraint on the effect of soil moisture on the regional hydrological cycle.

1. Introduction

Land–atmosphere interaction is manifest at a wide range of spatial and temporal scales (Van den Hurk and Blyth 2008). The planetary boundary layer is affected by the evaporation from the surface by changing the atmospheric humidity, which changes surface evaporation by a modified humidity gradient within a couple of hours. The lifting condensation level (LCL) is strongly coupled to the surface relative humidity and soil moisture (Betts 2004). The ability of the atmosphere to produce precipitation is partly dependent on surface fluxes affecting convective activity and moisture supply at a similar time scale (Findell and Eltahir 2003a; Taylor and Ellis 2006). Soil moisture anomalies may affect subsequent rainfall at synoptic (Koster et al. 2000, 2003) or monthly-to-seasonal (Koster et al. 2004) time scales. Seneviratne et al. (2006) show shifts in regions of strong coupling induced by climate change on multidecadal time scales. The monthly-to-seasonal time scale is particularly relevant, as it holds promises for improving seasonal forecasting by exploiting the predictability contained in the soil moisture state (Douville 2009).

The hydroclimate in areas with a strong annual cycle displays strong spatial and temporal gradients of the strength of land–atmosphere interaction at the seasonal time scale. Examples are areas within the annual cycle of the equatorial tropical convection belt and monsoon regions, where annual cycles of temperature contrasts between land and ocean drive a strong regional hydrological cycle (e.g., Sultan et al. 2003). Land–atmosphere interaction and, in particular, anomalous soil moisture under conditions of long soil memory, are held partly responsible for anomalies in the strength or duration of wet or dry episodes in, for example, the West African monsoon area (Koster et al. 2004; Douville et al. 2007).

Recently, Dirmeyer et al. (2009) developed an elegant framework in which a series of observable correlations are indicative of a strong land–atmosphere interaction and associated predictability. To have a strong and predictable atmospheric response to soil moisture anomalies, three conditions should be met. First, surface evaporation is positively correlated with soil moisture. Negative correlations point to conditions where evaporation is depleting the soil reservoir and is primarily limited by available energy or atmospheric demand. Second, evaporation is positively correlated with soil memory. This criterion ensures that anomalous wetness conditions are actually maintained in the subsequent period. Third, atmospheric recycling of water (i.e., the fraction of precipitation...
originating from local evaporation) is positively correlated with soil moisture. This ensures that the atmosphere is actually responsive to local wetness anomalies and that precipitation is not entirely driven by remote advection and/or atmospheric circulation. Their analysis based on a mixture of multiyear (surface) model results and observational datasets reveals that only few areas in West Africa show a strong land–atmosphere coupling in either of the main seasons.

The statistical correlations in the analysis of Dirmeyer et al. (2009) are well interpreted in terms of general behavior of the interacting land–atmosphere system. However, they do not reveal a lot of information on the responsible mechanisms that play a role in this interaction. Dirmeyer et al. (2006) and Guo et al. (2006) address the surface evaporation response to soil moisture and radiation to explain the multimodel spread in the analysis of regions of strong land–atmosphere coupling presented by Koster et al. (2004). A closer look at atmospheric processes is needed to understand and model this complex interaction properly. However, the large number of degrees of freedom in the atmosphere complicates a comprehensive (preferably observation based) experimental setup aiming at diagnosing the atmospheric control on land–atmosphere interaction. Findell and Eltahir (2003a, b) introduced a framework in which a single-column model was driven by observed atmospheric profiles and where the sensitivity of convective triggering to soil moisture conditions was evaluated. They presented maps of systematic positive and negative feedbacks over the United States, conditioned on the convective triggering potential (CTP), and a measure of the near-surface dewpoint depression \(\text{HI} \text{low}\). Betts (2004) carried out a range of evaluations of global circulation model (GCM) outputs focusing on the role of LCL, clouds, and radiation on the local land–atmosphere interaction. Also, regional climate models (RegCMs) have been used by a number of authors to assess the sensitivity of the regional hydroclimate to anomalous moisture conditions. Like the work of Findell and Eltahir (2003b), both positive and negative feedbacks are found depending on the governing conditions and processes (Schär et al. 1999; Kanamitsu and Mo 2003; Cook et al. 2006; Fischer et al. 2007; Bisselink and Dolman 2008). Most of these studies consider a number of cases, but a systematic evaluation of the coupling characteristics at multiyear time scales has hardly been carried out.

