Impact of Initial Soil Moisture Anomalies on Subsequent Precipitation over North America in the Coupled Land–Atmosphere Model CAM3–CLM3

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ABSTRACT

To investigate the impact of anomalous soil moisture conditions on subsequent precipitation over North America, a series of numerical experiments is performed using a modified version of the Community Atmosphere Model version 3 and the Community Land Model version 3 (CAM3–CLM3). First, the mechanisms underlying the impact of spring and summer soil moisture on subsequent precipitation are examined based on simulations starting on 1 April and 1 June, respectively. How the response of precipitation to initial soil moisture anomalies depends on the characteristics of such anomalies, including the timing, magnitude, spatial coverage, and vertical depth, is then investigated. There are five main findings. First, the impact of spring soil moisture anomalies is not evident until early summer although their impact on the large-scale circulation results in slight changes in precipitation during spring. Second, precipitation increases with initial soil moisture almost linearly within a certain range of soil moisture. Beyond this range, precipitation is less responsive. Third, during the first month following the onset of summer soil moisture anomalies, the precipitation response to wet anomalies is larger in magnitude than that to dry anomalies. However, the resulting wet anomalies in precipitation quickly dissipate within a month or so, while the resulting dry anomalies in precipitation remain at a considerable magnitude for a longer period. Consistently, wet spring anomalies are likely to be ameliorated before summer, and thus have a smaller impact (in magnitude) on summer precipitation than dry spring anomalies. Fourth, soil moisture anomalies of smaller spatial coverage lead to precipitation anomalies that are smaller and less persistent, compared to anomalies at the continental scale. Finally, anomalies in shallow soil can persist long enough to influence the subsequent precipitation at the seasonal time scale. Dry anomalies in deep soils last much longer than those in shallow soils.

1. Introduction

Many climatologists have speculated on the role of soil moisture in the midlatitude climate. Namias (1952) hypothesized that soil moisture could support month-to-month persistence in climatic anomalies over the United States. This hypothesis has been followed by numerous studies on land–atmosphere interactions, focusing on soil moisture–precipitation feedback. Precipitation influences soil moisture; the resulting anomalies in soil moisture feed back to impact precipitation, which further influences soil moisture, through the water, energy, and momentum exchanges between land and atmosphere. Anomalies of soil moisture can last for weeks or even months, while anomalies in the atmosphere are much shorter lived, often no longer than a few days. The atmospheric response to precipitation-induced soil moisture anomalies can prolong the precipitation anomalies. Such memory of the land–atmosphere system and the resulting climate persistence is important for the prediction of seasonal climate, especially for the prediction of extreme climate events like droughts and floods. Specifically, soil moisture can serve as a potential predictor for precipitation over regions where the land memory is long and the soil moisture–precipitation coupling is strong.

Land memory and the strength of soil moisture–atmosphere feedback vary from place to place. Koster and Suarez (2001) analyzed the lagged autocorrelation of soil moisture over the globe based on results from the National Aeronautics and Space Administration (NASA) Seasonal-to-Interannual Precipitation Project (NSIPP) atmospheric general circulation
model (AGCM), and documented significant soil moisture persistence over regions including the central United States. Entin et al. (2000) analyzed the soil moisture observations across the Northern Hemisphere midlatitudes, including Illinois and Iowa in the United States. Their results showed the persistence time of soil moisture is approximately 2 months, consistent with an earlier study based on model data (Huang et al. 1996). On the other hand, the U.S. region has been identified as a region of strong coupling between soil moisture and precipitation (Koster et al. 2004). Different studies have suggested to various extents a role of soil moisture in the 1988 summer drought and 1993 summer flood over the United States. For example, Namias (1991) indicated that the deficient precipitation in antecedent seasons, associated with extratropical sea surface temperature (SST) pattern, played a key role in generating the 1998 drought. Trenberth and Guillemot (1996) suggested that the 1988 drought and 1993 flood were mainly initiated by the anomalous tropical SST, but were persisted and/or enhanced by the resulting soil moisture anomalies. Nonetheless, many studies (e.g., Atlas et al. 1993; Sud et al. 2003) agreed that climate models perform better in simulating the 1988 drought and 1993 flood when more realistic soil moisture values are provided.

Many studies exploring the link between soil moisture and precipitation have been performed using GCMs (e.g., Shukla and Minz 1982; Rind 1982; Oglesby and Erickson 1989; Oglesby 1991; Oglesby et al. 2002; Sud et al. 2003) or regional climate models (e.g., Paegle et al. 1996; Seth and Giorgi 1998; Bosilovich and Sun 1999; Lu et al. 2001; Pal and Eltahir 2001; Georgescu et al. 2003). The topic of common interest is how the initial soil moisture anomalies impact the climate conditions throughout the model integration. Most of these studies support a positive feedback between soil moisture and precipitation over the United States; that is, positive (negative) anomalies in soil moisture leads to positive (negative) anomalies in precipitation, which can then enhance the initial soil moisture anomalies. Using the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM1), Oglesby and Erickson (1989) imposed “desert-like” initial soil moisture over an extensive area of North America. They found that in the summertime, specifically in July, soil moisture reduction prolonged and/or amplified drought over North America, and suggested that moisture advection from the Gulf of Mexico played an important role in determining the areas where the reduced soil moisture persisted. Using the Purdue Regional Model (PRM), Bosilovich and Sun (1999) found that the simulated 1993 flood became much less severe when initial soil moisture was reduced in two 1-month integrations starting from 1 June and 1 July. In addition, back trajectory analysis of the National Centers for Environmental Prediction (NCEP) reanalysis data (Dirmeyer and Brubaker 1999; Brubaker et al. 2001) and the water vapor tracer analysis in GCM simulations (Bosilovich and Schubert 2002; Bosilovich and Chen 2006) suggest strong precipitation recycling (the direct pathway through which soil moisture influences precipitation) over the Mississippi River basin during summer.

