The Response of Soil Moisture to Long-Term Variability of Precipitation

WANRU WU
School of Earth and Atmospheric Sciences, Georgia Institute of Technology, Atlanta, Georgia

MARVIN A. GELLER
Marine Sciences Research Center, State University of New York at Stony Brook, Stony Brook, New York

ROBERT E. DICKINSON
School of Earth and Atmospheric Sciences, Georgia Institute of Technology, Atlanta, Georgia

Abstract

Soil hydrology is a widely recognized low-pass filter for the interaction between land and atmosphere. However, the lack of adequate long-term measured data on soil moisture profiles has precluded examination of how soil wetness responds to long-term precipitation variations. Such a response can be characterized by its amplitude damping, phase shifting, and increasing persistence with soil depth. These should be correlated with the climate spectra through the interactions between the land and the atmosphere. The major objective of this study is to investigate how precipitation signals are manifested in vertical soil moisture profiles in the context of timescales. Thus, the natural variability of soil moisture profiles is documented using 16 yr of field observational data of soil moisture measured at 11 levels of various depths down to 2.0 m at 17 locations over Illinois. Detailed statistic analyses are made of the temporal variations of soil moisture profiles and concurrently measured precipitation over the 16-yr period of 1981–96. Cross-spectral analysis is performed to obtain the coherency pattern and phase correlation of surface and profile soil moisture time series to determine phase shift and amplitude damping. A composite of the drought events during this time period is analyzed and compared with the 16-yr climatology. The major findings are that 1) the amplitude decreases with soil depth, with the dryness signal penetrating more deeply than the wetness signal; 2) the phase shift with soil depth is correlated with the timescales of the variation, such that it is deeper with longer timescales; and 3) the seasonal variation of soil moisture is amplified in the drought-year composite, with an increased phase shift from soil surface to bottom. Hence, the observations provide a description of the soil moisture profile variability as a function of soil depth. Whether or not climate models can reproduce this variability should be a good test of their land process representations in the treatment of soil hydrology.

1. Introduction

Soil moisture has a memory considerably longer than that of most atmospheric processes, and, hence, climatic anomalies may persist through processes dependent on it. The temporal structure and vertical profile of soil moisture depends on the temporal variability of precipitation and evaporation. A general motivation for this study is the hypothesis that interaction between soil moisture profiles and the atmosphere makes a substantial contribution to climate variability. Numerous studies, both observational and computational, have investigated how soil moisture interacts with the atmosphere. This paper examines how precipitation signals are manifested in vertical soil moisture profiles in the context of timescales.

Corresponding author address: Dr. Wanru Wu, School of Earth and Atmospheric Sciences, Georgia Institute of Technology, 221 Bobby Dodd Way, Atlanta, GA 30332-0340.
E-mail: wwu@eas.gatech.edu

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Delworth and Manabe (1988, 1989) analyzed the temporal variability of soil moisture in a multiyear integration of a general circulation model of the atmosphere. They showed that, within a GCM, the time series of soil moisture contains variance on seasonal to interannual timescales, and that the soil layer acts as an integrator of short-timescale precipitation anomalies, transforming the almost white noise time series of monthly mean precipitation into the red noise time series of soil moisture. Vinnikov et al. (1996) and Entin et al. (2000) considered the variance of the soil moisture observations separating the time series into red and white noise components. They related the red noise component of variability to atmospheric forcing, estimated it has a timescale of roughly 2–3 months, and showed that the upper 1.0 m was in good agreement with the theoretical estimates of Delworth and Manabe (1988). Huszar et al. (1999) further indicated that the soil moisture responds relatively quickly to climate variations and that changes in its availability are an important aspect of

Soil moisture acts as a feedback on climate in various ways. It acts to delay and prolong the effects of meteorological drought (Nicholson 2000) and to enhance the severity and persistence of floods and droughts (Bonan and Stillwell-Soller 1998; Hong and Kalnay 2000; Pal and Eltahir 2001). The climate of summertime precipitation heavily depends upon the soil moisture availability (Schar et al. 1999; Findell and Eltahir 1999).