In this study, we use a variety of model diagnostics to identify the regions and seasonal timing of systematic land–atmosphere interaction in western Africa on a climatological time scale. This area includes regions of strong coupling, as indicated by the work of Koster et al. (2004), and is an ideal test bed for exploring a suite of coupling diagnostics due to strong seasonal and spatial gradients of both land surface and atmospheric processes acting on the moisture budget. Following largely the correlation framework of Dirmeyer et al. (2009), we first identify such regions in a 19-yr hindcast simulation (1989–2007) with a regional climate model. Subsequently, we evaluate the sensitivity of a number of atmospheric key properties (LCL, CTP, and \(\text{HI} \text{low}\)) playing a role in land–atmosphere coupling to soil moisture conditions. Although the model used is far from perfect, it reproduces the observed seasonality in precipitation very well while maintaining sufficient degrees of freedom in the interior to allow realistic land–atmosphere feedback processes. By doing so, we make use of the capability of regional climate modeling in representing land surface and atmospheric processes and their interaction under conditions that are constrained by observed large-scale atmospheric boundary conditions.

In the following, we will detail the diagnostics and frameworks. The modeling setup and the climatological verification are briefly described. After presenting the results on the identified areas and the response of the atmospheric key variables, we will argue that a systematic effect of soil moisture anomalies on local precipitation is strongly constrained by a mixed occurrence of positive and negative feedback processes, and by a strong control of the climatological seasonal cycles of the atmospheric key variables.

2. Definition of diagnostics

For a strong mutual interaction between the surface and the atmosphere, both surface evaporation and precipitation are strongly affected by local moisture conditions. In our setup, we retain two of Dirmeyer et al.’s (2009) three criteria: 1) a collocation of a strong positive anomaly correlation between evaporation and soil moisture and 2) a strong positive correlation between collocated soil moisture and recycling ratio. The criterion involving a strong correlation between evaporation and soil moisture is believed crucial for the potential contribution of soil moisture information to the hydrological predictability at monthly or longer time scales; however, a weak correlation does not preclude the existence of a strong interaction at shorter (daily to monthly) time scales.

a. Correlation between daily soil moisture and evaporation

For this study soil moisture \(W\) is taken as the model value of total water depth accumulated over the top 1 m of soil (see model description below). It thus reflects the
soil water that can be transported to the atmosphere via vegetation transpiration within a seasonal time scale. The correlation between monthly soil moisture \( W \) and evaporation \( E \) is calculated as follows: Daily anomalies of \( E \) and \( W \) relative to their 19-yr means are averaged over each month, and correlations are computed between the time series of these monthly means at each grid element. No smoothing or spatial filtering was applied to these fields. The significance of the correlation is determined by a one-sided \( t \) test (significance level 95%), where the number of degrees of freedom is determined by the number of valid values within a given averaging period (maximum about 90 days \( \times \) 19 years for seasonal averages) divided by a measure of the soil moisture memory time scale. This time scale was defined as the time lag at which the soil moisture autocorrelation drops below 1/e (the e-folding time scale). For each day in the entire record, the autocorrelation was calculated with a backward time lag, and a mean seasonal memory time scale was defined as the average e-folding time scale of all days of each simulation year within that particular season.

b. Correlation between recycling ratio and soil moisture

The recycling ratio was calculated by a “Lagrangian” method (Dominguez et al. 2006), in which for each 6-hourly model output interval with precipitation, the moisture budget of a series of upstream model columns was integrated to estimate the source area of the precipitation. The ratio between \( E \) and the atmospheric moisture column \( q \) is indicative for the degree to which the atmospheric moisture content is changed by local surface evaporation. Integrating \( E/q \) over a trajectory leads to the requested source area. By vertically integrating the product of the horizontal wind and atmospheric moisture content at each model level, a trajectory along which the moisture column \( q \) is advected was constructed, pointing toward the next upwind model column \( x \). The ratio \( E/q \) was integrated over a number of columns until a specified distance \( L \) from the originating column was reached. The recycling ratio \( R \) can then be shown to be given by (Dominguez et al. 2006; Bisselink and Dolman 2008)

\[
R = 1 - \exp \left[ -\int_0^L \frac{E(x)}{\Delta x} \frac{\Delta q(x)}{\Delta t} dx \right],
\]

where \( \Delta t \) is the processing time step (6 h) and \( \Delta x \) the horizontal distance traveled within a time step. Note that Eq. (1) leads to a recycling ratio of 1 when integration is carried out over the entire globe \( (L \to \infty) \), or when the atmospheric column does not move \( (\Delta x/\Delta t \to 0) \). For an infinitesimally small atmospheric column, the contribution from local evaporation to \( q \) approaches 0 and \( R = 0 \).

