Pathways Relating Soil Moisture Conditions to Future Summer Rainfall within a Model of the Land–Atmosphere System

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ABSTRACT

In this paper, the key pathways and mechanisms through which soil moisture conditions affect future rainfall over the U.S. Midwest are investigated using a regional climate model. A series of numerical experiments are performed to identify these pathways using the drought of 1988 and flood of 1993 as representative events.

The results suggest that the soil moisture–rainfall feedback is an important mechanism for hydrologic persistence during the late spring and summer over the midwestern United States. They indicate that the feedback between soil moisture and subsequent rainfall played a significant role in enhancing the persistence of the drought of 1988 and the flood of 1993. It is found that there is a pronounced asymmetry in the sensitivity of simulated rainfall to specified initial soil moisture. The asymmetry acts to favor a stronger soil moisture–rainfall feedback during drought conditions as opposed to flood conditions.

Detailed analyses of the simulations indicate that the impact of soil moisture on both the energy and water budgets is crucial in determining the strength of the soil moisture–rainfall feedback. Anomalously high soil moisture tends to 1) increase the flux of high moist static energy air into the planetary boundary layer from the surface via an increase in net surface radiation, 2) reduce the planetary boundary layer height thus increasing the moist static energy per unit mass of air, and 3) reduce the amount of entrained air of low moist static energy from above the planetary boundary layer. Each of these effects are additive and combine to increase the moist static energy per unit mass of air in the planetary boundary layer. This increase results in an increase in the frequency and magnitude of convective rainfall events and a positive feedback between soil moisture and subsequent rainfall.

1. Introduction

In 1988, the United States experienced its warmest and driest summer since 1936 (Ropelewski 1988). It resulted in approximately 10 000 deaths from heat stress and caused an estimated $30 billion in agricultural damage (Trenberth and Branstator 1992). The 1993 summer flooding over the midwestern United States was one of the most devastating floods in modern history (Kunkel et al. 1994). Record high rainfall and flooding occurred throughout much of the Upper Mississippi River basin and persisted for long periods. The National Oceanic and Atmospheric Administration (NOAA) estimated that the flood caused $15–20 billion in damages (NOAA 1993). Such extreme events result in large financial losses for agricultural communities and cause additional hardship and personal loss in many urban communities.

Figure 1 shows maps of observed rainfall anomalies averaged over May, June, and July 1988 and 1993 for the continental United States and parts of Canada and Mexico. In 1988, negative anomalies associated with the widespread drought covered most of the eastern two thirds of the United States. The drought was especially pronounced in the Great Plains and Midwest (light shading). Contrary to the late spring and summer of 1988, rainfall in the late spring and summer of 1993 was anomalously high over much of the United States except along the Gulf Coast and eastern seaboard states. The primary area of flooding was in the Upper Mississippi River basin (dark shading). Not surprisingly, extreme soil moisture conditions were associated with these extremes in rainfall. Figure 2 shows time series (1981–93) of the Illinois State Water Survey (ISWS) near surface soil saturation data averaged over the state of Illinois (Hollinger and Isard 1994). Clearly, 1988 (lower solid line) was the driest spring and summer, in terms of soil saturation, on record, while 1993 (upper solid line) was the wettest. These extremes in soil moisture are likely to be more apparent over Iowa and Missouri where the flood and drought were more extreme than in Illinois (where data are available) during their respective years.

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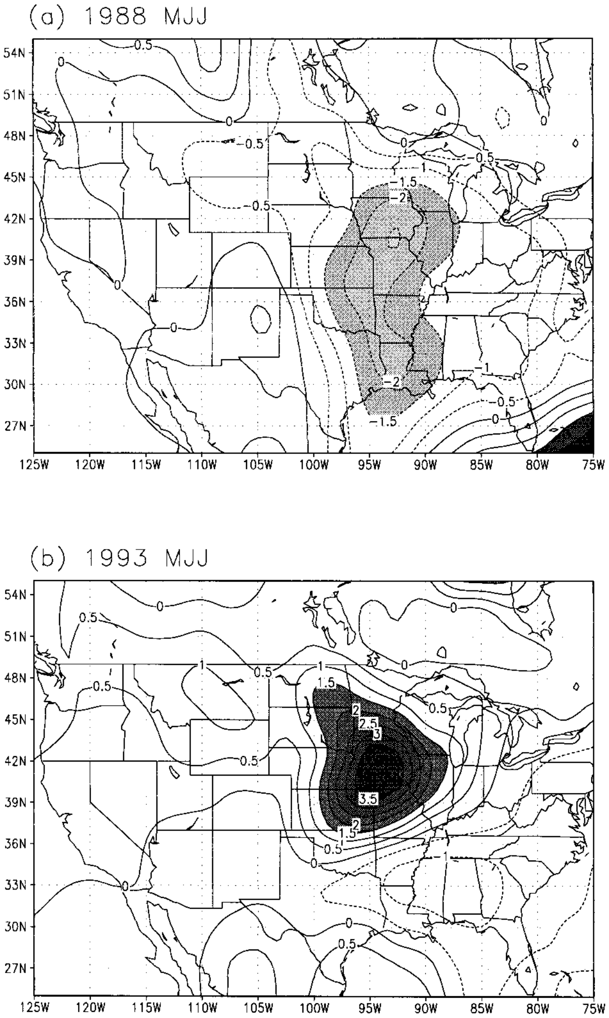

FIG. 1. Rainfall anomalies (mm day^{-1}) averaged over May, Jun, and Jul for 1988 and 1993 based on the Xie and Arkin (1996) merged dataset: (a) 1988 and (b) 1993. Shading occurs above and below anomalies of 1.5 mm day^{-1}.

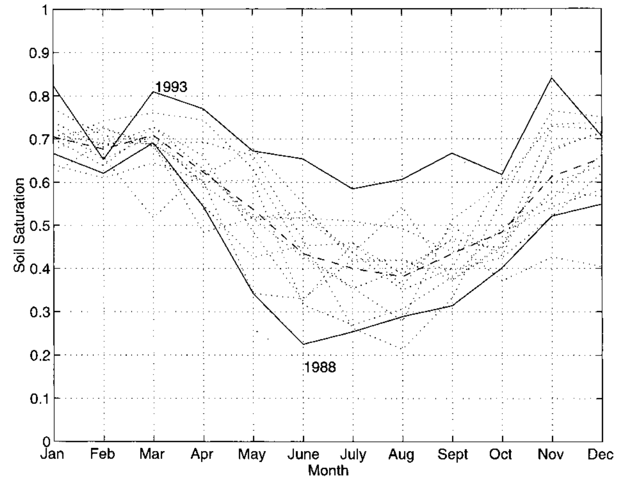

FIG. 2. Illinois State Water Survey monthly averaged soil saturation from 0 to 10 cm for 1981–93. The lower solid line is the soil saturation for 1988, the upper solid line is for 1993, the dotted lines are the rest of the years, and the dashed line is the average of all of the years.