Studies investigating the influences of soil moisture anomalies on monsoon systems (Matsuyama and Masuda 1998; Douville et al. 2001; Small 2001) have drawn similar conclusions. The seasonal predictability of atmospheric surface climate anomalies depends on the interannual variability of the soil moisture (Wang and Kumar 1998; Douville and Chauvin 2000), such that a better initialization of soil moisture would improve seasonal climate simulations (Dirmeyer 2000).

Many investigations of the two-way land–atmosphere interaction have used numerical models. These may be important for helping to isolate cause and effect, and for performing partial analysis of the role of soil moisture, but their results depend on their parameterizations. Thus, evidence from observations can identify relevant results from numerical modeling and indicate how land surface processes in climate models should be represented to obtain a realistic simulation of atmospheric variability. Especially important to test with observations is the variability of soil moisture with depth as an indication of coupling to the atmosphere on various timescales.

McNab and Karl (1989) schematically indicated that fluctuations in precipitation propagate down through the land branch of hydrological cycle; that is, they are seen in the surface runoff, the soil moisture, the streamflow, and the groundwater. The effects in the soil are characterized in terms of amplitude damping and phase shifting with soil depth. The lack of adequate long-term data on soil moisture profiles has precluded adequate study of the temporal variability of such characteristics of the soil moisture profile. However, this knowledge is crucial for understanding the soil contributions to the hydrological cycle and improving parameterizations of land surface processes. The objectives of this study are to identify and quantify with observations the variability of soil wetness at low frequencies and to show how this varies in phase and amplitude with soil depth in response to precipitation variations, and hence provide a basis for further investigating the impacts of the soil moisture profile on atmospheric variability.

This study uses 16 yr of field observations from 1981 to 1996, with 11 layers of soil moisture at 17 grass-covered sites across Illinois, one of the best long-term datasets of soil moisture profile measurements, and concurrent precipitation data. Detailed statistic analysis is made of the variations of soil moisture profile and precipitation anomaly time series. A composite of the drought events that occurred during this time period is analyzed and compared with the 16-yr climatology.

### 2. Data used and analysis method

The data used in this study (Robock et al. 2000) consist of the observed precipitation and the soil water content from the surface to 2-m depth in each of 11 soil layers (0–0.1, 0.1–0.3, 0.3–0.5, 0.5–0.7, 0.7–0.9, 0.9–1.1, 1.1–1.3, 1.3–1.5, 1.5–1.7, 1.7–1.9, and 1.9–2.0 m) from February 1981 to August 1996. The original soil moisture data were measured at irregular intervals of approximately three times a month at 17 stations over grass throughout the state of Illinois (an important agricultural region extending over approximately 38°–42° latitude and 88°–90° longitude). The dominant soil texture is silt loam in the top 1 m, and the average total porosity over the six layers of the top 1 m of the soil is 0.488 (therefore, a nominal saturated soil moisture value for the 2-m soil column is 976 mm). The methods of data measurement and their uncertainties are detailed by Hollinger and Isard (1994).

In our study here, we take the original time series of the soil moisture profiles and precipitation at the 17 sites from February 1981 to August 1996 and we average them 1) temporally in time to monthly means and 2) spatially across all 17 sites to obtain a statewide mean.

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**Table 1. The statistical methods used in this study and their purposes.**

<table>
<thead>
<tr>
<th>No.</th>
<th>Method</th>
<th>Reference</th>
<th>Purpose</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>High-pass filter*</td>
<td>Hamming (1989, chapter 12)</td>
<td>To remove long-term trend</td>
</tr>
<tr>
<td>2</td>
<td>Bandpass filter**</td>
<td>Murakami (1979)</td>
<td>To isolate frequency band</td>
</tr>
<tr>
<td>3</td>
<td>Periodogram analysis</td>
<td>Jenkins and Watts (1968, chapter 6)</td>
<td>To obtain power spectrum and identify dominant mode</td>
</tr>
<tr>
<td>4</td>
<td>Cross spectrum</td>
<td>Jenkins and Watts (1968, chapter 8)</td>
<td>To calculate coherency and phase between two time series</td>
</tr>
<tr>
<td>5</td>
<td>Wavelet analysis</td>
<td>Torrence and Compo (1998)</td>
<td>To identify dominant mode varying in time.</td>
</tr>
</tbody>
</table>