The procedure was applied to all 6-hourly intervals in a given month to calculate a monthly recycling ratio. Values of \( L \) were varied over a range between 100 and 1000 km. Although the overall value of \( R \) increases with \( L \), the spatial and temporal pattern of \( R \) is fairly insensitive to the choice of \( L \), and values calculated with \( L = 500 \) km were used. This length scale implies that effects of local variations of \( W \) on remote variables affecting \( R \) (i.e., \( E, q, \) and \( \Delta x/\Delta t \)) are included in the correlation between \( W \) and \( R \) for those features that are picked up by the chosen RegCM configuration and resolution (approximately 50 km; see below). As before, a significant correlation between monthly \( R \) and \( W \) was determined by a \( t \) test, with the number of degrees of freedom set by the number of days in the considered period divided by the soil moisture memory time scale.

c. Definition of regions with significant coupling

A seasonal mean correlation was calculated considering all monthly anomalies of \( W \) and monthly values of \( E \) and \( R \). The conventional meteorological seasons [December–February (DJF), March–May (MAM), June–August (JJA), September–November (SON)] were selected for each location, despite that at some locations (for example, the northern edge of the West African monsoon area) the typical annual cycle of succession of dry and wet episodes does not coincide with these time intervals. Areas where both \( W \) and \( E \) and \( W \) and \( R \) show a significantly positive correlation are identified as areas prone to strong land–atmosphere interaction at the climatological time scale. In these areas evaporation is higher when soil moisture is higher, showing a control of soil moisture on evaporation. Negative correlations point at high evaporation despite soil moisture depletion, facilitated by strong surface insolation. Also, a higher recycling ratio for wetter soil conditions points to atmospheric conditions where precipitation is at least partly controlled by the local surface conditions—and not entirely because of remote advection of moisture. A spatial smoothing to the significance levels was applied to yield spatially consistent patterns. Grid points are only considered if for six of the adjacent \( 3 \times 3 \) grid boxes a significant correlation of the same sign is found.

d. Atmospheric properties

For each grid point and each day, the LCL is computed from the daily mean air temperature \( T_2 \) and dewpoint temperature \( T_d \) both at 2-m height. The LCL is defined as the intersection of a dry adiabatic lapse rate from \( T_2 \) and a moist adiabatic lapse rate from \( T_d \).
Because the LCL is closely related to the surface relative humidity (Betts 2004), it generally increases with decreasing soil moisture content. However, this sensitivity is strongly controlled by atmospheric processes affecting the thermodynamic structure of the planetary boundary layer and above.

The CTP is a measure for the atmospheric ability to trigger convection (Findell and Eltahir 2003a). Although convective triggering is not a limiting factor for annual rainfall in the Sahel (Mathon et al. 2002; see also section 2), it is considered a useful diagnostic in this study because it expresses the (possibly surface state dependent) atmospheric ability to generate or sustain convection. CTP is defined as the vertically integrated early morning (0600 UTC) difference between the atmospheric temperature \( T \) and the moist adiabatic lapse rate \( T_m \) in the range between 100 and 300 hPa above the surface. This range is considered to be critical because surface processes may modulate the LCL and the atmospheric boundary layer (ABL) within this range. During most conditions, the ABL will reach 100 hPa above surface; however, 300 hPa above surface is rarely exceeded. The moist adiabat is calculated by assuming a saturated parcel ascent from 100 hPa above the surface and accounting for condensational heating on the parcel’s temperature as it ascends. The energy released when condensation occurs may be an additional source of buoyancy, possibly triggering free convection, depending on the CTP. Negative values of CTP (\( T \) on average exceeds \( T_m \)) indicate a stably stratified atmosphere in which no convection can occur. Very high CTP values are indicative of the likely occurrence of (wet or dry) free convection.

A second variable in the framework of Findell and Eltahir (2003a) distinguishes between wet and dry atmospheric conditions, in relation to the formation of precipitation from convective events. They define a humidity index \( H_{\text{Ilow}} \) as the sum of the dewpoint depression \( (T - T_d) \) at 50 and 150 hPa above the surface. Very high values of \( H_{\text{Ilow}} \) are indicative of dry conditions in which no precipitation will occur because the air is too far away from saturation. For very low values of \( H_{\text{Ilow}} \), the likelihood of precipitation is less dependent on triggers, such as surface conditions or thermal convection.