This study attempts to determine whether the anomalous soil moisture conditions observed in 1988 and 1993 played a role in initiating and/or enhancing the extremes in observed rainfall or whether they were simply a by-product of these extremes. More specifically, it investigates the pathways through which late spring and summer soil moisture conditions impact extreme North American summertime flood and drought. Understanding these pathways could potentially have significant implications for water resources management and cropping strategies and also help to reduce human and economic losses. If, for example, one can predict that a given summer will receive anomalously high rainfall, then water can be released from reservoirs in advance to reduce damaging floods. Furthermore, farmers can alter their cropping strategies. To investigate the soil moisture–rainfall feedback process, an extensive series of numerical experiments are performed with different initial soil moisture distributions using a regional climate model.

A study by Eltahir (1998) describes the physical mechanisms and processes resulting in the positive feedback between initial soil moisture and future rainfall. He suggests that anomalously high soil moisture conditions yield an increase in moist static energy per unit mass of boundary layer air and hence more rainfall in convective regimes. The mechanisms responsible for the feedback are directly linked to the surface energy budget and will be discussed in further detail in section 2. Zheng and Eltahir (1998) confirm these hypotheses using a two-dimensional model over West Africa. In this study, we further investigate the soil moisture–rainfall feedback mechanism by examining how late spring and summer soil moisture anomalies affect subsequent rainfall using a three-dimensional regional climate model over the midwestern United States. The hypotheses of Eltahir (1998) are tested using the Drought of 1988 and Flood of 1993 as examples.

The exact causes of extreme summertime flood and drought are relatively unknown. On one hand, Trenberth and Guillemot (1996) conclude that the cause of the drought of 1988 and flood of 1993 were related to La Niña and El Niño, respectively. They suggest that in 1993 the El Niño shifted the intertropical convergence zone (ITCZ) south of normal, which shifted the storm track southward and created a link to the moisture from the Gulf of Mexico. This link is what they believe to be responsible for the record rainfall observed over the Midwest in 1993. In contrast, for 1988 they suggest that the strong La Niña and the anomalously warm sea surface temperatures (SSTs) southeast of Hawaii caused a northward shift in the ITCZ, which, in turn, caused a northward shift in the storm track across North America. They argue that this shift did not allow the Gulf of

Mexico moisture link to form and resulted in drought conditions. On the other hand, Bell and Janowiak (1995) argued that anomalous SSTs in the tropical Pacific indirectly contributed to overall magnitude and extent of the 1993 flood. However, they state that no single factor alone caused the flooding. Furthermore, Namias (1991) argues that the La Niña observed in 1988 may have contributed to the drought, but was not the primary cause of the drought. Trenberth and Guillemot (1996) and Namias (1991) both agree that soil moisture may play an important role in increasing the persistence and magnitude of flood and drought. This study aims to isolate the effects of soil moisture on future rainfall by performing a series of numerical experiments where initial soil moisture conditions are varied and the large-scale forcing is kept the same.

Observations of soil moisture are extremely limited both on the spatial and temporal scale. As a result, conclusions regarding the initiation and persistence of flood and drought are difficult to make using observations alone. Recently, however, using observed soil moisture and rainfall data over the state of Illinois, Findell and Eltahir (1997) found that the feedback between soil moisture and future rainfall is a function of the time of year. More specifically, they found soil moisture and subsequent rainfall in the following three weeks to show a significant correlation in the summer and little or no correlation for the rest of the year. This summertime correlation is stronger than the serial correlation of rainfall, indicating that there is a positive feedback between soil moisture and rainfall over the state of Illinois during summer months. A study by Huang et al. (1996) somewhat contradicts the results of Findell and Eltahir (1997). They show using a soil moisture product simulated from U.S. meteorological station observations of precipitation and temperature that evaporation anomalies are smaller in magnitude than those for precipitation. In addition, they find that evaporation displays a strong correlation to soil moisture. Thus, they argue that anomalous evaporation resulting from anomalous soil moisture conditions has little impact on future precipitation. They further indicate that soil moisture is a better predictor of summer temperature than precipitation. One aspect of this study is to investigate how future precipitation responds to initial soil moisture using a numerical model. Another aspect of this study is to investigate how the timing of the soil moisture anomaly impacts the strength of flood and drought conditions.

Many numerical modeling studies, with a few exceptions, conclude that a positive feedback exists between soil moisture and rainfall [e.g., Atlas et al. (1993); Beljaars et. al. (1996); Mintz (1984); Oglesby (1991); Rind (1982); Rowntree and Bolton (1983); and Yeh et al. (1984)]. In general, these studies indicate that rainfall in the United States and other regions is sensitive to soil moisture conditions during months with pronounced convective activity, such as summers in midlatitudes. Most of these studies, however, prescribe unrealistic scenarios for soil moisture. In many instances, evapotranspiration is considered to be a surrogate for soil moisture and is maintained with no dynamic feedback at values close to zero for the dry simulations and close to the potential evaporation for the wet runs. In less extreme cases, soil moisture is prescribed uniformly both vertically and horizontally at the wilting point (~30% or less of soil saturation) for the dry runs and at the field capacity (~90% of soil saturation) for the wet runs where two-way interaction is allowed. Figure 3 presents the ISWS soil saturation profiles averaged over June, July, and August for Illinois. Although there are extreme interannual variations at or near the surface soil layer, the lower layers (below ~1 m) display considerably less variability. Thus, over Illinois and surrounding regions, it may be unreasonable to initialize soil moisture below the surface layer at values much less than 70%. Similarly, in arid regions, it is probably unreasonable to initialize soil moisture below the surface layer at values greater than 40% or so. Conceivably, more realistic anomalies in soil moisture have a relatively minor impact on rainfall. In this study, we investigate how the magnitude of the soil moisture anomaly affects the soil moisture–rainfall feedback. In doing so, we utilize a soil moisture dataset that combines information from the ISWS data and the pseudo-observed soil moisture dataset of Huang et al. (1996) (hereinafter referred to as HDG). In addition, this study differs from most of the above studies in that we focus on the mechanisms and pathways of the soil moisture–rainfall feedback.