The significance of the feedback between soil moisture and precipitation varies with several aspects, including timing, magnitude/direction, and soil depth of soil moisture anomalies. Oglesby (1991) documented that the timing of soil moisture anomalies is critical over North America. He found that the initial soil moisture reduction in early spring (1 March) did not persist through summer, while the reduction in late spring (1 May) did. Using a modified version of the NCAR Regional Climate Model (RegCM), Pal and Eltahir (2001) carried out 1-month simulations over the U.S. Midwest during May, June, July, August, and September for 1988 and 1993. They found that the relative sensitivity (i.e., the percentage change of precipitation in response to percentage change of soil saturation) does not depend on the timing of soil moisture anomalies. The absolute sensitivity (i.e., the precipitation change in millimeters per day in response to percentage change of soil saturation) is highest during June and July in the flood year 1993, but is nonexistent in the drought year 1988. They also documented an asymmetric response that the soil moisture–precipitation feedback is stronger during drought than during flood over the Midwest. Recently, Oglesby et al. (2002), using NCAR CCM3, documented that dry anomalies had larger impacts than wet anomalies, and anomalies in March had a larger impact than those in June. Similarly, based on the NASA Goddard Earth Observing System (GEOS-1) Data Assimilation System, Bosilovich and Schubert (2001) found that the recycling ratio is above average in the dry summer of 1988 and below average in the wet summer of 1993, suggesting a stronger influence of land conditions in droughts than in floods. Oglesby et al. (2002) also found that soil moisture in deep soil layers, where the water-holding capacity is large, powerfully influences upper soil layers and plays a more important role at interannual to decadal time scales than at monthly and seasonal time scales.

While previous studies collectively show that the impact of soil moisture initialization depends on several aspects, this dependency has not been systematically explored. In this study using a numerical modeling approach, we will systematically investigate how the im-
impact of soil moisture anomalies on subsequent precipitation depends on the timing, magnitude/direction, spatial coverage, and soil depth of the initial soil moisture anomalies. We will focus on not only the magnitude but also the persistence of the resulting climate anomalies. The model and methodology used are described in section 2. Section 3 explains the mechanisms through which soil moisture influences subsequent precipitation and focuses on the difference between spring and summer. Section 4 documents how the precipitation response varies with the timing, magnitude, spatial coverage, and depth of soil moisture anomalies. Conclusions and discussion are presented in section 5.

2. Model and methodology

a. Model description

This study is carried out using the coupled CAM3–CLM3 model, which consists of the Community Atmosphere Model version 3 (CAM3) (Collins et al. 2004) and the Community Land Model version 3 (CLM3) (Dai et al. 2003; Oleson et al. 2004), developed by NCAR in collaboration with scientists from the academic community. While CAM3–CLM3 can be integrated with ocean and sea ice models in the fully coupled framework [the Community Climate System Model (CCSM)], in this study we prescribe the climatological, monthly varying sea surface temperature and sea ice coverage.

CAM3 simulates the physics and dynamics of the atmosphere. For the dynamics of the atmosphere, CAM3 includes Eulerian spectral, semi-Lagrangian dynamics, and finite-volume (FV) dynamics. In this study, we employ the FV dynamical core with a horizontal resolution of 2° latitude by 2.5° longitude and a total of 26 levels in the vertical direction. CLM3 simulates energy, moisture, and momentum fluxes between vegetation, soil, and the atmosphere. It has 10 unevenly spaced vertical soil layers, up to 5 snow layers, and 1 vegetation layer. Land surface within each grid cell is represented by the fractional coverage of four types of patches (glacier, lake, wetland, and vegetated), and the vegetation portion of the grid cell is by the fractional coverage of up to four different plant functional types (PFTs). The leaf phenology scheme in the publicly available version of CLM3 is replaced with an improved scheme that has been validated against the latest observation data (Kim and Wang 2005). Leaf area index (LAI) is updated daily in response to both cumulative and concurrent hydrometeorological conditions, by scaling down the annual maximum LAI derived from the Moderate Resolution Imaging Spectroradiometer (MODIS) data (Tian et al. 2004).

b. Model performance

Performance of CAM3–CLM3 was assessed against observational data in Hack et al. (2006). While model validation is not the focus of this study, a few key hydrologic variables from a 20-yr integration of CAM3–CLM3 and their comparison with the North American Regional Reanalysis (NARR) data (Mesinger et al. 2006) are presented to help better understand the experimental simulations we perform in this study. Here the CAM3–CLM3 integration is driven with the inter-annually varying SST from 1979 to 1998 and runs at a 2° × 2.5° resolution. Although the NARR is available at the 32-km resolution for 25 yr (1979–2003), here the NARR data from 1979 to 1998 are regridded onto a 2° × 2.5° resolution to be consistent with the results of CAM3–CLM3.