In their CTP–\( H_{\text{Ilow}} \) framework, Findell and Eltahir (2003a) define a set of conditions in which the soil moisture state is likely to influence the occurrence of convective precipitation. In many cases (CTP < 0, \( H_{\text{Ilow}} \) too low or too high), precipitation is independent of the surface energy partitioning. For relatively low values of \( H_{\text{Ilow}} \) and positive CTP, precipitation is favored over wet soils because the surface provides a source of moist static energy that can convert into convective precipitation. Wetting an already wet soil implies a positive feedback loop. Note that this positive feedback has a different nature than the positive feedback (implied by, e.g., D’Andrea et al. 2006; Bierkens and Van den Hurk 2007), which is based on atmospheric moisture budget considerations rather than on the thermodynamic properties of the atmosphere. Under dryer conditions and higher CTP values, convection is favored over dry soils that force the LCL to reach to higher levels, a negative feedback.

3. Set up and brief verification of the RegCM

The RegCM considered here is an upgrade of the Regional Atmospheric Climate Model (RACMO), version 2.1 (RAMCO2.1; Van Meijgaard et al. 2008). It uses a semi-Lagrangian dynamical formulation and the physical parameterization package of the European Centre for Medium-Range Weather Forecasts (ECMWF), the so-called cycle 31 (ECMWF 2007). It carries the land surface parameterization presented by Van den Hurk et al. (2000), and the parameterized convection is described by Bechtold et al. (2004). Surface evaporation is calculated using a classical Penman–Monteith approach. It thus depends on—among others—available (radiative) energy and available soil moisture. Both canopy transpiration and direct evaporation from bare ground and leaf interception are considered. The convection scheme uses a convective available potential energy (CAPE) closure to determine the vertical mass flux and tests convective triggering by ascending a number of parcels from various levels within the atmosphere.

RACMO2.1 is used extensively for regional climate downscaling (e.g., Lenderink et al. 2007) and evaluating various components of the physical parameterization (e.g., Van Zadelhoff et al. 2007). The simulation used in the present study is a regional downscaling experiment focusing on the West African monsoon area applied in the context of the European Commission–sponsored project ENSEMBLES (Hewitt and Griggs 2004). The domain covers the area (23.32°S, 38.72°W; 38.72°N, 34.76°E) at 0.44° resolution (approximately 50 km) and 40 vertical levels. In all analyses and plots, a boundary relaxation zone of 10 grid points on either side was excluded. The hindcast simulation was carried out using ECMWF interim reanalysis data (Uppala et al. 2008) as lateral boundary conditions for the period January 1989–November 2007. A 3-yr spinup using 40-yr ECMWF Re-Analyses data (Uppala et al. 2005) was applied covering 1986–88. From an evaluation with observed precipitation fields, an apparent wet bias during the spin-up period was reduced considerably upon the introduction.
of ECMWF Re-Analysis (ERA)-interim lateral boundary conditions from 1989 onward.

The quantity $W$ is defined as the total water column in the top 1 m of the soil grid, which contains four levels up to a depth of 2.89 m. Although stratiform precipitation and convective precipitation were calculated separately, only total precipitation is considered in the analysis.

Although a full verification of the RegCM is not the scope of this study, a decent representation of the main hydrological cycle needs to be confirmed. Figure 1 shows a mean seasonal spatial distribution of precipitation compared with the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) data (Xie and Arkin 1997). Evidently both the location and order of magnitude of the seasonal mean precipitation are well reproduced by the RegCM. However, anomaly correlation coefficients between monthly RegCM and CMAP data vary across the land area between 0.2 and 0.6, which implies a poor reconstruction of the year-to-year variability of monthly-mean precipitation (not shown). Anomaly correlations between precipitation from the RegCM and the driving ERA-interim data are higher near the boundaries of the model domain but drop to similarly low values in the West African monsoon area. Apparently, the model has considerable degrees of freedom to adjust its interior hydroclimate away from the forcing boundaries. Therefore, the variability of hydrological quantities (evaporation, soil moisture precipitation) and land–atmosphere interaction, under investigation here, is more sensitive to modeled atmospheric and land processes than to the lateral forcing.