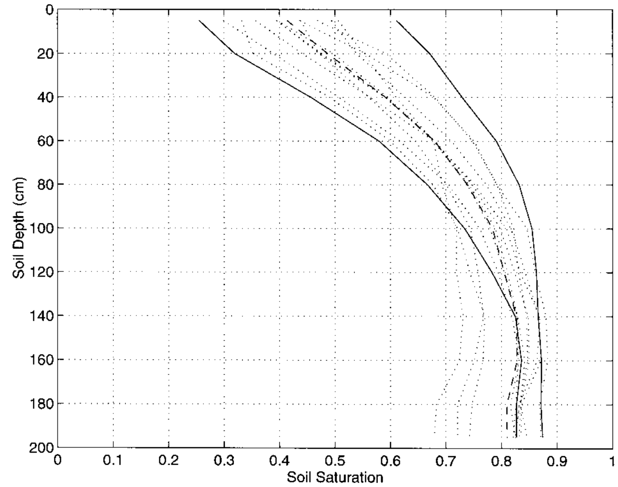

FIG. 3. Illinois State Water Survey soil saturation profile averaged over Jun, Jul, and Aug. The solid line on the left is the profile for 1988, the solid line on the right is for 1993, the dotted lines are the rest of the years, and the dashed line is the average of all of the years.

To summarize, this study addresses several questions: Are the persistent patterns in extreme hydrologic events maintained by external forcings or are they maintained by internal mechanisms involving the soil moisture–rainfall feedback? What role does the magnitude and direction of the prior soil moisture anomaly play in the soil moisture–rainfall feedback? How does the timing of this anomaly impact the strength of the feedback? And what are the pathways responsible for it? An extensive series of numerical experiments are performed to investigate these issues. Section 2 provides background on the theory of the soil moisture–rainfall feedback. Section 3 gives a detailed description of the numerical model used and the experiments performed in this study. The results and conclusions of the numerical experiments are described in section 4 and section 5, respectively.

2. Theory of the soil moisture–rainfall feedback mechanism

Soil moisture plays an important role in the climate system. In the scope of this study, it is defined as water that is available for evapotranspiration from bare soil and vegetated areas. It provides a long-term memory mechanism for the rainfall that occurs throughout the year. The soil moisture storage is depleted during warmer portions of the year when evapotranspiration tends to exceed rainfall (typically spring and summer) and replenished during the colder portions of the year (autumn and winter). The evapotranspiration from the land surface is used to moisten and cool the planetary boundary layer (PBL); the amount of moistening and cooling that occurs affects the energetics of the PBL.

Classically, researchers have focused on the soil moisture–rainfall feedback as a water recycling process, neglecting the radiative processes. Recently, however, Betts and Ball (1994), Entekhabi et al. (1996), Eltahir (1998), and Schär et al. (1999) have suggested that soil moisture not only impacts the water budget of the surface and the PBL, but it also impacts the energy budget. This section describes the mechanisms and pathways through which soil moisture impacts the near-surface variables that affect boundary layer processes and rainfall based on Eltahir (1998). Figure 4 provides a schematic diagram of these processes.

a. Radiative feedbacks

Anomalously wet soils are often associated with greener denser vegetation and darker soils. Both greener denser vegetation and darker soils yield a lower surface albedo. A lower surface albedo implies that more of the incoming solar radiation is absorbed at the surface. In other words, neglecting any cloudiness feedback, more incoming solar radiation (on average) is likely to be absorbed by the surface during periods of anomalously wet soils than during periods of anomalously dry soils. It is important to note that any feedback that results in an increase in cloudiness with increasing soil moisture tends to balance or in some cases outweigh the surface albedo effect, as will be discussed later.

Anomalously high soil moisture also tends to lower

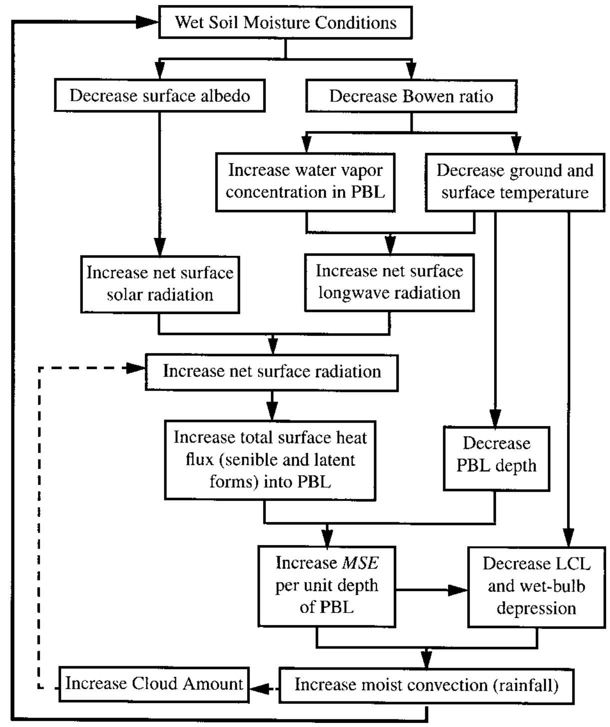

FIG. 4. Diagram relating the pathways through which anomalously wet soil moisture conditions lead to subsequent rainfall.

the Bowen ratio (ratio of sensible heat flux to latent heat flux) by increasing the surface latent heat flux and decreasing the surface sensible heat flux. The increase in latent heat flux moistens the lower atmosphere and hence, increases downward longwave radiation due to the greenhouse effect of atmospheric water vapor. The same process cools the surface and hence, reduces the outgoing surface longwave radiation due to the Stefan–Boltzmann Law. Overall, the decrease in the Bowen ratio with increasing soil moisture results in an increase in net longwave radiation at the surface.

As mentioned above, the cloud feedback may play an important role in the response of the surface radiation budget to changes in soil moisture. If more clouds result due to an increase in soil moisture, more of the outgoing longwave radiation is re-emitted toward the surface. This further increases the net surface longwave radiation and tends to balance a significant portion of the reduction in incoming surface solar radiation resulting from the increase in clouds. The strength of the cloud feedback is an important factor in determining the strength of the response of net surface radiation to changes in soil moisture.

In summary, anomalously wet soils tend to increase both the net surface solar radiation (via the albedo feedback) and the net surface longwave radiation (via the Bowen ratio feedback). This yields an increase in net all-wave radiation. The sensitivity of clouds to changes in soil moisture may alter this response.

b. Boundary layer and moist static energy feedbacks

The above subsection suggests that anomalously wet soils are associated with an increase in net surface radiation. Net surface radiation is balanced by the sum of the latent, sensible, and soil heat fluxes. On the long timescale, the soil heat flux can be considered negligible. Thus, the net radiation (or total surface heat flux) is equivalent to total fluxes of latent and sensible heat.

