On bimodality in warm season soil moisture observations

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[1] It has recently been suggested that the bimodality in warm season soil moisture observations in Illinois is evidence of a soil moisture-precipitation feedback. Other studies however provide little evidence for a strong feedback in this region. Here we show that seasonality in the meteorological conditions in combination with the non-linearity of the soil moisture response alone can induce this bimodality. The existence of preferred wet and dry soil moisture states may have implications for the understanding and modeling of soil moisture dynamics in mid-latitude regions. **Citation:** Teuling, A. J., R. Uijlenhoet, and P. A. Troch (2005), On bimodality in warm season soil moisture observations, *Geophys. Res. Lett.*, *32*, L13402, doi:10.1029/2005GL023223.

1. Introduction

[2] In continental climates, a significant but varying fraction of warm season precipitation can originate from recycled local evaporation [e.g., *Brubaker et al.*, 1993]. It has been argued that anomalous soil moisture conditions, through their effects on evapotranspiration and subsequent precipitation, might sustain themselves causing periods of enhanced floods and droughts. Although rainfall formation is a complex process and the effect of soil moisture is not necessarily positive [e.g., *Giorgi et al.*, 1996; *Ek and Holtslag*, 2004], positive feedbacks have indeed been reported in several Atmospheric General Circulation Model (AGCM) studies [*Shukla and Mintz*, 1982; *Oglesby and Erickson*, 1989; *Beljaars et al.*, 1996; *Bosilovich and Sun*, 1999; *Hong and Kalnay*, 2000; *Koster et al.*, 2004].

[3] Since Illinois (Figure 1) is one of the few continental regions where long-term soil moisture as well as precipitation records are available, many studies focus on this region. Although there is consensus about the significance of the soil moisture-precipitation feedback over the Great Plains region [*Findell and Eltahir*, 2003; *Koster et al.*, 2003, 2004], there is an ongoing debate about whether the feedback controls soil moisture and precipitation dynamics in Illinois [e.g., *Findell and Eltahir*, 1997; *Salvucci et al.*, 2002; *D'Odorico and Porporato*, 2004]. *Findell and Eltahir* [2003] showed that soil moisture can indeed influence the triggering of deep convection in this region, but *Salvucci et al.* [2002] were unable to detect a causal relation between soil moisture and subsequent precipitation in observations.

[4] *Rodríguez-Iturbe et al.* [1991] showed that precipitation recycling over large continental regions can lead to two modes in the steady-state soil moisture probability density function (pdf). Recently, *Kochendorfer and Ramírez* [2005] concluded that this bimodality does not occur when conditions typical to the central United States (including Illinois) are considered. *D'Odorico and Porporato* [2004] showed that bimodality in the steady-state pdf can occur in Illinois when causality between observed soil moisture and subsequent precipitation is assumed. Moreover they showed that a majority of the soil moisture stations in Illinois show bimodality in their warm season (May–September) soil moisture pdf, which they argued was experimental evidence of the existence of such a feedback.

[5] In this paper we investigate the origin of the bimodality in the Illinois soil moisture observations. We argue that the bimodality is not indicative for a strong soil moisture-precipitation feedback, but can be explained by the existence of soil moisture states that show little sensitivity to changes in forcing.

2. Effects of Seasonality

[6] *D'Odorico and Porporato* [2004, p. 8850] used "the fact that in this region the late growing season can be considered to be practically under statistically steady conditions" to support their explanation for the observed bimodality. While the assumption of stationary soil moisture conditions might be valid at a daily timescale, as in the work by *Salvucci* [2001], it loses its validity at longer (sub)seasonal time scales. Figures 2a–2c show the results of an analysis of meteorological data from the Illinois State Water Survey Water and Atmospheric Resources Monitoring Program (WARM, data available from http://www.sws.uiuc.edu/warm), averaged over the different stations. Precipitation (through storm frequency λ and depth α) as well as potential evapotranspiration (E_p) show a clear seasonal trend extending to well beyond the warm season. Moreover, as will be discussed later on, this results in a transition from a net precipitation surplus in May to a deficit in August.

