Intraseasonal Land–Atmosphere Coupling in the West African Monsoon

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ABSTRACT

Via its impact on surface fluxes, subseasonal variability in soil moisture has the potential to feed back on regional atmospheric circulations, and thereby rainfall. An understanding of this feedback mechanism in the climate system has been hindered by the lack of observations at an appropriate scale. In this study, passive microwave data at 10.65 GHz from the Tropical Rainfall Measuring Mission satellite are used to identify soil moisture variability during the West African monsoon. A simple model of surface sensible heat flux is developed from these data and is used, alongside atmospheric analyses from the European Centre for Medium-Range Weather Forecasting (ECMWF), to provide a new interpretation of monsoon variability on time scales of the order of 15 days. During active monsoon periods, the data indicate extensive areas of wet soil in the Sahel. The impact of the resulting weak surface heat fluxes is consistent in space and time with low-level variations in atmospheric heating and vorticity, as depicted in the ECMWF analyses. The surface-induced vorticity structure is similar to previously documented intraseasonal variations in the monsoon flow, notably a westward-propagating vortex at low levels. In those earlier studies, the variability in low-level flow was considered to be the critical factor in producing intraseasonal fluctuations in rainfall. The current analysis shows that this vortex can be regarded as an effect of the rainfall (via surface hydrology) as well as a cause.

1. Introduction

The state of the land surface has the potential to influence large-scale circulations (Charney 1975). Variations in rainfall influence soil moisture and vegetation, which in turn affect the partition of surface heat and moisture fluxes into the atmosphere. If the large-scale circulation is sensitive to surface fluxes, and the surface responds to climate anomalies with sufficient amplitude and spatial coherence, conditions for feedback may be met. Monsoon circulations potentially provide these conditions, driven by gradients in ocean and land surface fluxes, with the transport of oceanic moisture into the continent sensitive to land surface conditions (e.g., Eltahir and Gong 1996). It follows that where monsoons exhibit strong intraseasonal rainfall variability, as in India, there is the potential for intraseasonal feedbacks between soil moisture and rainfall (Webster 1983; Ferranti et al. 1999).

Intraseasonal variability in the West African monsoon has been a focus of considerable interest in recent years. Monthly rain gauge data indicate a gradual northward advance of the intertropical convergence zone (ITCZ) over the region during the summer months, with maximum rainfall occurring in August. However, from daily data, Sultan and Janicot (2000) and Le Barbé et al. (2002) identified an abrupt northward shift in the ITCZ from 5° to 10°N, which occurs each year typically in late June. This “monsoon onset” signals the beginning of the wet season proper in the Sahel (12°–18°N in the current study). The onset has been linked to various factors, notably enhanced advection of moist air into the continent associated with oceanic processes (Gu and Adler 2004), the dynamics of the Saharan heat low (Sultan and Janicot 2003; Sijikumar et al. 2006), topography (Drobinski et al. 2005), and surface albedo (Ramel et al. 2006). Intraseasonal fluctuations in West African rainfall throughout the seasonal cycle have been the subject of a number of studies (Janicot and Sultan 2001, hereafter JS01; Grodsky and Carton 2001; Sultan et al. 2003; Matthews 2004; Mounier et al. 2008). Mounier and Janicot (2004) identified two independent intraseasonal modes of convection from analysis of satellite outgoing longwave radiation (OLR) data. These consist of a quasi-stationary
mode linking the Guinea Coast with the equatorial Atlantic (the “Guinean” mode), and a westward-propagating mode centered farther north around 15°N (the “Sahelian” mode). The dynamics of the Sahelian mode were explored by JS01 and Sultan et al. (2003) using a combination of rain gauge data, atmospheric reanalyses from the National Centers for Environmental Prediction (NCEP), and OLR imagery over a 22-yr period. They found coherent variations in rainfall and winds with a periodicity of around 15 days. Compositing filtered data around maxima and minima in Sahelian rainfall, they identified a vortex at 925 hPa, which propagated westward at about 4° longitude per day with the rainfall feature. They showed that the rainfall fluctuations were consistent with forcing by the vortex through modulation of the monsoon flow into the Sahel.