The total energy of the PBL can be described by the moist static energy (MSE). (MSE $= gZ + C_pT + LQ$, where g is the acceleration due to gravity, Z is elevation, C_p is the specific heat capacity at constant pressure, T is temperature, L is the latent heat of vaporization, and Q is the water vapor mixing ratio.) At large spatial scales (i.e., neglecting advection), the MSE of the PBL is supplied by the total flux of heat from the surface and depleted by the entrainment of low MSE air existing above the PBL, radiative cooling, and negative heat fluxes associated with convective downdrafts. Thus, with all else being equal, anomalously wet soil moisture conditions tend to increase the MSE of the PBL via an increase in total heat flux supplied from the surface.

Soil moisture also has a pronounced impact on the depth of the PBL. As implied above, drier soils are associated with higher fluxes of sensible heat. A high sensible heat flux is associated with greater turbulent energy. This tends to increase turbulent mixing, which increases the PBL growth rate and hence the depth of the PBL. Thus, even neglecting the increase in MSE supplied by the surface, anomalously wet soil moisture conditions tend to result in an increase in the MSE per unit depth (mass) of PBL.

MSE within the PBL tends to be well mixed (height invariant). However, above the PBL in the troposphere, the vertical distribution of MSE tends to decrease with elevation. As the PBL grows, low MSE air from above the PBL is entrained. This tends to lower the overall MSE of the PBL (Betts and Ball 1994). In addition, as the PBL grows in height, it entrains air of increasingly lower MSE from above (since the MSE above the PBL decreases with elevation). This results in an enhanced reduction of the MSE of the PBL. It is suggested in the above paragraph that anomalously wet soils result in a decrease in the growth rate of the PBL. This tends to reduce the amount of low MSE air entrained from above PBL and hence increase the overall MSE of the PBL.

In summary, changes in soil moisture conditions contribute to the overall MSE of the PBL in three ways: 1) surface fluxes from below the PBL, 2) PBL depth, and 3) entrainment from above the PBL. Anomalously wet soils tend to increase the flux of high MSE air into the PBL from below, reduce the PBL height increasing the MSE per unit mass of air, and reduce the amount of entrained air of low MSE from above the PBL. Each of these effects are additive and contribute to a relative increase of MSE per unit mass of air in the PBL.

c. Moist static energy and moist convection

As stated above, soil moisture plays an important role in determining the total MSE in the PBL. The impacts of soil moisture above the PBL are relatively minor. This implies that the changes in soil moisture alter the stability profile of the atmosphere. Since an increase in soil moisture tends to increase the overall MSE of the PBL and has little impact on the MSE above the PBL, the MSE vertical profile becomes unstable. As a result, moist convection occurs to redistribute the MSE toward neutral conditions. Furthermore, the increase in PBL MSE should result in an increase in the convective available potential energy (CAPE) (Williams and Renno 1993). CAPE can be directly related to storm size. Thus, anomalously wet soil moisture conditions tend to increase the frequency (via an increase in instability) and magnitude (via an increase in CAPE) of convective rainfall events. A recent study by Eltahir and Pal (1996) confirms these ideas using data from the Amazon Forest. They indicate that an increase in wet-bulb temperature (a quantity proportional to MSE) increases the convective instability and hence, the frequency and magnitude of rainfall events in the Tropics and the summer hemisphere.

Soil moisture is also likely to have an impact on the cloud base. Since soil moisture exhibits a positive relationship with low-level water vapor and a negative relationship with low-level temperature, the wet-bulb depression (dry bulb temperature minus wet-bulb temperature) should display a higher sensitivity to soil moisture than water vapor and temperature taken separately. Wet-bulb depression is related to the lifting condensation level, which is an approximate measure of the cloud base. An increase in soil moisture tends to result in a decrease in the height of the cloud base via a decrease in wet-bulb depression. On average, this should increase the likelihood of occurrence of convective rainfall. In addition, a lower cloud base should result in a deeper cloud (assuming the cloud top height remains constant). This should result in an increase in the magnitude of convective rainfall events.

In summary, two additive factors contribute to soil moisture's impact on the frequency and magnitude of convective rainfall events: 1) MSE of the PBL and 2) cloud base height. Soil moisture exhibits a positive relationship with PBL MSE and a negative relationship with the cloud base height. The response of both these effects on the occurrence and size of convective rainfall events operates in the same direction. If the arguments presented in this section hold true, anomalously wet soil moisture conditions should lead to an increase in the frequency and volume of convective rainfall. This increase in rainfall wets the soil and therefore, results in a positive feedback between soil moisture and rainfall.

3. Description of numerical model and experiments

The purpose of this study is to investigate the pathways and mechanisms through which initial soil moisture conditions impact subsequent rainfall. A series of numerical experiments using a regional climate model are performed to investigate the impact of soil moisture on the energy and water balances of the PBL over the Midwest using the drought of 1988 and flood of 1993 as representative events. This section provides a description of these experiments, as well as, a description of the numerical model and data used for this study.

a. Numerical model description

In this study, we use a modified version of the National Center for Atmospheric Research's (NCAR) regional climate model (RegCM). The original model was developed by Dickinson et al. (1989), Giorgi and Bates (1989), and Giorgi (1990) using the Pennsylvania State University–NCAR Mesoscale Model, version 4, [MM4; (Anthes et al. 1987)] as the base dynamical framework. Here we provide only a brief description of RegCM; a more detailed description of RegCM can be found in Giorgi and Mearns (1999) and its associated references, and additional model modifications can be found in Pal et al. (2000). Like MM4, RegCM is a primitive equation, hydrostatic, compressible, σ-vertical coordinate model, where $\sigma = (p - p_{top})/(p_s - p_{top})$, p is pressure, p_{top} is the pressure specified at the top of the model, and p_s is the prognostic surface pressure.

Unlike MM4, RegCM is adept for climate studies. The atmospheric radiative transfer computations are performed using the CCM3-based package (Kiehl et al. 1996) and the planetary boundary layer computations are performed using the nonlocal formulation of Holtslag et al. (1990). The resolvable (large-scale) cloud and precipitation processes are computed using a modified/simplified version of the Hsie et al. (1984) scheme (Pal et al. 1999). The unresolvable precipitation processes (cumulus convection) are represented using a modified Kuo-type parameterization [Anthes (1977) and Anthes et al. (1987)]. Last, the surface physics calculations are performed using a soil–vegetation hydrological process model [Biosphere–Atmosphere Transfer Scheme (BATS), Dickinson et al. (1986)]. Briefly, BATS allows for the exchange of heat, moisture, and momentum between the atmosphere and land surface. It is composed of a vegetation layer, a snow layer, and three soil layers varying from 10 cm in depth to 3 m. (Soil moisture is interactive and allowed to vary in space and time.)