For precipitation and moisture convergence, the mean and standard deviation during June, July, and August are presented; for volumetric soil water content, the mean and standard deviation on 1 June are presented. This is so chosen because later in this paper, we mainly focus on the response of monthly precipitation to changes in soil moisture on the first day of each month. Dry biases of all three fields in CAM3 relative to NARR are evident particularly over the interior continent (Fig. 1). The standard deviation of volumetric soil water content in CAM is less than about 50% of that in NARR (Fig. 2). While the volumetric soil water content is presented here due to the data availability of NARR dataset, the saturation level will be used to represent the soil moisture condition throughout the rest of this paper.

c. Methodology

A 12-yr model integration is first carried out using the coupled CAM3–CLM3 model driven with climatological SSTs. With the first 2 yr devoted to model spinning up, the last 10 yr of simulation, however short it may be, can provide a rough estimate of model climatology. Due to the lack of soil moisture observational data, soil moisture climatology for the first day of each month is derived based on the last 10 yr of the model integration. This model-based soil moisture climatology is used to initialize the model in subsequent simulations. Except for soil moisture, the atmospheric and land surface states on the first day of each month in the last year of the 12-yr model integration are used. By driving the model with the climatological SST, the experimental simulations are carried out under normal large-scale forcing, so that we can focus on the impact of initial soil moisture anomalies.

A large group of sensitivity experiments are then per-
formed to examine the impact of initial soil moisture anomalies of different magnitudes applied in different months over domains of different sizes and at different soil depths. For any given soil moisture anomaly in a given month, we perform two five-member ensemble simulations: the Control ensemble and the SM Anomaly ensemble. In the Control ensemble, the initial soil moisture is set to a level close to the model climatology for the first day of the corresponding month (at 100%, 99%, 98%, 97%, and 96% of the climatological soil moisture, respectively); in the SM Anomaly ensemble, the initial soil moisture in each simulation is higher or lower than its counterpart in the Control ensemble by the same amount (e.g., set to 20%, 19%, 18%, 17%, and 16% of the climatological soil moisture, respectively, for an 80% dry anomaly case). While previous studies often initialize the model with uniform soil saturation degree over the experiment domain, here we initialize the model with a certain percentage of climatological soil moisture. As Georgescu et al. (2003) documented, changing the initial soil moisture spatial distribution influences future simulated precipitation. By using a percentage of climatology, we intend to focus on the impact of change in the soil moisture magnitude, and exclude the impact of change in the soil moisture spatial distribution.

Initial soil moisture anomalies are applied as a certain percentage increase or decrease of soil moisture
climatology on the first day of each month. Figure 3 shows the percentage increase and decrease of soil moisture climatology that is required in order for soil moisture to reach the field capacity and wilting point, respectively. Over most of North America, more than 80% increase of climatology is required to reach the field capacity; about 20% decrease of climatology is needed to reach the wilting point. Here, this estimate is based on the soil moisture in the third soil layer as an example, and results for other layers are similar (not shown).

The vegetation phenology is simulated in the Control with the predictive phenology scheme, but in the SM Anomaly the vegetation phenology is prescribed based on that simulated in the Control. Therefore, the climate anomalies in the SM Anomaly ensembles relative to the Control ensembles are attributed to the impact of soil moisture initialization through soil moisture–precipitation feedback alone (with vegetation feedback excluded). The impact of vegetation feedback will be explored in a separate paper (Kim and Wang 2007). Here the simulations are carried out from the first day of each month to the end of that year.

While CAM3 simulates the climate globally, the focus of this study is on North America only. As briefly introduced in section 1, in addition to the timing and magnitude/direction of soil moisture anomalies, the impact of the spatial extent (in both horizontal and verti-
cal directions) of the anomalies will be examined as well. We consider three domains of different sizes (Fig. 4), where soil moisture anomalies will be applied: North America (“NA”), the Mississippi River basin (“MR”), and the Upper Mississippi River basin (“UMR”). Unless otherwise specified, initial soil moisture anomalies are applied over NA throughout the whole soil depth in the model (up to tenth soil layer, ~3.4 m). Here the shaded area presents the Mississippi River basin with a resolution of 2° latitude by 2.5° longitude according to Global Energy and Water Cycle Experiment (GEWEX) Americas Prediction Project (GAPP) (Bosilovich and Chern 2006), where our results analysis in sections 3 and 4 will focus on.

Fig. 3. (a) Percentage increase of soil moisture climatology required for soil moisture to reach the field capacity and (b) percentage decrease of soil moisture climatology required for soil moisture to reach the wilting point on the first day of each month from April to June.
3. Role of soil moisture during spring versus summer

This section focuses on how the spring and summer soil moisture anomalies modify the regional climate using April and June as examples. A large magnitude of soil moisture anomalies, namely, 80% increase or decrease, is applied over the large domain (NA in Fig. 4) throughout the entire soil depth to ensure a clear signal. All results presented here as well as in subsequent sections are averages among five ensemble members. In the monthly and spatial results, shading is applied to areas where the anomalies pass the 10% significance test, which is performed with the precipitation statistics estimated based on the 20-yr integration driven by SST from 1979 through 1998 (see section 2b).

The impact of soil moisture anomalies at the beginning of June on subsequent precipitation is shown in Fig. 5. Although the positive (negative) soil moisture anomalies are applied to the initial condition on 1 June only, it leads to a substantial increase (decrease) in precipitation during the following months, indicating a positive feedback between soil moisture and precipitation. The impact of initial soil moisture anomalies persists for at least 2–3 months. This persistence will be further examined using the daily results in section 4.