4. Regions of strong land–atmosphere interaction

A strong systematic hydrological feedback between land and atmosphere requires the simultaneous sensitivity of evaporation to soil moisture and of precipitation to atmospheric water content (Dirmeyer 2006). In our set up, areas where these sensitivities occur are indicated by significant correlations between $E$ and $W$ and between $W$ and $R$.

Figure 2 shows the areas with significant correlation between $E$ and $W$ anomalies. In the core of the heavy precipitation area around the equator, the correlations tend to be negative, indicating evaporation to be controlling soil moisture content (by depletion of the reservoir).
rather than vice versa. In the dry Sahara and Kalahari Desert, the correlation is weak, owing to the small variability of either signal. But in a major portion of the domain, significant positive correlations are found. This is in broad agreement with Dirmeyer et al. (2009), who base their findings on offline simulations with an ensemble of land surface models driven by observed precipitation and radiative forcings.

Figure 3 shows a similar plot, except for the correlation between \( W \) and \( R \) using \( L = 500 \text{ km} \). Results are very similar for recycling ratios calculated with other values of \( L \). Here \( R \) is only defined when precipitation is present, which precludes any coherent signal in the deserts.

The strongest and most coherent positive correlation patterns are shown in JJA in the West African monsoon area and in the seasons preceding or following the main wet period: MAM and SON in the Southern Hemisphere and SON in the region south of the Sahel. The JJA monsoon signal is probably reflecting interannual variability in the spatial extent of the West African monsoon. Apparently, surface moisture conditions are rather variable and covarying with the recycling ratio. High recycling ratios during wet conditions imply conditions where precipitation is partly determined by local evaporation: increases in water supply from the surface increase precipitation. Negative correlations indicate wet soil anomalies by rainfall originating from remote areas. This particularly applies to the equatorial tropics in all seasons and to a narrow strip in the Sahel during SON. Again, results are similar to the findings of Dirmeyer et al. (2009), including the alternating negative–positive–negative signal between \( 10^\circ \text{N} \) and the equator in SON.
In Fig. 4 the areas where the correlations shown in Figs. 2 and 3 are significantly positive are indicated per season. Also shown is the seasonal mean soil moisture memory time scale (the $e$-folding time scale of lagged correlations). The simultaneous presence of the two positive correlations is indicative for a systematic influence of soil moisture anomalies on precipitation at seasonal time scales during the 19-yr analysis period. This influence is strongly constrained, and it is basically limited to the northern edge of the West African monsoon in JJA and some scattered areas during the subsequent drying stage in SON and the equatorward edge of the monsoon in the onset periods (MAM in Northern Hemisphere, SON in Southern Hemisphere). The South Africa area is near the boundary of the model domain and within the influence radius of the lateral relaxation of the RegCM to ERA-interim fields, which possibly introduces model tendencies that are difficult to interpret in terms of atmospheric responses to land surface conditions.

In the study of Koster et al. (2004), which was confined to a single JJA season, roughly the same Sahelian area is highlighted as in Fig. 4. They use a coupling strength diagnostic measuring the degree to which precipitation variability is affected by soil moisture anomalies. In Fig. 4 a different metric is plotted, but it is aiming at a similar connection between the surface wetness and precipitation.

Further analyses of the land–atmosphere interaction using RegCM model output concentrate on two contrasting but sensitive areas (see boxes in Fig. 4): 1) the Sahel located between $12^\circ-17^\circ$N and $12^\circ$W–$8^\circ$E and 2) the inland tropical area covering the Democratic Republic of Congo, Chad, Central African Republic, and southern Sudan between $15^\circ-26^\circ$E and $2^\circ-8^\circ$N. The former region shows positive correlations in JJA in 76% of the grid boxes within the domain and somewhat (7%) in the subsequent drying season, whereas the latter region has low correlations in all seasons but MAM (86%).

Figure 5 shows the monthly-mean values of evaporation...
and recycling ratio as function of $W$ averaged for these two areas. Monthly-mean values are labeled and colored by the season. For evaporation the seasonal cycle of the Sahelian area shows a larger amplitude and is shifted in time compared to the tropical domain. Evaporation gradually increases with $W$ in all seasons but saturates at high soil moisture levels in the wet season. However, in the drying stage of the season (SON), evaporation rapidly decreases while the column soil water shows a delayed response, leading to an apparent stronger sensitivity of $E$ to $W$ than in MAM and JJA. A similar plot expressing $E$ as function of soil moisture evolution in the top 7 cm of the model (not shown) does not display this hysteresis, pointing at an evaporation control by a shallow soil water reservoir. In the tropical subdomain, this rapid decline of evaporation does not take place until the boreal winter season (DJF), and little difference in sensitivity $\Delta E/\Delta W$ between JJA and SON is observed.