* Uses the second-order “Butterworth” transfer function (Hamming 1989).
** Uses the first-order “Butterworth” transfer function (Guillemin 1957).
The time and space averaging will smooth the data in time, that is, filter out high frequencies. Can this filtering affect the results of this study? Since the soil moisture profiles were measured only at irregular intervals of roughly three times a month, we decided that the sub-monthly variations in our time series would be inadequately represented, distorted, and untrustworthy, and therefore we averaged the data to monthly values. Further testing by comparing the spectral analysis results before and after this averaging process showed that the derived data generated virtually the same spectra as the original data, and using the area-averaged data instead of site data significantly reduced the measurement uncertainties and made the results more relevant for comparison with climate models.

The five statistical methods we used and our objectives for using them are summarized in Table 1. The trend and longer timescale oscillations of the series analyzed here are removed by the high-pass filter prior to any further operations on the data (method 1). The recursive Butterworth method (applied in both methods 1 and 2) is used to accurately locate the spectral peaks and filter the frequency bands, and to avoid the cutoff problem. Power spectral analysis (method 3) is performed on the soil moisture profile and precipitation time series to identify dominant frequencies. Moreover, we perform a cross-spectral analysis (method 4) to obtain the coherency patterns and phase correlations of surface and profile soil moisture time series. In addition, we use a wavelet analysis (method 5) to see the frequency-band distributions in the period-time domain. The decomposition of a time series into the period-time domain determines both the dominant modes of variability and how those modes vary in time.

Figure 1 shows the monthly anomaly time series of (a) soil water storage and (b) precipitation, after removing (via method 1 in Table 1) any long-term (low frequency) trends or variations whose periods exceeded the record length (i.e., 187 months). The standard deviation obtained over the 16-yr time series was 52.0 mm for soil moisture storage (2.0-m soil column) and 44.1 mm month$^{-1}$ for precipitation, whereas, for comparison, the 16-yr mean values of soil moisture and precipitation were 681.3 mm and 81.7 mm month$^{-1}$, respectively. The soil wetness is near maximum during the spring months and at this time displays little interannual variation, thus indicating that winter and early spring precipitation (while evaporation is low) is generally sufficient to fully recharge the soil moisture over Illinois. Inspection of Fig. 1a also reveals that summer droughts occurred in 1983, 1988, 1991, 1994, and 1995, with the strongest in 1983, 1988, and 1991. The summer soil moisture deficiencies for these five drought years exceed or approach twice the standard deviation. The area-average monthly soil water storage across Illinois during this analysis period was lowest in August 1983. It is noteworthy in Fig. 1a that for this 16-yr period, the summer soil moisture anomaly remained positive only during one summer, namely the summer of 1993, which is well known for the widespread flood event in the upper Mississippi region of Illinois and Missouri. Additionally, it is evident in Fig. 1 that, while the standard deviation of soil water storage and precipitation are of the same order, the temporal variability of the precipitation is substantially higher than that of the soil water storage. Moreover, while the largest anomalies of precipitation in Fig. 1b are positive anomalies (large precipitation), the largest anomalies in soil water storage are negative anomalies (drought). This dichotomy reflects the fact that, for large positive precipitation anomalies, much of the precipitation excess is often lost to either runoff and streamflow (in summer) or retained in the snowpack (in winter) and melted or sublimated later, rather than being retained in the soil. In contrast, for large negative soil moisture anomalies (droughts), the soil moisture deficit from relatively modest negative precipitation anomalies can be substantially amplified by high air temperatures and low humidity, and thus high evaporation.
3. Results and discussion

a. Seasonal amplitude damping

In order to initially illustrate the characteristics of soil moisture variability with soil depth, the soil moisture profile time series are temporally decomposed into three frequency bands: seasonal (period = 1 yr), intraseasonal (period < 1 yr), and interannual (period > 1 yr). For this purpose, the average of values in each series on the same calendar date (mean annual cycle) is the seasonal component, the average over an individual year gives the interannual component, and the residual after removing seasonal and interannual components is the intraseasonal component.