[7] Both leaf area index (ξ) and soil moisture (θ) respond to the seasonality in meteorological forcing. Leaf area index typically shows a strong seasonal cycle on the North American continent [e.g., *van den Hurk et al.*, 2003], with a peak in August. As an illustration, Figure 2c shows measurements made on a grassland site in Kansas in 2001 and 2002 (S. Gower, LAI field measurements for BigFoot MODIS land product validation, KONZ, 2001/2002, available at http://mercury.ornl.gov/ornldaac). Illinois soil moisture data are described by *Hollinger and Isard* [1994], and are available through the Global Soil Moisture Data Bank [*Robock et al.*, 2000]. Figure 2d reveals that the assumption of stationarity is not valid for soil moisture on seasonal timescales: the persistent dry-down during summer covers nearly the full mean soil moisture range [see also *Findell and Eltahir*, 1997, Figure 2].

[8] We study the effect of seasonality on the warm season soil moisture pdf by applying a modified version of the model used by *Laio et al.* [2001] and *D'Odorico and*

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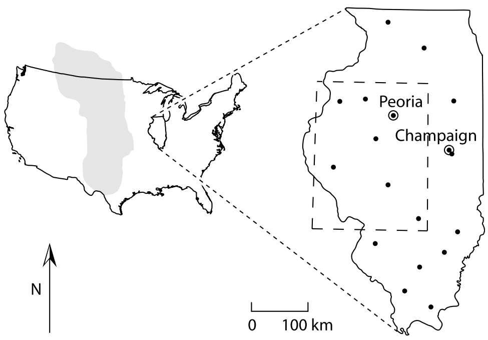

Figure 1. Map of Illinois (USA) showing the location of the relevant WARM sites and the selected ERA40 grid cell. The Great Plains, which have been reported to exhibit the strongest potential for soil moisture-precipitation feedback [*Koster et al.*, 2003, 2004], are shown in grey.

Porporato [2004]. The stochastic model solves the water balance at a daily time scale:

$$\frac{\mathrm{d}\theta}{\mathrm{d}t} = \frac{1}{L}[\varphi(\theta, t) - \chi(\theta)] \tag{1}$$

where θ is the volumetric soil moisture content of the active soil depth L, φ the infiltration, and χ is the loss function. Infiltration equals precipitation P that is not intercepted, or the remaining storage capacity $L(\theta_s - \theta)$, whichever is smaller. θ_s is the porosity of the soil. The size of the interception reservoir is taken proportional to ξ, with a proportionality constant of 0.2 mm per unit of ξ. Daily rainfall occurrence is modeled as a Bernoulli process with occurrence probability $\lambda(t)\mathrm{d}t$ ($\mathrm{d}t$ = 1 d), and depth drawn from an exponential distribution with mean $\alpha(t)$. A key difference with previous work is that we allow for seasonality in λ and α, assuming θ to be driven by meteorological forcing rather than θ to be the main driver of this forcing (as is explicitly assumed by *D'Odorico and Porporato* [2004]). Instantaneous losses $\chi(\theta)$ are modeled as:

$$\chi(\theta) = \begin{cases} E_w \dfrac{\theta - \theta_h}{\theta_w - \theta_h}, & \theta_h < \theta \leq \theta_w \\ E_w + (E_{\max} - E_w)\dfrac{\theta - \theta_w}{\theta_c - \theta_w}, & \theta_w < \theta \leq \theta_c \\ E_{\max} + k_s\left(\dfrac{\theta}{\theta_s}\right)^{2b+3}, & \theta_c < \theta \leq \theta_s. \end{cases} \tag{2}$$

where E_w is the residual evaporation at wilting point θ_w, θ_h the hygroscopic point, θ_c the critical moisture content marking the transition between soil and atmosphere controlled evapotranspiration, $E_{\max}$ the maximum evapotranspiration rate of the vegetation, k_s the saturated hydraulic conductivity, and b is a pore size distribution parameter. We differ from *Laio et al.* [2001] by incorporating the effect of varying ξ and E_p on $E_{\max}$ as [e.g., *Al-Kaisi et al.*, 1989]:

$$E_{\max} = \left(1 - e^{-c\xi}\right)E_p \tag{3}$$

where c is an extinction coefficient for global radiation (0.4 for grass). Other parameters are adopted from *Laio et al.* [2001] and *D'Odorico and Porporato* [2004]: k_s = 20 mm d^{-1}, θ_h = 0.06, θ_w = 0.08, θ_c = 0.24, θ_s = 0.45, b = 5.39, L = 0.5 m, and E_w = 0.2 mm d^{-1}. A comparison of (2) with independent estimates of $\chi(\theta)$ for Illinois confirms the non-linear shape of the loss function (Figure 3a). Field capacity θ_f in Figure 3 is defined as the point where drainage losses are 10% of $E_{\max}$ [*Laio et al.*, 2001].