For the recycling ratio the situation is different. In the Sahelian area, the overall recycling ratio is highest in JJA, and also the sensitivity $\Delta R/\Delta W$ is only clearly positive during the JJA season, allowing a systematic land–atmosphere feedback. In the tropical area, a rapid decline of $R$ from September to November takes place at continuously high values of $W$. In addition, a lack of soil moisture control on evaporation prohibits a strong land–atmosphere feedback in this area during JJA and SON.

The MAM signature in the tropical and Sahelian subdomains shown in Fig. 4 can also be understood from a different phasing of the seasonal cycle of $\Delta E/\Delta W$ in these two areas: although $E$ strongly increases with soil water in the Sahelian spring season, the limited water availability does not allow a sustained evaporation or recycling at the monthly time scales analyzed here. In the tropical area, the wet season has clearly started by the end of MAM, and the recycling ratios are positively affected by additional moisture availability.
5. Key surface–atmosphere relationships in interaction regions

The response of the precipitation features to soil moisture anomalies is governed by both land surface and atmospheric processes. While several studies have explored the sensitivity of precipitation to changing land surface conditions (see references in introduction), only few systematically address the atmospheric conditions that must be met to allow a strong land–atmosphere feedback.

a. Lifting condensation level and convective triggering potential

Among others, Betts (2004) uses the relation between lifting condensation level and soil moisture to diagnose regions of strong interaction while exploring results from the National Centers for Environmental Prediction (NCEP) and ECMWF reanalysis products. A strong sensitivity of the LCL to soil moisture anomalies is at the origin of a number of subsequent feedbacks: boundary layer mixing up to the level of free convection, cloud formation, and shortwave and longwave radiative responses when condensation occurs, or eventually triggering or fuelling of convection.

Figure 6a (top row) shows an aggregated picture of the relation between LCL and soil moisture in both subdomains. A general tendency of lower LCL with higher soil moisture is evident, but both areas expose a similar asymmetric seasonal cycle, with a low LCL and low sensitivity to soil moisture in the major wet season, and a steeply increased LCL and sensitivity in the subsequent drying phase. Thus, in this drying phase, a given soil moisture anomaly has a significantly stronger effect on the LCL than a similar anomaly during the period of wetting. Differences in the atmospheric configuration between these two phases are thus potentially reflected in the seasonal cycle of the sensitivity of precipitation occurrence to the surface conditions.

A higher LCL is also positively correlated to a CTP (Fig. 6a, second row). Both CTP and LCL values are
rather low and not very variable during the core wet season (JJA) in both regions. Larger systematic excursions are shown in both the wetting and the drying seasons. The relationship between CTP and LCL is fairly similar in both seasons in the tropical subdomain; however, in the Sahelian area, lower LCL values are associated with enhanced CTP in the SON season, presumably giving rise to easier convective triggering due to LCL rise.

Figure 6b, showing the relation between the low-level dewpoint depression $H_{\text{low}}$ and soil moisture, reveals an important constraint on strong land–atmosphere interaction during the entire seasonal cycle. For the Sahelian area, $H_{\text{low}}$ in JJA is on average confined to values $<15$ K (consistent with the range found by Findell and Eltahir 2003a; see next subsection) but quickly reaches very high values in the subsequent dry season.
particularly from October onward. Although relatively small soil moisture perturbations could reduce HI$_{low}$ considerably in this steep regime, the atmosphere is often too dry to form precipitation, which implies a strong constraint on the control exerted by soil moisture on the precipitation formation. In the tropical domain, HI$_{low}$ stays well within the “precipitable range” with values of $<15$ K in all seasons but DJF, but the variability in soil moisture conditions in particularly SON is very small because of the persistent precipitation, which also reduces the effect of soil moisture anomalies on regional precipitation to a minimum.

b. CTP, HI$_{low}$, and soil moisture anomalies

In the CTP–HI$_{low}$ framework of Findell and Eltahir (2003a,b), too low or too high values of HI$_{low}$ rule out any surface influence on convective triggering, as the atmospheric structure is overriding any surface anomaly. In a small range of HI$_{low}$–CTP values, surface conditions do matter: Low HI$_{low}$ (between about 5 and 10 K) and positive CTP favor convection over wet surface conditions, implying a positive feedback mechanism. Higher HI$_{low}$ values (between 10 and 15K), and CTP values well above zero are preferred for convection over dry soils. They use a modeling approach in which for a given day, surface wetness conditions are varied and the convective response is explored. Such an experiment could be repeated with a RegCM [by running a small ensemble with variable soil moisture conditions around a reference run, for example, applied by Fischer et al. (2007)], but this experiment was not carried out for the present study.