While initializing the model with a large magnitude of soil moisture anomalies (i.e., 80% decrease/increase of soil moisture climatology) favors a clear signal of climate response to the initial soil moisture anomalies, such large magnitude of anomalies may be beyond the realistic range of soil moisture anomalies. We therefore also experimented on various smaller magnitudes of initial soil moisture anomalies. Here precipitation anomalies during June, July, and August in the SM Anomaly ensembles initialized with a 30% decrease and a 30% increase of soil moisture climatology on 1 June, respectively (relative to the Control ensembles), are given in Fig. 6. Precipitation response to the initial soil moisture anomalies equivalent to 30% and 80% of soil moisture climatology (Fig. 5 versus Fig. 6) shows similar spatial distributions. As the magnitude of soil moisture anomalies decreases dramatically, the resulting precipitation anomalies decrease only slightly. Further details about the impact of magnitude of soil moisture anomalies will be examined later in section 4.

In terms of spatial coverage, although soil moisture anomalies are applied over NA, precipitation response is concentrated over only part of the domain. The region of the precipitation response is consistent with the region of strong soil moisture–precipitation coupling in CAM3–CLM3 (Wang et al. 2007). The dry (wet) soil moisture anomalies cause an increase (decrease) in surface temperature through the suppression (enhancement) of evaporative cooling (Fig. 7a) and consequently decrease (increase) in sea level pressure (Fig. 7b). Such local changes, in turn, modify the large-scale circulation, as shown by anomalies in the moisture con-

Fig. 4. Map of study area. Three boxes present three different domains of anomalies: North America (“NA”), the Mississippi River basin (“MR”), and the Upper Mississippi River basin (“UMR”). The shaded area defines the Mississippi River basin on a resolution of 2° latitude by 2.5° longitude (Bosilovich and Chern 2006).
vergence field (Fig. 7c). Changes of pressure field lead to weaker (stronger) westerlies and its northward (southward) shift in the dry (wet) SM Anomaly experiments. The belt of high precipitation, therefore, moves northward (southward) (Fig. 5a). These results are consistent with the findings of Oglesby and Erickson (1989), who used CCM1 to simulate the climate anomalies with “desert-like” initial soil moisture conditions over extensive areas of North America. In dry (wet) cases, decreases (increases) in pressure in the northern part of the domain lead to moisture divergence (convergence) over the southern part (Great Plain), intensifying the impact of local dry (wet) soil moisture anomalies.

As a comparison to Fig. 5, Fig. 8 demonstrates the impact of soil moisture anomalies at the beginning of April on precipitation in subsequent months, namely, April, May, June, and July. It differs from the experiments starting on 1 June in that most of the large-magnitude responses in precipitation are delayed. Major responses are found in June or later, although some responses in precipitation are present during April and May. Changes in local soil moisture condition propagate into changes in large-scale circulation. In the wet case, extensive (yet relatively small) increase of precipitation is found; in the dry case, however, the response in April is mostly small in magnitude and lacks spatial coherency. In both cases, large-magnitude precipitation response is generally delayed. Such delayed response is due to the fact that precipitation in April is still in the winter regime; that is, local convection is not a dominant mechanism in generating precipitation. In April,
the amount of local convective precipitation, about the
same as the amount of large-scale precipitation, is less
than half of that in summer months (Fig. 9a). Changes
in evapotranspiration due to soil moisture anomalies
influence the planetary boundary layer, which in turn
influence local convective activities that produce most
of the summertime precipitation over the domain of
study. On the contrary, such local anomalies of soil
moisture have minimal impact on local precipitation
during April when local convective precipitation does
not dominate. The persistence of late-spring soil mois-
ture anomalies (not shown) therefore leads to changes
of precipitation later (during summer) but not immedi-
ately following the anomalies.

4. Sensitivity to characteristics of soil moisture
anomalies

This section documents how the timing, magnitude,
spatial coverage, and depth of soil moisture anomalies
influence the precipitation responses they trigger, based on a series of sensitivity experiments tabulated in Table 1. Pal and Eltahir (2001), using a regional climate model (RegCM), examined the sensitivity of precipitation response to the timing and magnitude of soil moisture anomalies. They used a small Midwest domain, and integrated the regional model for one month. Here we use a global climate model CAM3–CLM3 and integrate the model from spring and summer months to the end of that year. The longer model integration allows us to examine the persistence of land–atmosphere anomalies in addition to their magnitudes.

a. Timing of anomaly

The impact of soil moisture anomalies varies with timing. As presented in section 3, local soil moisture plays a critical role during summer when convective activities in the study domain are most pronounced, and is less important in other seasons (e.g., early spring). Such seasonality of the feedbacks between soil moisture and precipitation is investigated here. We examine the impact of initial soil moisture anomalies imposed on 1 April, 1 May, 1 June, 1 July, and 1 August. For each starting date, the Control and 80% dry/wet SM Anomaly simulations are carried out until the end of December (Table 1). Figure 10 presents the 10-day running average of precipitation difference between the SM Anomaly and Control ensembles, averaged over the Mississippi River basin (shaded area in Fig. 4). Overall, the positive feedback between soil moisture and precipitation clearly exists during summer, and the resulting memory of the coupled land–atmosphere system lasts for more than 2 months except for the August case. As suggested in section 3, the precipitation responses over the Mississippi River basin are delayed with the springtime anomalies, although slight precipitation changes exist during spring. Further, the persis-
tence of precipitation anomalies is significantly shorter in August than in other summer months. Starting from September, local convective precipitation over this region decreases dramatically to about half of its summertime amount in the model as shown in Fig. 9a. Therefore, the soil moisture anomalies imposed in August cannot be translated into precipitation anomalies after (including) September.