Figure 2 shows the seasonal component of soil moisture at the 11 levels analyzed down to 2 m (each layer is 0.2 m thick except for the top and bottom, which are 0.1 m). It illustrates the amplitude decay with depth ("amplitude damping") and phase shifts to latter times progressing downward ("phase shifting"), well-known characteristics of diffusionlike propagation of a periodic disturbance from a boundary. In the 11 plots, the same coordinate scales are used in order to compare them with each other. From the soil surface (0–0.1 m) to the bottom (1.9–2.0 m), the maximum value of soil water content (in "volumetric" units, i.e., percent of unit soil volume occupied by water) decreases from 37.89% to 33.73%, whereas the minimum increases from 20.61% to 33.04%, and the date of the minimum moves from August to October.
The seasonal cycle of soil moisture in Fig. 2 is substantially more pronounced than that for precipitation (not shown) because of the impact of evapotranspiration, which is highest in the summer months because of the warmest temperatures, highest solar radiation, and peak greenness (discussed further in section 3d). The amplitude of the annual cycle is quite small below 1 m because the root density of the overlying grassland becomes vanishingly small below the 1.0-m soil depth (Jackson et al. 1996). Figure 2 shows that the month of minimum soil moisture in the top four soil layers, which hold most of the roots, is August. Maximum evapotranspiration is also in midsummer. Hence, the amplitude damping and phase-shift effect shown here could be substantially dependent on vegetation and root density profile and may be different at sites with different vegetation types, vegetation densities, or annual cycles of vegetation greenness (e.g., a winter wheat area over Oklahoma, where greenness peaks in April and harvest occurs in early June).

Figure 3 provides an alternative depiction of the amplitude damping and phase shifting of the seasonal cycle of soil moisture with increasing depth. Figure 3a shows, layer by layer, the monthly departure of the soil moisture from its layer annual mean. Figure 3b shows in blue (red) for each soil layer, the ratio (denoted by vertical line segment) of the annual maximum of the positive (negative) anomaly of soil moisture for that layer to that of the top soil layer. We refer to the blue (red) plot in Fig. 3b as the wetness (dryness) signal. Thus, Fig. 3b quantifies the damping of positive and negative soil moisture anomalies with depth. This latter ratio, the “damping ratio,” decreases with soil depth, with about 80% of the amplitude damping for both the wetness and dryness signals occurring within the top 1 m of the soil column. The dryness signal extends deeper into the soil than does the wetness signal. Nearly half of the damping for the wetness signal (to damping ratio of about 0.6) occurs in the first layer below the top layer (widths in Fig. 3b), whereas the dryness signal takes two layers below the top layer to reach a similar damping ratio. Evidently the roots, which are concentrated in the top 0.5 m, extend the depth of the release of subsurface soil water into the atmosphere, especially in the season of vegetation peak greenness.

b. Soil moisture phase shift associated with timescale of variation

To quantify the phase shift, a cross-spectral analysis (method 4 in Table 1) was applied. This method provides two different measures of the dependence between two given time series. The first is the coherency spectrum, which effectively measures the correlation between the two times series at each frequency, and the second is the phase spectrum, which measures the phase difference between the two time series at each frequency.