RegCM requires initial conditions and time-dependent lateral boundary conditions for the wind components, temperature, surface pressure, and water vapor. The initial conditions (except for soil moisture and SST) and lateral boundary conditions for each simulation are taken from the NCEP reanalysis data (Kalnay et al. 1996). These data have a spatial resolution of 2.5° × 2.5°, are distributed at 17 pressure levels (8 for humidity), and are available at time intervals of 6 h. The SST is prescribed using data provided by the U.K. Meteorological Office [Rayner et al. (1996); 1° grid]. Last, soil moisture for the control simulations is initialized using the merged HDG/ISWS dataset as described below.

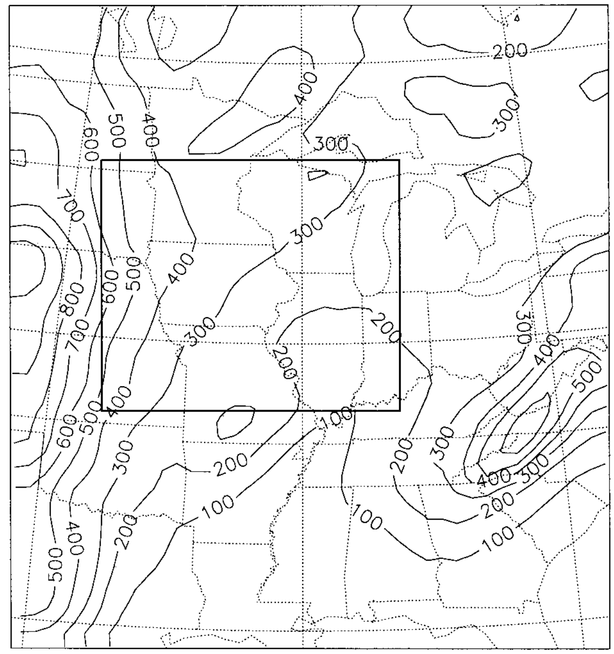

FIG. 5. Map of the domain and terrain heights (m) used for the numerical simulations. The outlined box is the region over which the averages are taken and corresponds to the areas where the drought of 1988 and flood of 1993 were most severe. The contour interval is 100 m.

b. Model domain

The model domain is centered around the state of Illinois at 40.5°N and 90°W and is projected on a Lambert conformal grid. The domain size is 2050 km × 2500 km with horizontal grid point spacing of 50 km. At this resolution, the main topographic features of the domain are captured. The model domain and topography are shown in Fig. 5. This region captures the primary area of drought of 1988 and flood of 1993 (see Fig. 1). The western boundary lies along the eastern slopes of the Rocky Mountains. The boundaries were not moved farther west to avoid the impacts of complex topography and lack of adequate data in mountainous regions to force the model. As a result, the western boundary lies close to the drought and flood regions.

The vegetation is characterized using the Global Land Cover Characterization (GLCC) data (Loveland et al. 1999). The predominant vegetation type of the domain is crops that are characterized by a shallow root zone depth (1.0 m). Hence, the majority of the water for evapotranspiration is extracted from the upper portions of the soil, which is consistent with Fig. 3 in that there is little variability in the soil column below 1 m. During the peak growing season (summer), crops achieve a relatively high maximum fractional vegetation coverage (85%) and leaf area index (6) and a relatively low minimum stomatal resistance (40 s m^{-1}). All of these parameters suggest that crops transpire more than the remaining vegetation types. The other dominant vegetation types within the model domain are short and tall grass, disturbed forest, evergreen needleleaf trees and deciduous broadleaf trees, and mixed woodland. The soil texture class is prescribed according to the vegetation characterization. The primary soil texture over the domain is comparable to a loam soil. The values for porosity, wilting point, and saturated hydraulic conductivity associated with this texture class are 48%, 33%, and 6.3×10^{-3} mm s^{-1}, respectively. Although the soil properties vary in the horizontal, they do not vary in the vertical.

c. *Soil moisture datasets*

The initialization of soil moisture in models is important for an accurate representation of the land surface energy balance. To initialize soil moisture in the control simulations, this study takes advantage of a merged dataset created by the authors that combines three datasets: ISWS; HDG; and a vegetation-based climatology (denoted hereinafter as HDG/ISWS).

The ISWS is responsible for a network of direct soil moisture measurement stations across the state of Illinois [see Hollinger and Isard (1994)]. Since 1981, biweekly measurements have been taken at 11 depths from the surface to 2 m over 17 grass-covered sites across Illinois. To date, it provides the most accurate and the most spatially and temporally extensive soil moisture dataset of its kind in North America. Although this is the best dataset available in North America, it does not cover a large enough spatial range to initialize the entire domain of the experiments described above.

The HDG soil moisture data used in this study come from a simulated dataset created over the continental United States generated from the 344 climate divisions data (Huang et al. 1996). It spans the period from 1931 through 1993. The model used to generate these data is based on the water budget of the soil and uses monthly station rainfall and temperature as inputs. Its four parameters are calibrated using observed rainfall, temperature, and runoff from an area in Oklahoma. The soil moisture model performs remarkably well when compared with the ISWS soil moisture data (correlation coefficient = 0.84). Therefore, this dataset should be adequate in initializing the domain of the regional climate model.

BATS requires soil moisture data at three soil levels, however, the HDG data are given as one volumetric value. A relation between the ISWS and HDG data is developed to generate a more realistic initial vertical soil moisture profile over the entire domain. The soil saturation at a given layer k and location i, j ($S_{i,j,k}$) is the ratio of the average ISWS value at the given layer ($\mathrm{ISWS}_k^{\mathrm{avg}}$) to the HDG value over Illinois ($\mathrm{HDG}_{\mathrm{ILL}}^{\mathrm{avg}}$) multiplied by the HDG value ($\mathrm{HDG}_{i,j}$) according to the following relation:

$$S_{i,j,k} = \frac{\mathrm{ISWS}_k^{\mathrm{avg}}}{\mathrm{HDG}_{\mathrm{ILL}}^{\mathrm{avg}}}\mathrm{HDG}_{i,j}. \quad (1)$$

This approximation translates the vertical profile structure from the Illinois data to the rest of the domain. Although, vertical soil water profile structures are regionally dependent, this method provides a more reasonable way to initialize soil moisture that accounts for the vertical variability.

The land portions of the domain that do not include the United States (i.e., Canada and Mexico) are initialized with a soil moisture climatology based on the vegetation type. Moister values are associated with trees, and drier values are associated with deserts and grasslands according to Giorgi and Bates (1989).

TABLE 1. Description of each simulation performed in this study.