In simulations starting on 1 April, in terms of magnitude, the precipitation response to dry soil moisture anomalies is more significant than to wet soil moisture anomalies during summer. In simulations starting from the summer months, however, the precipitation response to wet soil moisture anomalies is more significant in magnitude during the first month following the initial soil moisture anomalies than to dry anomalies.
Such asymmetry is highly related to the differences in the condition between spring and summer, with spring being much wetter than summer (see Fig. 9b). A wet regime tends to be more sensitive to dry anomalies and a dry regime tends to be more sensitive to wet anomalies. While the magnitude of precipitation response to the wet soil moisture anomalies is generally larger than to the dry ones, the persistence of precipitation response does not seem to differ when results from a single ensemble are measured against the 90% confidence interval as presented in Fig. 10. However, in summer cases, it is noticeable that the precipitation response to initial wet soil moisture anomalies peaks during the first month and quickly dissipates (within a month or so), while the precipitation anomalies in response to initial dry soil moisture anomalies remain at a considerable magnitude for a longer time. This asymmetry in the persistence of a climate anomaly, to be further illustrated in section 4b, has to do with the fact that soil dries up from spring to summer in this region (see Fig. 9b). Evaporation is higher than precipitation and potential evaporation is even higher. When a wet anomaly occurs, the extra water in the soil is quickly removed by the fast evapotranspiration process; when a dry anomaly occurs, the void soil storage cannot be replenished before evaporation slows down (to a level below precipitation) and soil moisture rebounds during the fall.

In the April simulations, snow cover can potentially modify the interaction between soil moisture and precipitation. Snow melting may offset some of the dry soil moisture anomalies or reinforce wet soil moisture anomalies. On the other hand, snow cover may decouple the interaction between soil moisture and precipitation by preserving the underlying soil moisture anomalies. However, in our simulations, snow presence on 1 April is rare south of 60°N (not shown). The impact of snow cover in the April simulations is therefore minimal.

b. Magnitude of anomaly

As mentioned in section 2, initial soil moisture anomalies are expressed as a percentage of climatology. A series of simulations are carried out with the decrease of 15%, 30%, 50%, and 80%, and increase of 30%, 50%, 80%, and 100% of soil moisture climatology using 1 May, 1 June, and 1 July as examples (see Table 1). Based on the daily output of the 20-yr CAM3–CLM3 integration (1979–98) (see section 2b), the mean saturation level of soil (i.e., the volumetric water content divided by the maximum volumetric water content) is 43% in the top 50 cm of soil depth over the Mississippi River basin (i.e., the shaded area in Fig. 4).

Among the grid cells within the Mississippi River basin, in a cell where soil saturation level varies most, its mean, maximum, and minimum are 43%, 81%, and 32%, respectively (i.e., the maximum and minimum are 88% above and 25% below the mean soil moisture, respectively). In the least variable cell, the mean, maximum, and minimum saturation levels are 44%, 57%, and 38%, respectively (i.e., the maximum and minimum are 29% above and 13% below the mean, respectively). Note that the variability of soil moisture is largest in the top soil layer and decreases with depth. For example, in the most variable cell, the maximum saturation level is 172% above the mean in the second layer (between 1.8 and 4.5 cm of soil depth) and 60% above the mean in the seventh layer (between 49 and 83 cm). Here we also provide the statistics from the Illinois State Water Survey (http://climate.envsci.rutgers.edu/soil_moisture/illinois.html; Hollinger and Isard 1994) to compare the model soil moisture with observations, although they are not directly comparable due to their discrepancies in the spatial and temporal resolutions. The Illinois State Water Survey provides the roughly biweekly measurements of soil moisture from 1981 to 2004 at 19 stations. At a station where soil saturation level varies most among the 19 stations, the mean, maximum, and minimum saturation levels averaged for the top 50 cm
are 66%, 100%, and 8%, respectively (i.e., the maximum and minimum are 52% above and 87% below the mean soil moisture, respectively). In the least variable cell, the mean, maximum, and minimum saturation levels are 78%, 98%, and 49%, respectively (i.e., the maximum and minimum are 26% above and 37% below the mean, respectively).

Figure 11 presents monthly precipitations corresponding to the initial soil saturation over the Mississippi River basin, based on the different SM Anomaly experiments. Precipitation is almost linearly correlated to antecedent soil moisture conditions in a certain range of initial soil moisture anomalies (i.e., within $\pm 30\%$ to $\pm 50\%$ of soil moisture climatology). Beyond this range, the sensitivity drops substantially. At the low end of soil moisture, although the evaporation regime is soil moisture controlled, further decrease of initial soil moisture beyond the wilting point is not