Instead, the potential role of the soil wetness condition on the atmospheric ability to form precipitation has been investigated by directly relating soil moisture to the CTP–HI$_{low}$ regimes. Assuming the CTP and HI$_{low}$ thresholds proposed by Findell and Eltahir (2003a) to be generally applicable, we can separate the occasions where convection is favored over wet soils from conditions with dry soil advantage or no soil control. Figure 7 shows the probability of wet soil advantage conditions $p_{WSA}$, defined as the relative number of days in every season where CTP > 0 J kg$^{-1}$ and 5 < HI$_{low} < 10$ K. Figure 8 shows the analogous probability $p_{DSA}$ for atmospheric conditions with a dry soil advantage (CTP > 150 J kg$^{-1}$ and 10 < HI$_{low} < 15$ K). Only points are shown where >5% of the days in the indicated seasons are within the indicated regime. Although the general applicability of the chosen thresholds may be disputed, the spatial signature of the results is not very sensitive to the choice of the threshold values. Reducing, for example, the critical value of HI$_{low}$ where a transition from a wet to a dry soil advantage regime occurs from 10 to 8 K obviously decreases $p_{WSA}$ and increases $p_{DSA}$, but these changes occur in the same domains as shown in Figs. 7 and 8. In most of the domain, the atmospheric conditions were either in the wet or dry soil advantage regime for more than 85% of the days when rain occurred (not shown).

In general, wet soil advantage conditions are more frequent than dry soil conditions. The areas where a mixture of wet and dry favoring conditions occurs do coincide near the outermost limits of the wet seasons, and dry soil advantage is barely seen in the deep convective core of the rain seasons. Thus, according to this diagnostic, positive feedback occasions (where convection is triggered over wet soils) are more frequent than negative feedback. In particular, positive feedbacks remain possible in the northern edge of the Sahelian rainfall region in SON, when the major rainbelt is retreating southward. However, outside the tropical rainfall zone, positive and negative feedback occurrences are collocated in the same areas and seasons. This implies that conditions with wet or dry soil advantage do alternate and that a systematic positive or negative effect of soil moisture anomalies on convective triggering is difficult to establish. This introduces an extra constraint for a strong systematic role in soil moisture on precipitation events.

The results show that for CTP > 0, relatively low near-surface dewpoint depressions ($<10$ K) occur more often than higher values of HI$_{low}$. To assess whether this is related to the soil moisture conditions in the RegCM runs, we checked whether the occurrence of precipitation under either wet or dry soil conditions actually matched the soil moisture anomalies at the beginning of the day. A statistical analysis was performed that tested the hypothesis that initial soil moisture conditions for “wet soil advantage precipitation” events $W_{WSA}$ were higher than for dry soil advantage precipitation events $W_{DSA}$. The results (not shown) did confirm that for most areas where $W_{WSA}$ differed significantly from $W_{DSA}$ this difference pointed into the right direction. However, a clear pattern over the seasons or areas was not apparent.

6. Discussion and conclusions

In various recent studies, land–atmosphere interaction has been investigated by a variety of observation- and model-based diagnostics, using statistical relationships between variables (Koster et al. 2004; Dirmeyer et al. 2009; Douville 2009) or physical concepts, such as convection and LCL (Betts 2004; Findell and Eltahir 2003a,b). Statistical measures have the advantage of displaying general features on large spatial scales, whereas diagnostic studies involving physical processes are often limited to regional areas and/or case studies because of data...
constraints. This study uses a RegCM as a pseudo-observational dataset to explore the statistical indicators for land–atmosphere interaction and complements these with an analysis of physical properties of the land–atmosphere system. Although the relations found here depend on the structure and quality of the model components (particularly the parameterizations of evaporation and convection, e.g., Hohenegger et al. 2009), the study provides new insights of the nature of land–atmosphere interaction. We confirm the existence of regions of strong interaction in the northern edge of the West African monsoon in JJA and a number of smaller regions, including a tropical Northern Hemispheric region in MAM. However, from a statistical point of view, the role of surface conditions in the precipitation dynamics is generally not overwhelming, even under conditions where soil moisture exerts a significant influence on surface evaporation. The study provides additional evidence for this limited extent of the surface control on precipitation.