Simulation months:
• May, Jun, Jul, Aug, and Sep 1988
• May, Jun, Jul, Aug, and Sep 1993
Simulation duration: 1 month
Soil saturation (initialized on the 1st of each month):
• Observed HDG/ISWS.
• 10% uniformly over entire domain and depth
• 25% uniformly over entire domain and depth
• 50% uniformly over entire domain and depth
• 75% uniformly over entire domain and depth
• 90% uniformly over entire domain and depth

d. *Design of numerical experiments*

To develop an understanding of the physical mechanisms involved in the soil moisture–rainfall feedback, a series of numerical experiments for several different months are performed each of which differ in the initial soil saturation. Month-long simulations are performed for May, June, July, August, and September of 1988 and 1993. The control runs for each month and year are initialized using the merged HDG/ISWS dataset described in the previous subsection. In addition, a set of integrations are performed with the soil moisture initialized uniformly in the vertical and horizontal at values of 10%, 25%, 50%, 75%, and 90% of saturation. In total, there are six different soil moisture initializations for each month and year. Table 1 provides a description of the experiments performed in this study.

4. Results

This section describes the results of the simulations performed in this study. In the first subsection, we brief-

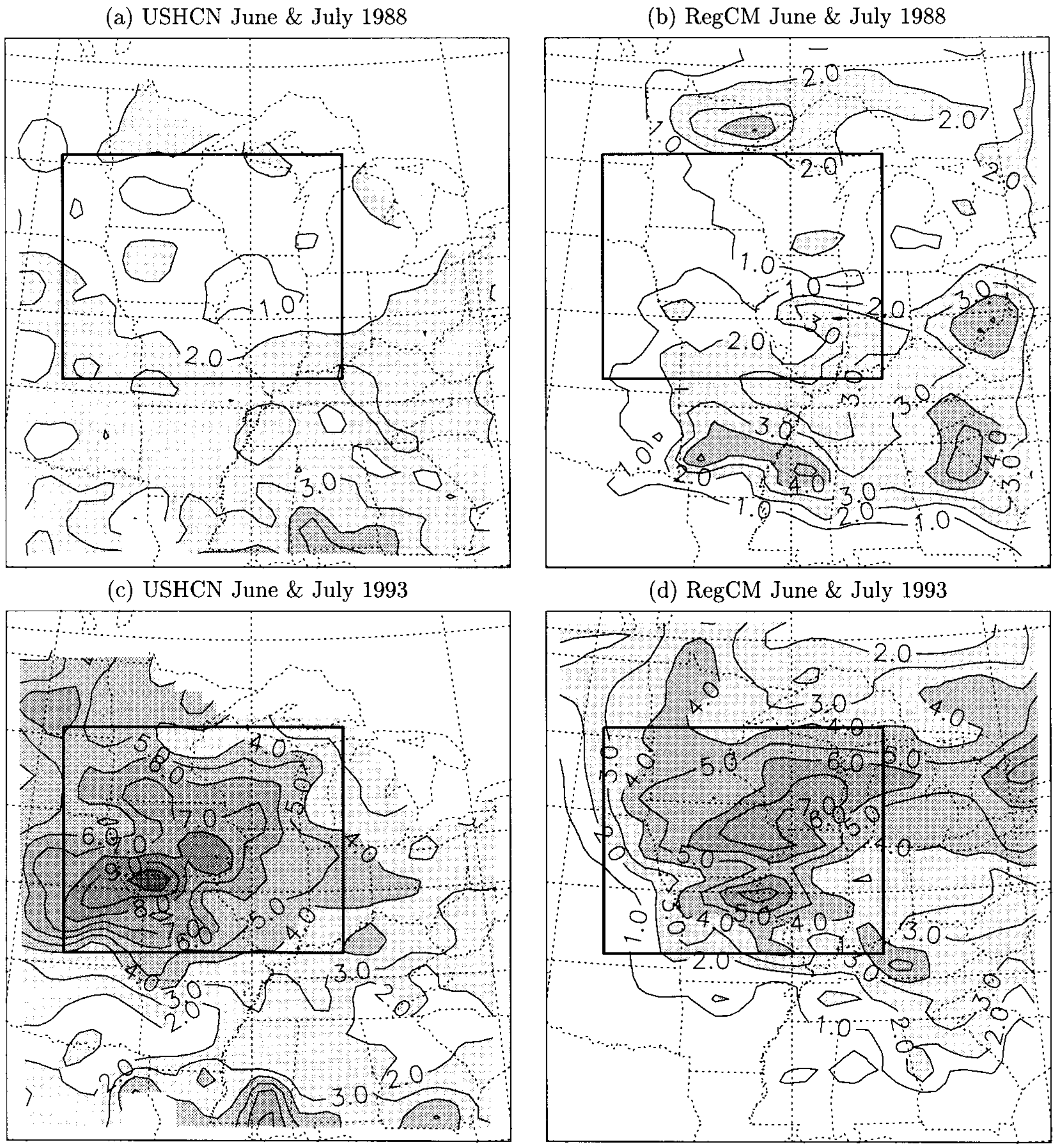

FIG. 6. USHCN observed and RegCM simulated rainfall (mm day^{-1}) averaged over Jun and Jul of 1988 and 1993: (a) USHCN Jun and Jul 1988; (b) RegCM Jun and Jul 1988; (c) USHCN Jun and Jul 1993; (d) RegCM Jun and Jul 1993. The outlined box is the region over which the averages are taken and corresponds to the areas where the drought of 1988 and flood of 1993 were most severe. The contour interval is 1 mm day^{-1}, and the shading interval is 2 mm day^{-1}.

ly compare the modeled rainfall to the observations, and in the second subsection, we investigate the sensitivity of the model to initial soil moisture.

a. Brief model comparison to precipitation observations

Because this is a mechanistic (not predictive) study, we do not perform a rigorous comparison of the model simulations to the observations. More rigorous model comparisons can be found in Giorgi and Mearns (1999) and Pal et al. (2000), among others. This section compares the control simulations to the United States Historical Climatology Network (USHCN) data (Karl et al. 1990). This dataset consists of 1221 high quality stations from the United States Cooperative Observing Network within the 48 contiguous United States. The data are interpolated onto the RegCM grid defined in section 3 by exponentially weighting the station data according to the distance of the station from the center of the RegCM grid cell, with a length-scale of 50 km.

Figure 6 compares the observed gridded USHCN rainfall averaged over June and July for both 1988 and 1993 with the model-simulated rainfall for the same months. June and July of 1988 and 1993 are selected for comparison because they represent the most severe months of the drought and flood. When making the comparisons, it should be noted that USHCN observations do not exist over Canada.