The cross spectra between the surface soil moisture (depth of 0–0.1 m) and each layer in the soil moisture profile (11 layers, down to 2.0 m) were calculated, and the results are shown in Fig. 4a. The shaded area denotes where the coherency coefficient (this parameter, ranging from 0.45–1.0, measures the degree of coherency between two time series) is above the 99% confidence level, and arrows represent phase lag (clockwise direction) or phase advance (anticlockwise direction), with the upward vertical direction denoting zero lag. The vertical color bars in Fig. 4a denote the periods of 3.5 (red), 12 (green), and 42 (blue) months, corresponding to the intraseasonal, seasonal, and interannual variations, respectively, that are used for calculating the corresponding phase shifts with soil depth in Fig. 4b. At these three periods (as well as at the fourth period of about 5 months), the soil moisture variations are well correlated throughout the soil profile. The phase shift results are displayed in Fig. 4b. From the land surface (0–0.1 m) to the bottom (1.9–2.0 m), the phase shifts (lags) are about 1, 3.6, and 7.6 months for the intraseasonal, seasonal, and interannual variations, respec-
FIG. 4. (a) Coherency (contour interval of 0.1, decreasing downward from surface contour value of 1.0) and phase correlation (arrows) between the surface soil moisture and each layer in the soil moisture profile, and (b) phase shift of soil moisture temporal variation with soil depth. Arrows represent phase lag (clockwise direction) or phase advance (anticlockwise direction). The shaded area is above the 99% confidence level for the coherency. The three frequency values of intraseasonal (red), seasonal (green), and interannual (blue) are highlighted.

c. Characteristics of drought

The drought years are defined from Fig. 1 when the deficiencies of soil water storage exceed or approach two standard deviations. They occur in 1983, 1988, 1991, 1994, and 1995. The seasonal variation of the soil moisture during drought is compared with that for the 16-yr
climatology in Fig. 5. With drought, the anomaly amplitudes at the soil surface increase from 10.71 to 12.70 for the wetness signal and from 6.68 to 12.90 for the dryness signal [all units are volumetric content (v/v)%], while the annual mean volumetric soil water contents at the surface for the 16-yr climatology and the 5-yr drought composite are 29.44% and 28.58%, respectively. In Fig. 5b, the phase of the soil moisture minimum shifts from July at the soil surface to October at the bottom, rather than the August to October shift of the 16-yr climatology in Fig. 5a. Figure 6 compares the damping ratio for the dryness and wetness signals. They differ more in drought years (triangles in Fig. 6) as a result of the dryness signal damping more slowly and the wetness signal damping more quickly than for the 16-yr climatology.

Comparison of Figs. 5 and 6 shows that in seasonally dry environments, annual cycle perturbations of soil moisture profiles are amplified for both positive and negative anomalies rather than for the negatives only, and this amplification extends to the bottom of the soil column. The phase shifts also increase during drought.

d. “White” atmospheric forcing and “red” land surface response

The concepts of white noise and red noise have been commonly used in meteorology for decades (e.g., Gilman et al. 1963; Jones 1975; Delworth and Manabe 1988). Simply speaking, white noise is a random process, with a constant spectral density for all frequencies, while red noise has a spectral density that is relatively large at low frequencies, which then decays quadratically with increasing frequency.

Wavelet analysis has become a common tool for analyzing the localized variations of power within a time series to see the frequency-band distributions in the period-time domain. The decomposition of a time series into time–frequency space determines both the dominant modes of variability and how those modes vary in time. Methods 5 and 3 in Table 1 were used to obtain the wavelet power spectra of Fig. 7 and the power spectra of Fig. 8, respectively.

Figure 7 shows the wavelet power spectrum of the precipitation, and the soil water storage over the 2-m soil column for the years 1981–96. The vertical axes are the Fourier period (in months), and the horizontal axes are time (in calendar years). The shaded areas denote the regions of 95% confidence level or above for a red-noise process. Anything below the line of demarcation (“ship’s hull”) is of dubious reliability due to edge effects and insufficient record length for longer periods (details in Torrence and Compo 1998). The most basic features of the vertically averaged soil moisture spectra are their resemblance to red noise (as stated in Delworth and Manabe 1988) and an annual variation (12-month period) that is dominant and consistent throughout the entire time domain (Fig. 7b). The spectra of precipitation are “white” on timescales of a month or longer; its annual variation (12-month period) is usually strong but has differences between years and thus is not always dominant (Fig. 7a). Since we have shown
earlier that the bottom 1-m soil moisture has little seasonal variability, the vertical averaging over the entire 2 m for Fig. 7b probably suppressed some meaningful frequencies that characterize the top 1 m (such as 3.5 or 5 months). In order to clarify this concern, we did the same calculation for just the top 1-m soil column. The result showed that its “redness” patterns were similar to those in Fig. 7b, and no significant high frequencies (period < 1 yr) occurred.