For instance, during the heart of the wet seasons in the Sahel, soil moisture is abundant and hardly constraining evaporation, precluding a strong land–atmosphere coupling. A strong correlation between soil moisture and recycling ratio points to the local origin of moisture for precipitation. In the tropical zones around the equator, moisture advected from the oceans is presumably the primary source of moisture. Away from the equator, significant correlations between soil moisture and the recycling ratio exist during the transition seasons SON and MAM.

The lifting condensation level (LCL) displays a pronounced seasonal cycle in harmony with the migration of the precipitation systems. During the drying stage, the LCL rapidly rises with decreasing soil moisture, pointing to a strong sensitivity of the atmosphere to soil moisture conditions. However, a strong atmospheric sensitivity to soil moisture conditions is found apparent for areas without significant correlations between soil moisture,
evaporation, and the recycling ratio. Thus, other mechanisms must be in place as well to explain a strong atmospheric sensitivity to surface wetness conditions.

From a comparison between two key regions, the relationship between LCL and convective triggering potential (CTP) shows a slightly different seasonal signature. In the region with strong coupling in JJA (northern Sahel), a hysteresis between LCL and CTP appears: during the drying phase, the LCL is lower at a given CTP value than during the wetting season, MAM. This may favor the release of more convective potential energy, providing a link between the soil state and the convection. However, the generally dry state of the atmosphere outside the main wet season forms an extra constraint on the potential role of soil moisture in convective triggering. In the tropical area, the LCL – CTP relation is more symmetric, and the LCL rise in the drying season does not affect CTP differently.

From an evaluation of previous-day soil moisture under conditions of rainfall, it was seen that there are many locations where soil moisture is indeed higher for wet soil advantage conditions than for dry soil advantage. However, this feature is not very coherent. Wet soil advantage conditions—associated with a positive feedback, as precipitation is promoted under wet soil conditions—are more widespread and frequent than dry soil advantage conditions, and they remain present in the northern Sahel during the retreat of the rainfall in SON. The dominant wet soil advantage in this study is in contrast to analyses of convective activity over wet or dry soils by Taylor and Ellis (2006), who find a preference of dry soils for convection. Their analysis is based on satellite observations, which probably pick up relevant processes at much smaller scales than in our study.

In general, the areas where this coupling is strong are fairly small. Precipitation in the tropical convergence zone and the adjacent trade wind regimes are dominated by migrating and growing mesoscale precipitation systems (Mathon et al. 2002), whose dynamics are dominated by large-scale atmospheric features. Local moisture
anomalies may influence the activity or trajectory of these rainfall mechanisms (Taylor et al. 2007) or change the atmospheric circulation affecting the subsequent advection of moisture (Taylor 2008); however, the spatial scale at which this interaction is active is not included in the local diagnostics used in this study. A new set of nonlocal diagnostics needs to be developed that takes the surface control on the initiation, migration, and decay of major mesoscale convective systems into account.

Land–atmosphere interaction is governed by a combination of surface and atmospheric processes. The seasonal cycle of the atmospheric properties is an important modulator of the degree to which hydrological surface anomalies extend affect the precipitation formation process. Combined analysis of both surface and atmospheric processes and variables is necessary to construct a coherent picture of systematic land–atmosphere interaction.

The study was designed to demonstrate a number of diagnostics that in principle could be observed and that do not require model sensitivity runs, like the setup of Findell and Eltahir (2003a), Koster et al. (2004), and Hohenegger et al. (2009). However, further work is needed to verify the RegCM results with true observations. Also, the diagnostic framework could still be developed further. Conditional predictability in cases where systematic land–atmosphere interaction is small but where during individual episodes a strong soil control on the regional hydrological cycle may exist (e.g., Ferranti and Viterbo 2006) should be addressed more carefully. Also, diagnostics that—other than this study—measure the potential influence of anomalous soil moisture conditions on precipitation in remote areas (see, e.g., the case studies by Beljaars et al. 1996; Haarsma et al. 2009) remain a relevant topic for research.

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