Several of these studies have considered variations in land surface heating as playing a role in West African intraseasonal variability. Grodsky and Carton (2001) identified variability on the 10–15-day time scale in the zonal wind off the coast of West Africa and found a strong correlation in the premonsoon period (April–June) with rainfall to the south of the Sahel (essentially the Guinean mode of Mounier and Janicot (2004)). They proposed a feedback cycle that was consistent with variations in surface air temperature and pressure in reanalysis data. Enhanced monsoon winds led to increased cloud and rainfall in the interior, suppressing surface heating via reduced insolation and increased evaporation, and raising the surface pressure such as to reverse the zonal winds. Similarly, Mounier et al. (2008) linked changes in surface temperature and pressure associated with the Guinean mode with variations in cloud cover over the continent, though they did not speculate on the role of changes in the partition of insolation between sensible and latent heat.

There are significant uncertainties in the simulation of land surface fluxes in atmospheric reanalyses resulting from the poor representation of key drivers, notably cloudiness, precipitation, and soil water. Furthermore, direct observations of surface fluxes are unrepresentative at the larger scale because of variations in soil and vegetation properties and atmospheric forcing (notably rainfall and insolation). To better understand the extent to which the West African (and other) monsoon circulations are sensitive to surface fluxes, it is therefore important to exploit alternative observational proxies for regional-scale surface fluxes. In particular, one would wish to know whether climate-induced intraseasonal fluctuations in fluxes were of sufficient magnitude and spatial extent to feed back on the monsoon circulation, and thus influence subsequent rainfall, as proposed in studies already mentioned.

In the context of West Africa, the sparseness of the vegetation in the Sahelian zone makes surface fluxes particularly sensitive to variations in antecedent rainfall in this region. In the day or two after rainfall, soil water in the top few centimeters of the soil is readily available for evaporation directly from the soil surface (e.g., Gash et al. 1997; Wallace and Holwill 1997). The resultant high latent heat flux from a wet soil is accompanied by reduced sensible heating. The atmosphere responds directly to soil moisture forcing during the daytime through changes in the temperature, humidity, and depth of the planetary boundary layer (PBL). The daytime winds within the PBL are weak (Parker et al. 2005) and surface heat fluxes dominate the evolution of the daytime PBL heat budget. Measurements from the special observing period (SOP) of the African Monsoon Multidisciplinary Analyses (AMMA; Redelsperger et al. 2006) demonstrated how the spatial pattern of antecedent rainfall and soil moisture map rather closely onto variations in the properties of the overlying PBL (Taylor et al. 2007). Indeed, the mesoscale soil moisture gradients typically found in the Sahel were shown by Taylor et al. (2007) to be strong enough to induce a sea-breeze-type response during the afternoon. After dark, when dry convection switches off, the winds intensify in response to the pressure field generated by the pattern of daytime heating (Racz and Smith 1999). If significant soil moisture contrasts exist at the synoptic scale (as shown by Taylor et al. 2005), then one might expect an anticyclonic circulation to develop overnight in the region of the wet soil. This effect would be maximized by dawn the following morning.

The aim of this study is to reexamine the Sahelian mode of intraseasonal variability identified by JS01 from the perspective of soil moisture. In particular, the paper explores whether the 15-day rainfall signal produces a spatially coherent soil moisture response, which might feed back on the development of low-level vorticity via surface heating. Passive microwave retrievals from satellite provide a characterization of near-surface soil moisture, and their use is introduced in the next section. Section 3 presents characteristics of the soil moisture variability and its relationships with atmospheric temperature, as depicted by atmospheric analyses. A simple model of surface heating based on satellite soil moisture data is developed in section 4, and used to make quantitative estimates of the dynamical response of the atmosphere to soil moisture. These estimates are then compared with the intraseasonal variability in atmospheric analysis data.
2. Data

Passive microwave imagery from a satellite provides a method of detecting soil moisture (e.g., Jackson 1993). At low microwave frequencies, the ratio of horizontally to vertically polarized brightness temperatures is sensitive to moisture in the top few centimeters of the soil. The signal is attenuated by the presence of a vegetation layer, but in sparsely vegetated regions such as the Sahel, the soil moisture signal is strong. In this study, data at 10.65 GHz from the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI; Kummerow et al. 1998) are used to qualitatively infer soil moisture. TRMM data have been used by Bindlish et al. (2003), among others, to estimate soil moisture in the top centimeter of the soil. The satellite typically provides data once per day, and this is gridded spatially at 0.25° resolution in 6-h time steps. In the current study, Sahelian wet seasons (1 June–30 September) from 9 yr (1998–2006) are analyzed over a domain of 10°–30°N, 17°W–40°E. The approach is the same as that described in Taylor and Ellis (2006) and involves calculating a polarization ratio anomaly (Δpr) relative to a characteristic “dry” value (Fig. 1). The dry background value captures the small-amplitude trend in polarization during the season resulting from the developing vegetation layer. Figure 1 illustrates the sharp drop in polarization ratio after rainfall, followed after several days without further rain, by a relaxation back to the dry value as the near-surface soil dries. The signal is insensitive to cloud, though pixels where rain is falling are identified and removed from the analysis using an algorithm demonstrated by Ferraro et al. (1998).

Analysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF) are used to characterize aspects of intraseasonal atmospheric variability for the 9 yr of available satellite data. Previous studies (e.g., Thorncroft et al. 2003) have found that the dynamic fields depicted by ECMWF analyses are reliable in this region, benefiting from assimilation of radiosondes and synoptic station data. On the other hand, the representation of soil moisture and surface fluxes in the analyses are not considered to be realistic. In this study, 40-yr ECMWF Re-Analysis (ERA-40) data are used for the first four wet seasons (1998–2001), and operational analysis data subsequently. All data have been regridded at 1° resolution.

Two further satellite datasets are used. Cloud-top temperatures from Meteosat thermal infrared imagery provide an indication of deep convection. Within a 1° grid box, the percentage of Meteosat pixels (−0.05° × 0.05° resolution) below a threshold temperature (set to be −40°C) is used to compute the fractional cold cloud coverage. The seasonal evolution of vegetation leaf area index is characterized by 8-day data from the Moderate Resolution Imaging Spectroradiometer (MODIS) Terra satellite, available since 2000 and obtained from the National Aeronautics and Space Administration (NASA) Langley Research Center Atmospheric Science Data Center. Finally, the nonrecursive filter used in JS01 is applied to the microwave, cold cloud, and atmospheric analyses datasets. This bandpass filter retains periodicities between 10 and 60 days.

3. Linking soil moisture and atmospheric temperature

a. Seasonal evolution

Aspects of the seasonal evolution of the land–atmosphere system are depicted in Fig. 2 for a typical year in the Sahelian zone, averaged over 20° longitude. The year chosen is 2006, and covers the wet season AMMA SOP. The cold cloud cover (Fig. 2a), taken as
a proxy for rainfall, exhibits marked variability on a range of time scales. The daily data show strong fluctuations linked to synoptic-scale modulation of mesoscale convective systems (MCSs). At the seasonal time scale, the northern extent of the cloud field moves to its maximum latitude around mid-August, consistent with peak Sahelian rainfall in that month in the long-term climatology. Superimposed on these features there are some intraseasonal fluctuations evident even in the daily data, most notably around the time of the monsoon onset (14 July). There is little penetration of convection into the northern Sahel (north of 15°N) prior to this time.

Figure 2b shows the daily polarization ratio data $\Delta_{pr}$ averaged over the same region. Note that while individual pixels are not sampled every day by TRMM, within the region there are data available on every day to compute an areal average polarization ratio anomaly.

The data exhibit strong day-to-day variability linked to fluctuations in the cold cloud cover. When the daily signal is filtered out, intraseasonal variability in soil moisture emerges, with a period of the order of 15 days. Such a signal is to be expected as rainfall dominates the surface moisture budget, and (as highlighted by JS01) there are important variations in rainfall at this time scale.

The vegetation develops rapidly during the wet season, as depicted in Fig. 2b by the leaf area index. Once established, the vegetation cover suppresses the soil moisture contribution to surface polarization detected by satellite. This effect contributes to a weakening in the intraseasonal fluctuations of $\Delta_{pr}$ later in the season, and explains the poor translation of cold cloud features (and presumably rainfall) into $\Delta_{pr}$ in the southern Sahel. Also of note are hints of intraseasonal fluctuations in leaf area superimposed on the growing cycle, although the 8-day data are too infrequent to effectively capture such variability.

Figure 2c explores the links between intraseasonal soil moisture variability and atmospheric temperature at 925 hPa. This is diagnosed from the 6-hourly ECMWF analyses at 1800 UTC (same as local time in this region) in order to capture the impact of a full day of surface heating on a substantial depth of the lower atmosphere. The sensible heat flux is suppressed when the surface is wet because of the availability of near-surface moisture for evaporation. Hence, the PBL should be cooler above wet soil. Figure 2c shows that periods of enhanced surface moisture tend to occur at the same time as lower temperatures. The correlation between the two datasets is not perfect and, in particular, the latitudinal extent of the temperature anomalies is greater than that of the surface data.