[9] In order to allow for direct comparison with observations, we evaluate (1)–(3) over the period May–September using the regressions in Figures 2a–2d. As initial condition we assume stationary conditions on 1 May. Since an analytical solution of this problem is not easy to obtain, Figure 3a shows the soil moisture pdf $p(\theta)$ based on Monte Carlo simulations of the model (50,000 seasonal realizations). The pdf shows a distinct bimodality. This bimodality is also apparent, although less pronounced, in the shorter period June–August (not shown). The pdf compares well to observations from Peoria, which have the strongest bimodal tendency according to *D'Odorico and Porporato* [2004]. However the origin of the observed bimodality differs from that provided in previous interpretations.

3. Characterizing Wet and Dry Modes

[10] A visual comparison between Figures 3a and 3b reveals that the bimodal pdf for the period May–September is a near-perfect mixture between the steady-state solutions of $p(\theta)$ for wet (May) and dry (August) regimes. This is confirmed by the observations from Peoria for the individual wet and dry months. These show no bimodality (Figure 3b), but follow the skewness predicted by the

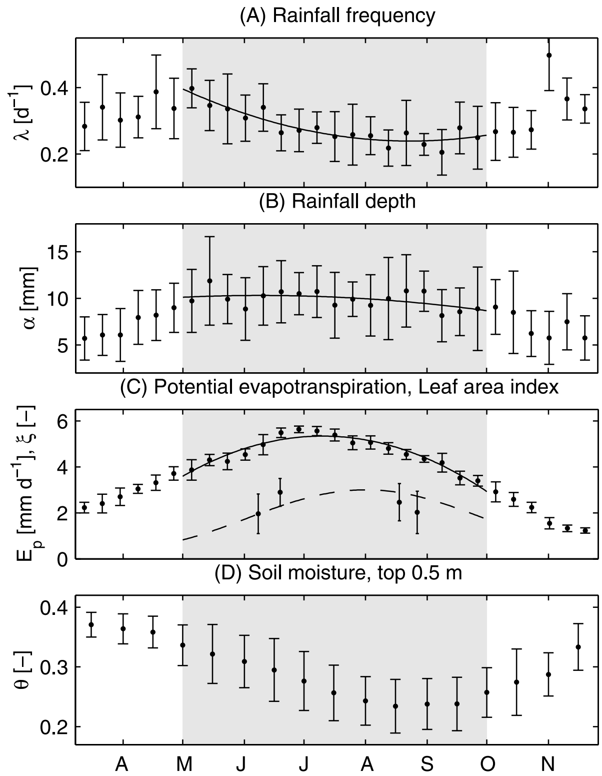

Figure 2. Basic climatology of Illinois. Error bars denote interannual variability (standard deviation) for the years (a–c) 1989–2004 or (d) 1981–2004. Forcing was binned into 17 nine-day periods spanning the period May–September. Solid lines are second-order polynomial regressions for the warm season, and the dashed line is the assumed sine curve for ξ.