According to the soil water balance equation

$$\frac{\partial w}{\partial t} = P - E - R,$$

the changes in time of total soil water $w$ must balance precipitation $P$ minus the sum of evapotranspiration $E$ and runoff $R$. If $E$, with an evaporation scale $E_0$, can be assumed proportional to $w$ with a soil water scale $w_0$, then the timescale at which redness occurs is proportional to $w_0/E_0$. Since $E_0$ has a pronounced maximum in summer and a pronounced minimum in winter, soil moisture will be more persistent in winter than in summer (Delworth and Manabe 1988). Runoff $R$, if also proportional to soil water, will also increase the timescale of soil water variability.

Figures 8 displays the power spectrum of soil moisture compositied into four typical soil layers: 0–0.3, 0.3–0.9, 0.9–1.5, and 1.5–2.0 m. The spectra are normalized by their mean values over frequencies at each layer. The dominant annual cycle mode has been removed (via method 2 in Table 1) in order to show the spectral characteristics of the other timescales. The redness of the soil moisture spectra increases with increasing soil depth, and spectral regions of large variance shift to lower frequencies; that is, the soil hydrology is acting essentially as a low-pass filter (Entekhabi et al. 1992) throughout the soil moisture profile. The significant spectral peaks are dominated by low frequencies (periods of 1 yr or more) at deep soil depths (below 1.0 m). These soil moisture profile variations may in turn influence climate variations on these timescales.

This paper has emphasized precipitation as the random forcing for soil moisture. We have largely neglected the additional contribution from evapotranspiration, ET. Although the ET contribution most likely dominates that of precipitation for the seasonal cycle over Illinois, we believe ET’s random contribution, for example, from random variation in net surface radiation or temperature because of clouds or frontal passages, is quite a bit smaller than that of precipitation. If so, the nonseasonal cycle random variations of ET are primarily a response to changes of soil moisture, or snow cover, in turn responding to precipitation.

4. Concluding remarks

This study has quantified with observations, for a 16-yr period over Illinois, the low-pass filter climatology of soil hydrology in terms of amplitude damping, phase shifting, and increasing persistence with soil depth, and it has differentiated the variability of soil moisture during drought years. The leading mode of variation for the soil moisture profile is the seasonal cycle. The most basic feature of the soil moisture profile spectra is its “redness,” or low-frequency nature (while the precipitation spectrum is “white”). The redness of the soil moisture spectra increases with increasing soil depth, that is to say at deeper depths the largest variances are shifted to lower frequencies. The amount of phase shifting depends on the timescales of the variation and depth within the soil. Almost no significant phase shift is observed for the intraseasonal component below approximately 1.1-m soil depth. For the seasonal and interannual components, the rate of phase shift increase with depth is more substantial and nearly constant throughout
the entire 2-m soil profile. The amplitude of the soil moisture seasonal cycle decreases with increasing soil depth, with around 80% of its amplitude being reduced within the top 1.0-m soil layer for both the wetness and dryness signals. The amplitude damping and phase-shift effect derived here could be significantly dependent on the vegetation and root density profile and may be different at sites with different vegetation properties. The seasonal variation in soil moisture is amplified in droughts, with an increased phase shift from soil surface to bottom, while the dryness signal damps more slowly and the wetness signal more quickly compared to the 16-yr climatology.

This study has used data from Illinois, a midlatitude and grass-covered region. It remains to be addressed how those characteristics may differ for different land cover and climate regions for which long-term global soil moisture profile measurements are not available. The observations we used here have provided a description of phase shifting, anomaly amplitude damping, and redness increasing with soil depth, and allow us to assess their potential contribution to the climate spectra. Hence, whether or not climate models can reproduce these characteristics should be a good test of the adequacy of their land process representations in the treatments of soil hydrology.

Acknowledgments. We appreciate the valuable remarks by anonymous reviewers, and the Soil Moisture Data Bank data source established by Dr. Alan Robock et al. (2000). This research was supported by a NASA grant through the EOS IDS program and an NSF grant.
on Land–Ocean–Atmosphere Interaction (ATM-0096099).

REFERENCES