b. Event scale

A negative correlation between filtered soil moisture and PBL temperature is a necessary but not sufficient condition to indicate a feedback. Low PBL temperatures may coincide with high soil moisture for other reasons, notably enhanced cool southerly flow, and downdrafts from the precipitating systems. To explore the temporal nature of the link further, a subset of the data are composited according to the occurrence of rain events, as identified by a sharp decrease in observed surface polarization. A wet event is defined for a TRMM pixel when a midday value of $\Delta_{pr}$ falls below a threshold of $-0.03$, which is the same threshold used by Taylor and Ellis (2006). In addition, the pixel must have not experienced a wet event in the previous 5 days. When more than 25% of the TRMM pixels within a 1° ECMWF grid cell satisfy these criteria, the data are included in the “wetting event” composite for that
ECMWF grid cell. The results presented are rather insensitive to the threshold values chosen.

The results of this compositing are shown in Fig. 3a, based on 777 events over 9 yr in a region of 12.5°–17.5°N, 15°W–15°E. The composite mean sequence of 

Δp indicates minimum values (i.e., maximum soil moisture) around time 0 (midday on day 0). Composite mean cold cloud cover (white bars; %) at a threshold of −40°C. (b) Temperature anomalies at 925 hPa when all cases are included in the composite [as in (a)], and when only those cases where modeled soil moisture increases by at least 1 mm are included.

ECMWF grid cell. The results presented are rather insensitive to the threshold values chosen.

The composite time series indicates that cool temperatures in the ECMWF analyses lag wet events, and not the other way around. The composite temperature sequence differs significantly (at the 95% level according to a Student’s t test) from the mean diurnal cycle for several days, beginning approximately at the time of the observed maximum in convection. The ECMWF model may not capture individual events particularly well, and indeed there is a strong dry bias in the northern Sahel (Hagemann et al. 2005). Nevertheless, when averaged over many observed wet events, the temperature structure of the lower atmosphere reveals a physically plausible response to rainfall. There is cooling at 925 and 850 hPa overnight between days −1 and 0, consistent with downdraughts from the convective event. During the daytime, once the soil is wet (days 0–2, and to a lesser extent on day 3), the PBL cools relative to the mean diurnal cycle. On the other hand, overnight on days 0–3, there is a warming tendency within the PBL back toward the mean state. This implies that the low temperatures in the days after rain are maintained by daytime processes, that is, weaker-than-normal surface heating.

Advection of cooler air from the south may also influence the low-level heat budget, most notably at night when the monsoon winds are strong and any surface effects on the PBL heat budget are small. There are significant southerly wind anomalies at 925 hPa of up to 2 m s⁻¹ on days −2 and −1 in the composite (not shown), which may account for the overnight cooling at 925 hPa on day −1. However, variability in heat advection is unlikely to play a role in the low temperatures after the rain, when southerly flow is close to the mean. Considering the PBL moisture budget on the other hand, the strengthened monsoon flow on days −2 and −1 may be critical for providing sufficient low-level moisture for the convection to occur.

Figure 3b presents the 925-hPa temperature signal from an additional composite in which, of the original 777 events, only cases where the ECMWF model simulated an increase in top-level soil moisture of at least 1 mm were retained. This second composite only contains 113 events, confirming that the model does not capture many of the observed wet events. However in this subset, the post-rain temperature response is approximately twice as strong as in the larger, drier composite. This suggests that because the model does not simulate many of the wet events, the analyses underestimate the temperature response of the PBL to soil moisture in the original sample of 777 events. If the model does not simulate accurately the dominant process (rainfall), what mechanism produces the temperature signal? A full answer to this question would re-
quire a detailed analysis of diagnostics from the ECMWF model. However, it seems likely that the assimilation of observational data implicitly provides information on daytime heating rates. Prior to the enhancement of the radiosonde network funded by the AMMA program (Parker et al. 2008), there were only six radiosonde stations in the region used to construct this composite (approximately 2 000 000 km²), which sent data to operational centers such as ECMWF. By contrast, there are over 50 synoptic stations, the large majority of which transmit data more than twice a day. At ECMWF, temperature and relative humidity data from synoptic stations are used to nudge the soil moisture in their analysis, using a radius of influence of 300 km around the stations. In addition, surface pressure data from the synoptic stations are assimilated directly into the atmospheric model, and thus provide information to constrain the low-level temperature field.