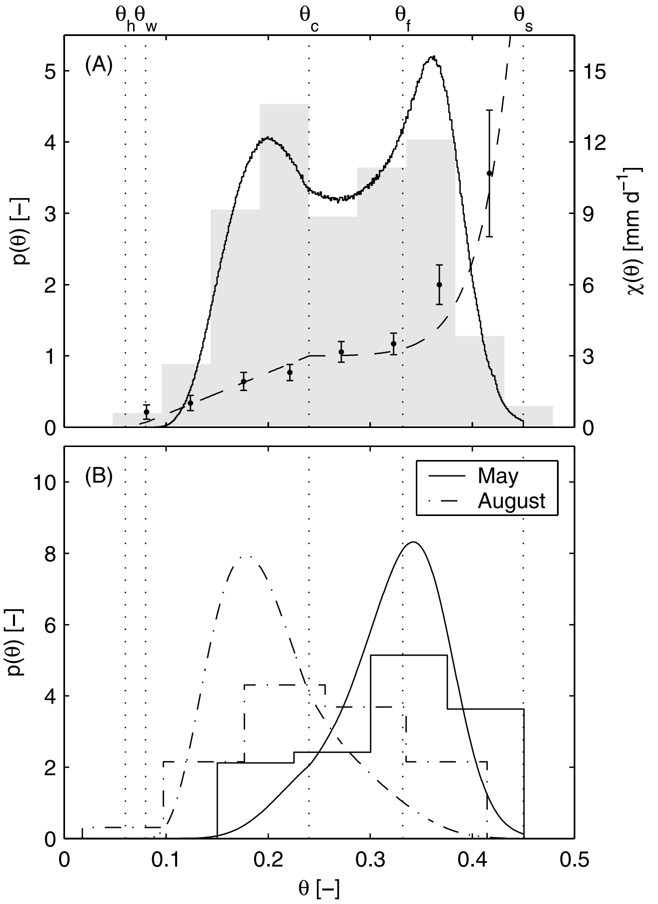

Figure 3. (a) Simulated soil moisture pdf (solid line) and observations (top 0.5 m) from Peoria (1981–2004, grey), both for the period May–September. The dashed line is $\chi(\theta)$, with $E_{\max}$ = 3 mm d^{-1}, with estimates of $\chi(\theta)$ by *Salvucci* [2001] for the period June–August. (b) Steady-state pdf's from *Laio et al.* [2001] with parameters taken from Figure 2 on 31 May ($P > E_{\max}$) and 31 August ($P < E_{\max}$), and observations (top 0.5 m) from Peoria for May and August. Histogram bin widths are adjusted to the number of observations [*Scott*, 1979].

steady-state solutions without feedback. This conflicts with the view of *D'Odorico and Porporato* [2004, p. 8850] that "summer soil moisture dynamics evolve toward either a dry or a wet state in which the system may remain locked for the rest of the warm season". Instead, soil moisture is in a wet state at the beginning of the warm season, and switches to a dry state whenever actual evaporation starts to exceed the precipitation. This switch generally takes place in the period June–July.

[11] This can be explained as follows. Since, in climatic average terms, $P \neq E_{\max}$, no steady-state pdf exists with a mode between θ_c and θ_f (since $d\chi(\theta)/d\theta \approx 0$). In this region, θ is always in transition. Outside this region, the inequality between P and $E_{\max}$ can be balanced by either increased drainage caused by the strong non-linear dependence of the hydraulic conductivity on θ (in case $P > E_{\max}$) or soil moisture limitation on transpiration ($P < E_{\max}$). Hence these regions will act as "attractors" in the soil moisture probability density space. The wet "attractor" is often referred to as field capacity.

[12] The transition between wet and dry modes can occur very rapidly. In the absence of rainfall, the time span of this transition is the time needed for the soil to dry from θ_f to θ_c. With the parameters of Figure 3a, this yields a value of ~20 d. The occurrence of bimodality is controlled by the ratio (Ψ) between the maximum precipitation deficit in the course of the warm season D_P (relative to $E_{\max}$) and the amount of water between θ_f and θ_c that is accessible for transpiration (through L):

$$\Psi = \frac{D_P}{L(\theta_f - \theta_c)} \tag{4}$$

[13] In climates with precipitation surplus in winter, no bimodality occurs if $\Psi < 1$. Figure 4 shows the effect of varying Ψ on $p(\theta)$, obtained by increasing L. The resulting pdf's have a shape similar to observations from Champaign, one of the stations that does not show bimodal behavior. This confirms that local variations in Ψ through either climate, soil, or vegetation properties control the variations in soil moisture dynamics in Illinois. Since the shape of $\chi(\theta)$ in Figure 3a is not unique for Illinois (nor is the climate), similar dynamics are likely to occur in other mid-latitude regions. A rapid transition between wet and dry states has also been observed in studies of spatial soil moisture patterns in a small Australian catchment [*Grayson et al.*, 1997].