<table>
<thead>
<tr>
<th>Name of ensemble</th>
<th>Start date</th>
<th>Initial soil moisture (% of climatology)</th>
<th>Region of anomaly</th>
<th>Soil layers of anomaly</th>
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<tr>
<td></td>
<td>C_May</td>
<td>1 May</td>
<td>NA</td>
<td>Up to 10th layer</td>
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<tr>
<td></td>
<td>C_Jun</td>
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<td>(343.3 cm)</td>
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<td></td>
<td>C_Jul</td>
<td>1 July</td>
<td>NA</td>
<td>Up to 10th layer</td>
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<td></td>
<td>C_Aug</td>
<td>1 August</td>
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<td></td>
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<tr>
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<tr>
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<td>SM_D50_Jun</td>
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<td>50, 49, 48, 47, and 46</td>
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physical and does not actually reduce evapotranspiration. At the high end, atmospheric conditions, especially radiation, are limiting and potential evaporation controls the actual evapotranspiration from the surface. Increase of soil moisture beyond a certain level therefore can no longer promote evaporation. At both the low and high ends of soil moisture, low sensitivity of evapotranspiration to soil moisture changes results in low sensitivity of precipitation. Therefore the high sensitivity to changes of soil moisture within $\sim 30\%$ to $\sim 50\%$ of its climatology is consistent with Fig. 3 in section 2, which shows that the soil moisture climatology on the first day of each month is closer to wilting point (about 20\% lower than climatology) than to the field capacity (more than 80\% higher than climatology). In addition, precipitation anomalies in the 30\% dry and wet cases are more than half of precipitation anomalies in the 80\% dry and wet cases as presented in Fig. 6. This indicates that our conclusions based on results from experiments with a large magnitude of soil moisture anomalies would remain valid for moderate anomalies (30\% decrease or increase).

From Fig. 12 (which uses experiments starting on 1 July as an example), we make two observations. First, changes in precipitation due to dry anomalies in initial soil moisture last longer than those due to wet anomalies during the summer months. For example, precipitation anomalies became statistically insignificant by
mid-August in the W30 and W50 ensembles, and by the end of August in the D30 and D50 ensembles. Although such asymmetry in response is somewhat evident in Fig. 10 and mentioned in section 4a, results based on soil moisture anomalies of different magnitudes confirm the robustness of this finding. It is also consistent with observations (Diaz 1983) that droughts tended to last longer than floods over much of the United States. Second, the magnitude of precipitation response to summer wet anomalies is larger than to dry anomalies of the same magnitude during the first month (July), but dissipates rather quickly. This also confirms some of the observations from Fig. 10.

c. Spatial coverage of anomaly

So far the results presented were based on ensemble simulations with different initial soil moisture conditions on 1 May (circle), 1 June (square), and 1 July (diamond). The error bar represents a standard deviation of area-average precipitation among five ensemble members; the x axis indicates saturation level of soil (%) in the top 50-cm soil depth corresponding to different initial soil moisture conditions. The filled circle, square, and diamond present results from the Control ensembles (i.e., ensembles initialized with the soil moisture climatology).

mid-August in the W30 and W50 ensembles, and by the end of August in the D30 and D50 ensembles. Although such asymmetry in response is somewhat evident in Fig. 10 and mentioned in section 4a, results based on soil moisture anomalies of different magnitudes confirm the robustness of this finding. It is also consistent with observations (Diaz 1983) that droughts tended to last longer than floods over much of the United States. Second, the magnitude of precipitation response to summer wet anomalies is larger than to dry anomalies of the same magnitude during the first month (July), but dissipates rather quickly. This also confirms some of the observations from Fig. 10.

c. Spatial coverage of anomaly

So far the results presented were based on ensemble simulations with different initial soil moisture conditions imposed across NA. Here we examine the impact of different spatial coverage of soil moisture anomalies centering around the Mississippi River basin, and focus on simulations starting on 1 May as an example. Initial anomalies with an 80% decrease of climatology are imposed over NA, MR, and UMR, respectively (see Fig. 4; Table 1). Here the MR and UMR are chosen according to areas showing the most pronounced precipitation response to initial soil moisture anomalies over NA. Figure 13 presents the simulated monthly precipitation difference between the SM Anomaly and Control ensembles. When the dry anomalies are applied to NA, subsequent decrease of precipitation in May is found over only part of NA, including primarily the UMR. Reducing the anomalies’ domain to the MR leads to a smaller magnitude of the precipitation response, but the spatial extent and location of the precipitation response especially in June do not change much. When a dry anomaly of the same magnitude is applied to the UMR only, rainfall decrease in May has a similar spatial pattern to the case of large domain anomaly, but is of a much smaller magnitude and is less persistent—little response is detected in June. It suggests that precipitation response to soil moisture anomalies of the same magnitude on subsequent precipitation can vary significantly with the spatial coverage of such anomalies. A large-scale drought/flood will last longer and is generally more severe. Note that the location of domain will matter and the finding in this subsection may hold only when the area of small domains is chosen according to the precipitation response to soil moisture anomalies over large domain.

d. Depth of anomaly

The impact of soil moisture vertical profile in the context of land–atmosphere interactions has been studied using observational data (Wu et al. 2002) and climate models (Oglesby et al. 2002; Wu and Dickinson 2004). Oglesby et al. (2002) suggested that the soil moisture profile plays the most significant role in soil moisture–precipitation feedback, and soil moisture in the deep soil zone has a more powerful and persistent impact. CLM has 10 unevenly spaced soil layers as introduced in section 2. The thickness of each layer exponentially increases with depth, and the total soil depth is ~3.4 m. All results presented up to this point are based on simulations with soil moisture anomalies imposed over the entire depth.