4. Impact of soil moisture on monsoon dynamics

a. A simple model of sensible heat flux

In the previous section, qualitative links were made between surface polarization observations (a proxy for soil moisture) and atmospheric temperature in ECMWF analyses. To assess whether intraseasonal variations in soil moisture are of sufficient amplitude and spatial coherence to drive the fluctuations in Sahelian PBL temperature (and thereby influence the dynamics of the monsoon) requires a quantitative relationship between the polarization observations and surface heating. A simple model is therefore developed here. Estimates of surface heat fluxes, expressed as anomalies from a flux over a dry surface, are derived solely from the satellite retrievals of polarization, based on simple assumptions of the surface physics. No explicit account is taken of geographical variations in soil and vegetation properties, though the suppression of the polarization signal by denser vegetation provides an implicit sensitivity to leaf area.

The observations are gridded in time to the nearest 6 h, and a sensible heat flux anomaly is calculated for every available observation $\Delta H_{pr}$ at observation time step $i$ according to

$$\Delta H^i = H_0[1 - \exp(-\Delta H_{pr}/K_{pr})].$$  \hspace{1cm} (1)

A value of $H_0 = -100$ W m$^{-2}$ is assumed to represent the difference in daytime mean heat flux between dry and saturated conditions. Because the observations are irregular in time, and because after rain the polarization ratio relaxes back to dry values more rapidly than the surface energy balance, a second equation is required to capture the temporal evolution of the heat flux at a 6-hourly time step. This evolution is based on empirical knowledge of bare soil evaporation; typically for 1 day after heavy rain, evaporation rates are close to potential, before decaying rapidly as the soil dries (e.g., Daamen and Simmonds 1996; Wallace and Holwill 1997). Thus, the modeled heat flux is held constant for a period $\tau_1$ (=1 day or 4 time steps). Subsequently, at model time step $j$, the flux anomaly associated with that observation ($\Delta H^j$) is assumed to decay exponentially in time as

$$\Delta H^j = H_0[1 - \exp(-\Delta H_{pr}/K_{pr})] \exp[-(j - i - \tau_1 + 1)/K_1]$$

$$i + \tau_1 = j. \hspace{1cm} (2)$$

In a final step, the influence of all preceding observations on the heat flux at model time step $j$ is accounted for by taking the most negative value of (2):

$$\Delta H_j = \min_{1 \leq i \leq j}(\Delta H^i). \hspace{1cm} (3)$$

Values of $K_1 = 1$ day and $K_{pr} = -0.03$ are adopted to capture the plausible evolution of fluxes after rainfall. Heat flux anomalies are presented in Fig. 4a, estimated as a function of polarization ratio and time since the observation. The inset presents an example of a sequence of measured polarization values [converted to estimated flux anomalies via (1)] and their temporal interpolation using (2) and (3).

The heat flux model contains parameters for which values have been assumed a priori. Choices for $H_0$ and $K_{pr}$ will affect the estimated fluxes, and values for these parameters are not readily identifiable from the literature. Thus, in Fig. 4b, an illustration of the day-to-day evolution of sensible heat anomalies estimated from the above equations is presented in comparison to in situ eddy correlation observations. The observations were made at three sites during the wet season of 2005 as part of the AMMA extended observing period (F. Timouk et al. 2008, manuscript submitted to J. Hydrol.) in the Gourma region of Mali over contrasting land cover. Such a comparison requires careful interpretation. The measured fluxes depend strongly on vegetation cover and soil properties. Thus, there are large monthly mean differences between the three sets of observed heat fluxes, primarily due to contrasts in tree and grass cover. In addition, the sites are located several tens of kilometers apart and thus do not receive the same atmospheric forcing, a particular problem for rainfall. On the other hand, the model estimate based on satellite data has coarse spatial resolution (of the order of 50 km) and no explicit sensitivity to land cover.
The heat flux anomalies estimated from satellite (and shown here relative to an arbitrary value of 300 W m\(^{-2}\)) indicate a sequence of nine major rain events during June and July for that region, each resulting in reductions of heating in excess of 75 W m\(^{-2}\). There is a tendency for these to coincide with marked negative departures in observed \(H\) at one or more sites (4 and 21 June, and 10, 14, 21, and 31 July), followed by a recovery to higher values over several days. The reductions in measured daytime heat flux can exceed 100 W m\(^{-2}\) (i.e., larger than the maximum from the model, as defined by \(H_0\)), though these extreme values tend to coincide with reduced insolation (not shown). The model does not represent the impact of cloud cover on heat flux, which can be significant when storms occur during daylight hours. Furthermore, the temporal resolution of the satellite data means that the model can sometimes “miss” rain events.