[14] To investigate if coupled land-atmosphere models are capable of reproducing the observed soil moisture dynamics, we analyze the ERA40 reanalysis soil moisture data (obtained from the ECMWF data server) for the warm seasons of the years 1981–2002 (Figure 1). Whereas observations tend to be bimodally distributed, the ERA40 data show one distinct peak and less dynamical range (Figure 5). Since we concluded that the bimodality is related to the shape of $\chi(\theta)$, we investigate to what extent this shape controls the soil moisture pdf by repeating our model runs with soil parameters and a simplified loss function of the ERA40 land surface scheme (Figure 5). This (daily average) function was derived from the model equations [see *van den Hurk et al.*,

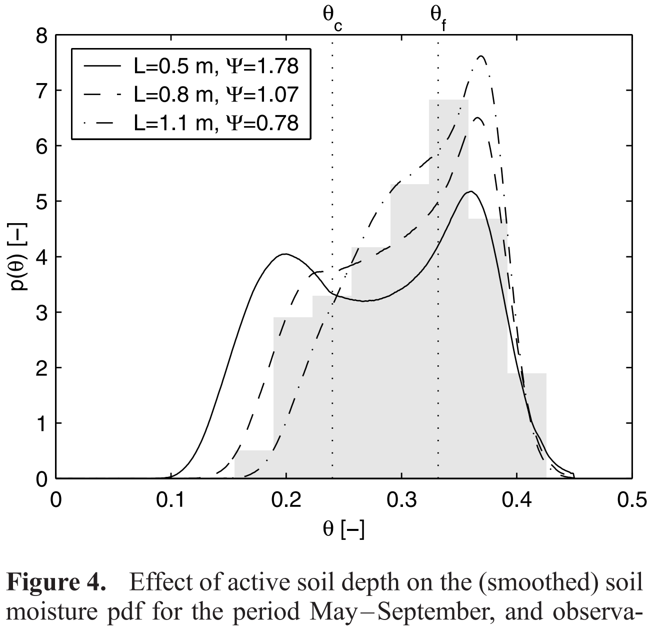

Figure 4. Effect of active soil depth on the (smoothed) soil moisture pdf for the period May–September, and observations (top 0.5 m) from Champaign for the same period in the years 1984–2004 (grey).

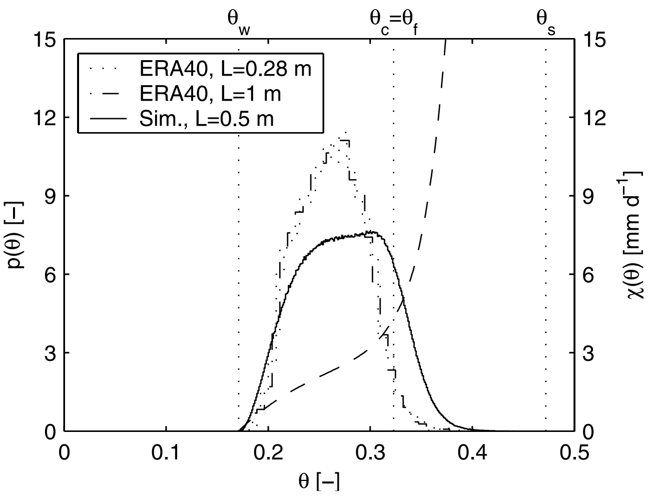

Figure 5. ERA40 and simulated soil moisture pdf for the period May–September, and simplified $\chi(\theta)$ (dashed line) with E_{max} = 3 mm d^{-1}.

2000] under assumptions of instantaneous vertical soil moisture redistribution, full vegetation cover, stomatal conductance only reduced by soil moisture, and equal aerodynamic and minimum surface resistance. The results, although showing a slightly higher variability, show similar dynamics as the ERA40 data and no bimodality. Although other factors contribute to the damped soil moisture dynamics in the ERA40 [e.g., *Seneviratne et al.*, 2004], this shows that the simulated soil moisture dynamics are strongly controlled by the parameterization of the soil moisture losses.

4. Conclusion and Discussion

[15] In this paper we show that soil moisture bimodality cannot be considered as conclusive evidence for the existence of a soil moisture-precipitation feedback, since no feedback is necessary to explain the existence of the wet and dry modes. However, our results do not exclude the possibility that such a feedback exists, and some feedback effects might exist in the data that was used. Although many processes (including soil moisture-precipitation feedback) can lead to bimodal soil moisture distributions, we think the process described in this paper dominates possible other processes in their effect on soil moisture dynamics in Illinois. Since the conditions that generate the bimodality are not typical to Illinois nor to regions were a land-atmosphere feedback may exist, similar soil moisture dynamics are likely to exist in many mid-latitude regions. Understanding these dynamics is crucial for the development, parameterization, and validation of land surface models.

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