To examine the role of soil moisture anomalies in deep soil versus those in shallow soil, we carry out SM Anomaly simulations with anomalies applied to the top 7 layers only (top 83 cm) instead of all 10 layers (i.e., whole depth: ~3.4 m). They are initialized with 80% decrease and increase of soil moisture climatology, respectively, starting on 1 May (SM_D80_May_7 and SM_W80_May_7 in Table 1). Compared with results from the experiments on initial anomalies over the whole soil depth, the spatial pattern of precipitation anomalies is quite similar to each other (not shown). Figure 14 shows that precipitation anomalies following soil moisture anomalies in shallow soil are slightly
smaller in magnitude and similar in persistence, compared to those following moisture anomalies in the whole soil depth. Precipitation anomalies damp quickly after about 2 months when the initial anomalies are applied to the top seven layers only.

Figure 15 presents the daily soil moisture difference in different soil layers. The deeper the soil layer is, the slower the soil moisture anomalies are to dissipate (especially for the dry cases). Dry anomalies persist for more than 7 months (the length of the whole model integration). It is evident that the shallow soil anomalies alone can persist and impact the future climate in the monthly and seasonal time scale, while the deep-soil anomalies further enhance and prolong this impact. Results from all previous experimental simulations, therefore, would be qualitatively the same if we were to apply initial soil moisture anomalies to shallow soil layers only. It is mainly because grasses and crops, the dominant plant types over the Mississippi River basin, have about 90% and 95% of their roots, respectively, in the top seven layers (top 83 cm). Consequently, soil moisture in the top several layers plays a critical role in controlling evapotranspiration, and thus precipitation.

5. Conclusions and discussion

In this study, we carried out a series of numerical experiments using a modified version of CAM3–CLM3 to investigate the impact of soil moisture anomalies on subsequent precipitation over North America. First, we examined the mechanisms related to the impact of spring and summer soil moisture on summer precipitation by starting simulations on 1 April and 1 June. The positive feedback between soil moisture and precipitation was shown and its relation with the local and large-scale dynamics was examined. We then investigated how the impact of initial soil moisture anomalies depends on the characteristics of such anomalies, including the timing, magnitude, spatial coverage, and vertical depth. The main findings are as follows. 1) The primary impact of spring soil moisture anomalies is not evident until early summer, although the impact of anomalies on the large-scale circulation results in slight changes in precipitation during spring. This delay is due to the difference in precipitation regimes between spring and summer. 2) The amount of precipitation is almost linearly correlated with the amount of antecedent soil

Fig. 12. Ten-day running average of precipitation anomalies (SM Anomaly − Control) as a response to soil moisture anomalies applied on 1 July. D15 represents the dry SM Anomaly ensemble starting with a 15% decrease of soil moisture climatology, D30 for 30%, D50 for 50%, and D80 for 80%. W30 represents the wet SM Anomaly ensemble starting with a 30% increase of soil moisture climatology, W50 for 50%, W80 for 80%, and W100 for 100%. Each line presents the ensemble mean of five members averaged over the Mississippi River basin. The shaded area presents the 90% confidence interval for the 10-day average of precipitation, estimated from a 20-yr model integration with SST varying from 1979 to 1998.
moisture within the active range of soil moisture. The precipitation response shuts off as the initial soil moisture approaches the wilting point on the dry end and field capacity on the wet end. 3) During the first month following the onset of summer soil moisture anomalies, the precipitation response to wet anomalies is larger in magnitude than that to dry anomalies. However, the resulting wet anomalies in precipitation quickly dissipate within a month or so, while the resulting dry anomalies in precipitation remain at a considerable magnitude for a longer period of time. Wet spring anomalies, however, are likely to be ameliorated before summer, thus have a smaller impact (in magnitude) on summer precipitation than dry spring anomalies. 4) Soil moisture anomalies of smaller spatial coverage lead to precipitation anomalies that are smaller and less persistent, compared to anomalies at the continental scale. 5) Anomalies in the shallow soil can persist long enough to influence the subsequent precipitation at the seasonal time scale. Dry anomalies in deep soils last much longer (by months) than those in shallow soils.

Note that the model sensitivity to certain changes (e.g., soil moisture anomalies) depends on the mean climate of model. Such dependence on the model climatology is important for understanding different results from different models. Therefore, it would be worthwhile to examine our findings, particularly the finding that precipitation response during the first month following summer dry anomalies is slightly smaller in magnitude than that following wet anomalies, after correcting the dry biases of CAM3–CLM3 over North America. The results with different spatial coverage of soil moisture anomalies imply that the spatial extent of droughts or floods matters. Large-scale droughts or floods would be more severe and more persistent than small-scale ones.

The response of precipitation to soil moisture anomalies in an individual model such as CAM3–CLM3 depends on the strength of soil moisture–precipitation coupling in the model. The latter is highly model dependent. Among the 12 widely used GCMs examined in Koster et al. (2004, 2006), CAM3–CLM3 is one of a few models in which the soil moisture–precipitation coupling is strong. Considering the impact of soil moisture from both shallow and deep soil layers, the regions of strong coupling in CAM3–CLM3 include a large portion of the Mississippi River basin (Wang et al. 2007). Other models showing strong coupling in North America include the Geophysical Fluid Dynamics Laboratory (GFDL) model and the NSIPP model.

Fig. 13. Monthly precipitation anomalies (SM Anomaly – Control) in (top) May and (bottom) June as a result of an 80% decrease of soil moisture climatology applied on 1 May to different regions: (a) NA, (b) MR, and (c) UMR. Shading is applied to areas where the anomalies pass the 10% significance test. The numbers in the left bottom of each panel indicate averages over the Mississippi River basin (shaded area in Fig. 4).
Therefore, the response of precipitation to soil moisture anomalies we document here is probably comparable to that in GFDL and NSIPP, but stronger than that found in many other GCMs with a weak soil moisture–precipitation coupling. Whether the coupling in the real world is strong or weak has yet to be addressed.