A visual comparison of the time series with the flux observations suggest that when averaged over many storms, the model predicts variations in heat flux that are plausible, in terms of both their amplitude and their evolution in the days after rain. Simulations using alternative values of the sensitivity parameter \(K_{pr}\) have been performed to identify realistic bounds that this parameter can take. The number of days the model predicts negative flux anomalies exceeding 75 W m\(^{-2}\) is 15 when \(K_{pr}\) is set to \(-0.02\), and 1 when \(K_{pr}\) equals \(-0.08\). In the former case, the model is overestimating the heat flux variations, and in the latter, the model predictions are underestimates. The sensitivity of the overall results of the study to this range of values for \(K_{pr}\) will be assessed in section 4c.

b. A simple model of the atmospheric response to daytime heating patterns

The impact of variations in surface heating on the daytime warming of the PBL can be estimated with a simple heat budget calculation over a depth \(z\). If one assumes for simplicity that there is no change in the vertical heat flux at height \(z\), the anomalous surface contribution to daytime warming is

\[
\frac{\partial T}{\partial t}_{surf} = \frac{\Delta H_z}{\rho c_p z},
\]

where \(\rho c_p\) is the heat capacity of the layer and \(T\) is the PBL temperature. The layer depth \(z\) is assumed constant for simplicity, set to a value of 1.5 km, which is a typical wet season PBL depth. In reality the PBL depth may vary substantially with soil moisture, as found in observations during the AMMA SOP (Taylor et al. 2007). The heating rate from (4) can now be compared directly to the daytime warming in the ECMWF analyses at 925 hPa.

As discussed in the introduction, the daytime generation of synoptic-scale PBL temperature and surface pressure gradients may produce a significant dynamical response overnight. To gain an estimate of the strength of soil moisture forcing on the dynamics of the system, the daytime heat fluxes modeled in the previous section are used to quantify changes in the surface pressure field, in the absence of advection. Considering a low-level layer bounded by two pressure levels, under hydrostatic balance surface heating will affect the thickness and temperature of the layer such that the relative change in thickness equals the relative change in temperature of the layer (i.e., \(\Delta z/z = \Delta T_{surf}/\bar{T}\)). Assuming that the change in thickness occurs equally at the top and bottom of the layer, so that the upper pressure level ascends and the lower pressure level descends by an
amount equal to half of the thickness change, then the change in hydrostatic pressure at the surface becomes

\[ \Delta p_{\text{surf}} \sim -\rho g \left( \frac{z \Delta T_{\text{surf}}}{2T} \right) = -\frac{g \Delta H \Delta t}{2 \rho \bar{T}}. \]

where time \( \Delta t \) is 12 h, the period over which the heat flux anomaly acts. An order of magnitude estimate of the impact of the resulting pressure field on the vorticity can then be gained from assuming a geostrophic response,

\[ \zeta_{\text{surf}} = \frac{1}{\rho f} \left( \frac{\partial^2 \Delta p_{\text{surf}}}{\partial x^2} + \frac{\partial^2 \Delta p_{\text{surf}}}{\partial y^2} \right). \]

Equation (6) provides an estimate of surface-generated vorticity, which can be compared with the total vorticity in the ECMWF model. This is diagnosed at 0600 UTC, when the dynamics have had 12 h to respond to the daytime heating field.

c. Composited intraseasonal variations in surface heating

The study of JS01 presented the evolution of the surface rainfall and wind field at 925 hPa based on a composite of intraseasonal fluctuations in Sahelian rain. They identified a cyclonic vortex in the composite, which they argued to be the cause of the rainfall variability. A similar method is applied here, but instead of compositing on rainfall, surface heat flux is used to define the timing of the intraseasonal signal. The aim is to explore whether the vorticity can also be interpreted as a response to the rainfall, via soil moisture and surface heating. The calculations are performed using all 9 yr of satellite data and are analyzed in a pair of wet and dry composites defined by minima (wet) and maxima (dry) in filtered surface heat fluxes averaged over a 12.5°–17.5°N, 2.5°W–2.5°E subdomain (shown as a box in Fig. 5).