For June and July of 1988, the model performs well in capturing the lack of observed rainfall in the northern portion of the domain. In addition, the model captures the general region of rainfall greater than 2 mm day^{-1} in the southern portion of the domain. There are, however, small isolated regions of rainfall in excess of the observations. Furthermore, there is a considerable underestimation of precipitation along the southern and southwestern boundaries. It is believed that this is a result of the proximity of the boundaries to these regions. In addition, it is likely that the suppression of rainfall at the southern boundary may have resulted in the unobserved isolated rainfall peaks that formed downwind.

For June and July of 1993, the model performs well in reproducing the anomalously high rainfall in the upper Midwest. However, the model underestimates the overall magnitude of the flood peak by approximately 2 mm day^{-1}. In addition, the location of the peak is simulated too far to the north and east of the observed peak. This shift may be a consequence of the proximity of the flood peak to the western boundary. Similar to the 1988 simulations, the model does not perform well in reproducing the observed rainfall distribution along the southern boundary of the domain. Again, this may be due to the location of the boundaries.

In summary, the model performs well in reproducing the drought and flood events over the upper Midwest observed in 1988 and 1993, respectively. Some deficiencies do exist in simulating the precise location and overall magnitude of the flood peak in 1993. Although only June and July of 1988 and 1993 are compared to observations, the model exhibits a similar performance in the other months simulated. Of significant note, the model can clearly distinguish between the two extreme years. This is important because it indicates that the model is able to capture the interannual variability. With this in mind, it is reasonable to proceed with the sensitivity experiments.

On a final note, Seth and Giorgi (1998) found that the strength of the soil moisture–rainfall feedback can be dependent on the location of the domain boundaries. The point of this study is to develop an understanding of the key processes and mechanisms in the soil moisture–rainfall feedback. By using the small domain, in addition to computation savings, we are able to constrain the spatial location of the drought and flood in their respective years. This allows us to better identify the key mechanisms and processes the responsible for the soil moisture–rainfall feedback.

b. Model response to initial soil moisture

In this section, we investigate the model's sensitivity to initial soil moisture. Recall from section 3d that for each month and year, there are six different initial soil moisture conditions. The control simulation is initialized using the HDG/ISWS soil moisture dataset described in section 3c for 1988 and 1993 and the remaining five simulations are initialized uniformly in the horizontal and vertical at soil saturations of 10%, 25%, 50%, 75%, and 90%.

The comparisons are made using monthly and spatial averages of a set of hydrologic fields against initial soil moisture. The spatial average is made over the region severely affected by the flood and drought outlined in Figs. 5 and 6. This region is centered around Iowa and extends to include most of Illinois and parts of Indiana, Wisconsin, Minnesota, South Dakota, Nebraska, Kansas, and Missouri. It is selected because it was most affected by the drought of 1988 and the flood of 1993.

For brevity, averages for the May–September simulations for 1988 and 1993 are presented. The results for the individual months (except July 1988) are qualitatively the same. In these plots, the character F denotes 1993 and D denotes 1988; the large bold F and D denote the control simulation for their respective years.

1) SENSITIVITY OF FUTURE RAINFALL TO INITIAL SOIL MOISTURE

Figure 7 depicts a summary of the total, convective, and nonconvective rainfall for the July 1988 and 1993 simulations as a function of initial soil moisture. On the whole, a considerable sensitivity of total rainfall to changes in initial soil moisture is displayed in both years; there is a 50%–60% increase when increasing soil saturation from the low to high extremes (10%–90% of saturation; Fig. 7a). This sensitivity is primarily a result of an increase in convective rainfall where there is nearly a 250% increase between the extremes in soil moisture (Fig. 7b); nonconvective rainfall tends to remain virtually constant with soil moisture for both years (Fig. 7c). It should be noted that Pan et al. (1996) and Pal and Eltahir (1997) suggest that the choice of cumulus convective parameterization can have an impact on the strength of the soil moisture–rainfall feedback.

The response of future rainfall to initial soil moisture is nonlinear. Figure 8 is a plot of the relative sensitivity of total rainfall to initial soil saturation expressed as the percent change in rainfall per percent change in soil moisture. Each individual bar is computed from two simulations. For example, the July of 1988 bar in the 25%–50% soil saturation range is computed using the 25% and 50% soil saturation simulations. In nearly every case, rainfall in the 25%–50% soil moisture range is most sensitive. The 50%–75% range also exhibits considerable sensitivity (except July 1988), however, it is significantly weaker than the 25%–50% range. The weakest sensitivities are displayed in the highest and lowest soil saturation ranges. Note that the relative sensitivity of convective rainfall to initial soil saturation (not shown) is considerably larger than that of total

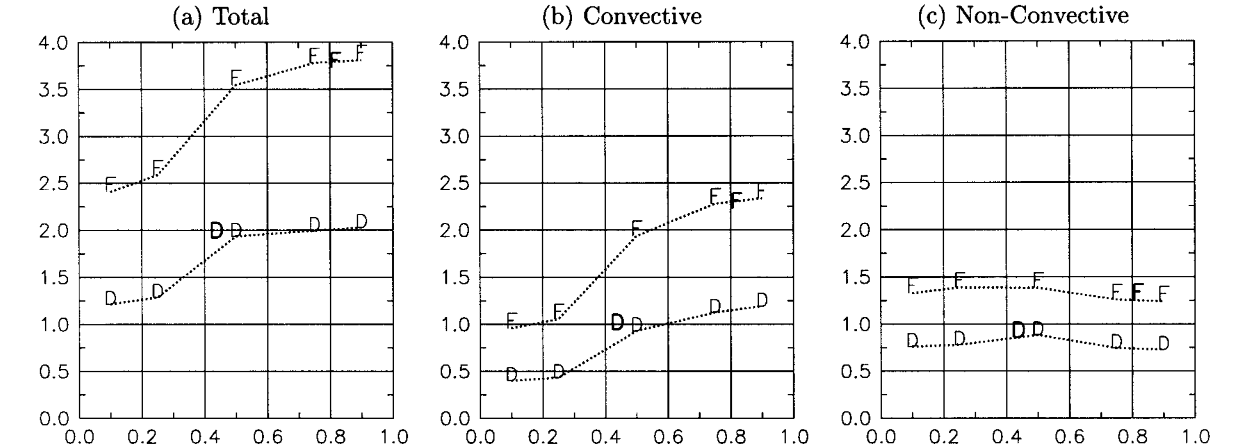

FIG. 7. Simulated monthly rainfall (mm day^{-1}) for the 1988 and 1993 simulations as a function of initial soil saturation: (a) total rainfall, (b) convective rainfall, and (c) nonconvective rainfall. Each data point represents the average of the May, Jun, Jul, Aug, and Sep simulations over the Midwest region outlined in Fig. 5 given the initial soil moisture. The D denotes the drought year (1988) and the F denotes flood year (1993). The boldface D and F denote the control simulations for each year.

rainfall since nonconvective rainfall is unresponsive to changes in soil moisture (except for July 1988 in the 50%–75% soil saturation range) (see Fig. 7c). The July 1988 simulations in the 50%–75% soil saturation range behave differently than the rest of the simulations in that a significant negative feedback to soil saturation is displayed. In this particular case, this can be explained by the fact that nonconvective rainfall significantly decreases when soil saturation is increased from 50% to 75% and convective rainfall remains virtually constant; thus the significant negative sensitivity displayed in Fig. 8a in this range.