The impact of soil moisture on subsequent precipitation has rarely been examined based on observational data (e.g., Findell and Eltahir 1997; Salvucci et al. 2002; D’Odorico and Porporato 2004), due to the limited soil moisture data in both the temporal and spatial domains. At the local scale, Findell and Eltahir (1997) and Salvucci et al. (2002) studied the soil moisture data from the Illinois State Water Survey (Hollinger and Isard 1994), but drew contradictory conclusions about the relationship between soil moisture and future precipitation. Findell and Eltahir’s (1997) linear correlation analysis suggested a significant positive lagged correlation between soil moisture and precipitation during the summer months, while Salvucci et al.’s (2002) Granger causality analysis found no evidence for the impact of soil moisture on precipitation. At the continental scale, Koster et al. (2003) compared the variances of observed precipitation with those simulated by the NSIPP AGCM with different land–atmosphere feedback treatments. The model reproduced the spatial pattern of observed precipitation variance when the land–atmosphere feedback was enabled, and lost its capability to reproduce this observed precipitation characteristics when the land–atmosphere feedback was turned off. This provided suggestive, although not definitive, evidence of feedback between soil moisture and precipitation. However, Ruiz-Barradas and Nigam (2005) showed that the contribution of local evaporation to local precipitation (precipitation recycling) is overwhelmingly high in CAM and NSIPP, compared to observationally constrained data such as NCEP reanalysis. Therefore, our simulated response of precipitation to soil moisture initialization may be stronger than observations, if the soil moisture–precipitation coupling strength in the reanalysis data is realistic. Unfortunately, lack of soil moisture data makes it difficult to draw a firm conclusion.

Snow cover in our ensemble simulations is initialized with the value on the first day of each month in the last year of the 12-yr initial integration similar to other at-

Fig. 14. Ten-day running average of precipitation anomalies (SM Anomaly – Control) for simulations with an 80% decrease (solid lines) or 80% increases (dashed lines) of soil moisture climatology applied on 1 May. The thin lines represent the simulations with anomalies in the whole soil depth (~3.3 m) and thick lines in the top seven layers only (83 cm). Each line presents the ensemble mean of five members averaged over the Mississippi River basin. The shaded area presents the 90% confidence interval for the 10-day average of precipitation, estimated from a 20-yr model integration with SST varying from 1979 to 1998.
Fig. 15. Daily soil moisture anomalies (soil saturation level in %) in the 1st, 3d, 5th, and 7th soil layers (SM Anomaly – Control) for simulations with an 80% decrease (solid lines) or 80% increase (dash lines) of soil moisture climatology applied on 1 May. The thin lines represent the simulations with anomalies over the whole soil depth (~3.3 m) and thick lines up to the top seven layers only (83 cm). Each line presents the ensemble mean of five members averaged over the Mississippi River basin. The shaded area presents the 90% confidence interval for the daily average of soil moisture in each soil layer, estimated from a 20-yr model integration with SST varying from 1979 to 1998.
mospheric and land surface state variables except for soil moisture. Although spring-to-summer soil moisture may be closely related to snow cover during winter and spring, only initial soil moisture is varied to focus on the impact of initial soil moisture condition on climate. Changes in snow cover during spring would change the soil moisture condition due to snow melting; changes in snow cover also lead to changes in surface reflectivity, and thus surface temperature, which is significant with ample supply of shortwave radiation (Schlosser and Mocko 2003). If initial snow cover is varied together with initial soil moisture in our study, we expect to see largest changes in simulations starting on 1 April. Since the widespread ablation of snow cover over the Northern Hemisphere occurs during April (Schlosser and Mocko 2003), adding the snow cover anomalies is equivalent to changing the magnitude of soil moisture anomalies in April. In the dry anomaly case, decrease in snow cover will reduce water supply to soil through snow melting, and also reduce albedo and then increase surface temperature. Such changes in soil moisture (primary) and temperature (secondary) will amplify the impact of decrease in initial soil moisture.

Convective precipitation is dominant during summer over North America, and is critical in the feedback between soil moisture and precipitation. While convective precipitation takes place at the small scale [less than 100 km$^2$ (Pitman et al. 1990)], the model in our study is run with a horizontal resolution of 2° latitude by 2.5° longitude. In this study, the use of a rather coarse-resolution GCM allows for a sufficient integration time and adequate feedback between the local condition and large-scale dynamics. GCM simulations with a finer resolution or regional climate simulations are desirable to solve the processes at the convective storm level. Moreover, we investigated the impact of initial soil moisture anomalies under the climatological SST. Since extreme climate events such as the 1930s Dust Bowl, 1988 U.S. drought, and 1993 U.S. flood are often shown to be associated with anomalies in both land condition and oceanic forcing (e.g., Schubert et al. 2004), numerical experiments in specific years with the observed time-varying SST are desirable to better understand the role of soil moisture anomalies in droughts and floods (Dirmeyer 2005). Finally, SM Anomaly simulations in this study are carried out with the prescribed seasonal variation of LAI as mentioned in section 2. The anomalies in soil moisture conditions, however, also impact the state of vegetation. The resulting response in LAI, in turn, will further impact the soil moisture and precipitation. Further research will tackle how the feedback from vegetation influences the land–atmosphere interactions at the seasonal time scale (Kim and Wang 2007).

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