The spatial structure of the wet-minus-dry composite at day 0 (i.e., minimum minus maximum heating) is shown in Fig. 5. The strong contrast in surface heating is centered approximately over the compositing box and coincides with a larger-scale heating minimum across the Sahel. This heating pattern is well correlated with a pattern of negative PBL temperature at 1800 UTC in the ECMWF data. The similarity of the two patterns supports the temporal analysis in section 3, namely, that the intraseasonal variations in PBL temperature are qualitatively consistent with soil moisture forcing. It should, however, be noted that the strongest temperature anomalies at 925 hPa are located approximately 250 km to the north of the maximum surface forcing. Also shown in Fig. 5 is the wet-minus-dry composite wind field at 925 hPa at 0600 UTC the following day. An anticyclonic circulation is apparent, which is maximized at low levels (not shown). This cool high anomaly closely resembles the anticyclonic vorticity structure in the days following a rainfall maximum identified by JS01 using alternative datasets and compositing.

In Fig. 6, the impacts of soil moisture forcing on daytime PBL warming and geostrophic vorticity are com-
pared quantitatively with the total warming and vorticity diagnosed from ECMWF data, averaged over the compositing region. The estimated surface heating effect is approximately in phase with and of the same order of magnitude as the total daytime heating rate, as represented in ECMWF analyses (Fig. 6a). This implies that the soil moisture forcing is indeed strong enough to influence intraseasonal fluctuations in PBL temperature. Similarly, the geostrophic vorticity forced by the soil moisture is of a notable magnitude, and its evolution resembles that of the total vorticity diagnosed at 0600 UTC in the ECMWF model (Fig. 6b).

The shading in Fig. 6 indicates the range of heating rates and vorticity obtained by varying the heat flux sensitivity parameter $K_{pr}$ between its realistic upper and lower bounds. This band is narrow relative to the amplitude of the signal, and implies that the above conclusions can be drawn for all plausible values of the sensitivity parameter. This suggests that the impact of the space–time variations in rainfall forcing on the surface dominates over any uncertainty in the simulation of heat flux. All the same, the level of quantitative agreement in Fig. 6 is somewhat surprising, given the simplicity of the assumed atmospheric response to heating patterns. In particular, the model effectively assumes that there is no horizontal diffusion during the day, and that the atmosphere reaches geostrophic balance in response to this forcing. These assumptions tend to exaggerate the effect of the surface on the vorticity field. On the other hand, the surface-forced vorticity estimate is due to only 1 day of heating; the effects of previous days of heating are neglected. In that sense, the total surface impact on vorticity is underestimated.

These calculations demonstrate that not only does the temporal and spatial pattern of surface heating qualitatively match that required to induce the intraseasonal vortex, but the amplitude of the forcing is also consistent. Thus, intraseasonal soil moisture variability appears to play a key role in the development of the vorticity anomaly identified by JS01 as the cause of the rainfall fluctuations. That is, the land–atmosphere system is closely coupled. According to Fig. 6b, this surface hydrological feedback provides an explanation for the lag of 2–3 days between maximum rainfall and maximum anticyclonic vorticity.

d. Space–time evolution of the surface and atmosphere

The evolution of the soil moisture anomaly and its associated vortex across North Africa is presented in Fig. 7. This shows an intraseasonal sequence of anomalies in surface heating, atmospheric temperatures, and vorticity based on the composite developed above. A similar sequence was presented by JS01 showing rainfall and 925-hPa winds, though the structures shown here lag those of JS01 by 2–3 days. At day −7 (Fig. 7a), the target box is characterized by a dry anomaly and hence enhanced heating of the PBL by the surface. This is accompanied by a heat low circulation centered roughly 200 km to the north of the surface feature. These features are essentially the converse of those depicted in Fig. 5 at day 0. At the larger scale, there is a
significant enhancement of cyclonic vorticity over North Africa, with northerly flow close to the Atlantic coast and southerly flow near the Red Sea. Two days later (day −5) the large-scale southerlies have triggered convection (identified from surface wetting) in the mountainous region of Darfur (12.5°–17.5°N, 20°–25°W). Negative temperature anomalies develop around 15°N, consistent with the weak surface heating.
there. After a further 2 days, the minimum in atmospheric temperature has intensified and spans the region from Darfur to the Greenwich meridian. This feature is accompanied by an elongated surface heating minimum, with two prominent centers. The wet surface signal is noticeably weaker over Chad, including Lake Chad and the desert region to the north. The lack of soil moisture in the desert is due to the very arid conditions, where few long-lived convective systems occur, while in the lake region, daily variability in rainfall will have little impact on surface moisture availability.

In subsequent days, the temperature feature continues to propagate westward, but starts to extend farther north in the process. The location of the temperature anomaly to the north of the surface heating pattern at this phase of the cycle may reflect the role of advection by the background southwesterly flow. By day +3, the surface heating feature is located over Mali and Senegal, and is less spatially coherent than in earlier phases. It is unclear whether the weakening of the signal to the west of the Greenwich meridian is either a physical feature, for example, resulting from the dominance of easterly wave dynamics in that region, or simply an artifact of the compositing process. Finally, by day +7, dry conditions have returned to the target box, accompanied by a local heat low there.

The sequence agrees qualitatively with the composite shown by JS01 using independent datasets (NCEP reanalyses and rainfall data for the 1968–90 period). The cool high develops at 20°–25°E and propagates westward at a speed of the order of 4° longitude per day. Essentially the same sequence is captured when different target boxes in the Sahel are used to define the composite (not shown). A strong feature develops in Darfur and propagates westward. The evolution of the composite is summarized in Fig. 8, averaged over Sahelian latitudes. A relationship between the anomalies in meridional wind and convection is readily apparent in Fig. 8a. Maxima in cold cloud tend to follow maxima in southerly flow with a lag of the order of 1 day. This is consistent with the model of JS01 whereby moisture advection associated with the low-level vortex controls subsequent convection. In Fig. 8b, the impact of that convection on surface heating is illustrated. A pattern of strong negative heating anomalies caused by the convection starts over Darfur on day −5 and crosses the continent at a rate of 4°–5° day−1. As anticipated from the above results, anomalies in atmospheric temperature are well correlated with the surface heating pattern. Between 20° and 10°E, there is no intensification of the composite cool high anomaly. This is consistent with a lack of strong surface forcing in that desert region, as discussed above.

From the above analysis one can construct the following plausible sequence of events: Favorable low-level wind conditions in Sudan produce a wet spell there associated with convective systems triggered by topography in Darfur. The resulting soil moisture contributes to a cool high, favoring southerly flow to the west and northerly flow to the east. In the absence of further rainfall, such a temperature anomaly might be expected to propagate westward as a Rossby wave, because of the strong meridional temperature gradient (Taylor et al. 2005). However, the induced meridional flows also influence rainfall via moisture advection, triggering a westward shift in rain, soil moisture, and PBL temperature. Both diabatic and adiabatic processes favor westward propagation, counter to the low-level flow. In reality this rather simplistic model, based on bandpass-filtered data, will be complicated by fast-moving MCSs and easterly waves, with their own time and space scales.
5. Discussion

The use of passive microwave data from a satellite provides an accurate picture of land surface variability, which is missing from atmospheric analyses. By exploiting satellite data to scale up surface fluxes, this study has provided strong evidence to support earlier findings by (among others) Webster (1983) and Grodsky and Carton (2001) of land–atmosphere monsoon feedbacks. The recent observational study of Taylor et al. (2007) confirmed earlier modeling studies, showing the relationship between mesoscale soil moisture features, PBL temperature, and atmospheric dynamics in the Sahel. The current study shows that the correlation between soil moisture and temperature also extends to larger time and space scales in this region. Using simple assumptions about the impact of soil moisture on sensible heat flux, and the response of the atmosphere to variability in that heating, it has been demonstrated that intraseasonal soil moisture fluctuations are intense enough to feed back on the low-level vorticity structure, which produces the initial rainfall variability.

The frequency of the variability studied here is not linked simply to the surface drying time scale, as suggested in the model of Webster for the Indian case. In the Sahel, the dominant evaporative response is that of bare soil with a time scale of only a few days. In the atmosphere, that time scale corresponds more closely to African easterly waves, where similar soil moisture feedbacks may play a role (Taylor et al. 2005). Further work is needed to assess whether slower intraseasonal modulations in root zone soil moisture and leaf area also affect the monsoon dynamics. Such processes may be more important further south in the monsoon region, where vegetation is denser.

A more complete depiction of the atmospheric response to soil moisture is needed in order to fully understand the nature of the coupling, and its interactions with other key sources of variability. This requires climate model simulations to assess the impact of soil moisture patterns on the dynamics of the monsoon. Such patterns could be imposed as boundary conditions in order to identify the vorticity and rainfall response. Equally, it would be useful to investigate how well climate models capture intraseasonal fluctuations in soil moisture and fluxes when the land and atmosphere are fully coupled, and subject to external forcing such as the Madden–Julian oscillation (Matthews 2004). An important question raised by this and other studies (Grodsky and Carton 2001; Mounier et al. 2008) is the relative importance of cloud and surface hydrological controls on sensible heat fluxes over land, and their importance relative to ocean-forced variability. The observations and analysis from the AMMA project should help to answer this question for the West African case, and provide some insight into coupling processes in other monsoon regions.

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