The low sensitivity of future rainfall to initial soil saturation in the 10%–25% soil saturation range is due to transpiration ceasing when the soil saturation falls below its wilting point (approximately 33% of saturation in these simulations). In this regime, the response of rainfall to changes in soil moisture is largely due to changes in bare soil evaporation, some interception loss, and some transpiration from the upper soil layer when the wilting point is exceeded shortly after rainfall events (not shown). Little sensitivity of rainfall to soil moisture is also exhibited at the higher values of soil moisture (>75%). This is a result of the evapotranspiration reaching the potential evaporation rate when the soils are unlimiting (atmosphere limiting).

Summertime soil saturation values over Illinois typically range from 40% near the surface to 80% at depth (see Figs. 2 and 3). With this in mind, it would be expected that the depth-averaged (surface to 1 m) soil saturation values over the upper Midwest would lie around 60% during normal years. (One might expect this value to be slightly lower since more of the water for evapotranspiration is extracted near the surface.) The

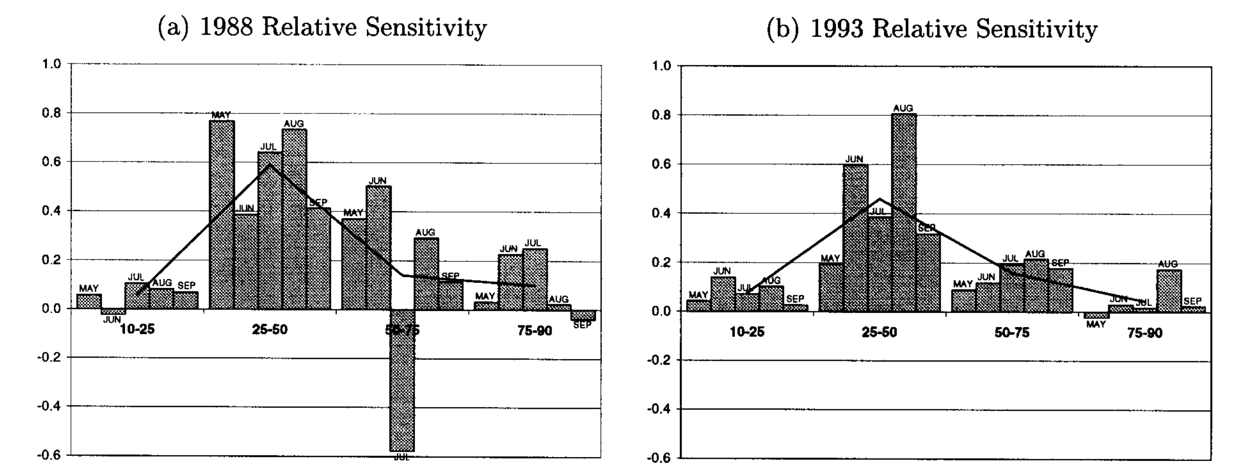

FIG. 8. Plot of relative sensitivity vs soil saturation: (a) 1988 and (b) 1993. Each bin on the x axis represents a range of soil saturations; the end points of each bin are individual simulations where each bar within the bin represents a month. The values on the y axis represent the percent change in rainfall per percent change in soil moisture. The solid line represents the average of all the months for the given year.

TABLE 2. Average number of hourly rainfall events within specified intervals occurring between 1 May and 30 Sep of 1988 and 1993 for the 10%, 50%, and 90% soil saturation simulations. The number in parentheses indicates the percentage of total rainfall. The last row provides the totals over all the intervals with the number in parentheses indicating the total rainfall (mm). The values are computed over the upper Midwest region defined in Fig. 5.

Interval	1988			1993		
(mm day^{-1})	10%	50%	90%	10%	50%	90%
0.1–0.5	1185 (4.1)	1192 (2.7)	1127 (2.7)	1003 (1.9)	787 (1.3)	678 (1.1)
0.5–1.0	384 (5.8)	385 (3.7)	496 (4.5)	407 (3.3)	337 (1.9)	371 (2.0)
1.0–2.0	446 (13.6)	482 (9.3)	498 (9.0)	446 (7.0)	455 (5.3)	481 (5.2)
2.0–4.0	428 (24.8)	501 (19.4)	585 (21.7)	552 (17.4)	652 (14.6)	684 (14.3)
4.0–8.0	230 (25.9)	456 (34.1)	458 (32.7)	564 (35.2)	849 (37.7)	801 (33.9)
8.0–16.0	98 (21.4)	170 (23.5)	164 (21.6)	282 (32.2)	435 (34.8)	516 (38.4)
≥16.0	10 (4.4)	28 (7.3)	31 (7.8)	16 (3.0)	32 (4.4)	37 (5.1)
Totals:	2781 (199)	3214 (313)	3359 (332)	3270 (382)	3547 (539)	3568 (577)

distinct asymmetry displayed in Fig. 8 suggests that a perturbation (say 20%) to this value would lead to different responses depending on the direction of the perturbation. On one hand, a perturbation to drier conditions would place the soil saturation in the most sensitive (soil controlled) regime displayed in Fig. 8. This would likely cause the rainfall to display significant response to the drier surface conditions. On the other hand, a perturbation to wetter conditions would result in a significant, but less dramatic response. A similar assymetric response is also exhibited in Brubaker and Entekhabi (1996) using a conceptual land–atmosphere model. Overall, the results presented here and in Brubaker and Entekhabi (1996) indicate that rainfall over the Midwest is more responsive to negative soil moisture anomalies than positive. This also suggests in the context of these experiments that the soil moisture–rainfall feedback would tend to favor more persistent drought conditions in comparison to flood conditions.

Table 2 presents the number of rainfall events occurring between 1 May and 1 October of both 1988 and 1993 within the specified interval for the 10%, 50%, and 90% soil saturation simulations over the upper Midwest region shown in Figs. 5 and 6. The numbers in parentheses indicate the percentage of the total rainfall occurring within the indicated interval (except the bottom row). The bottom row provides the totals over all the intervals with the number in parentheses indicating the total volume of rainfall (mm).

Moving from dry to wet conditions results in both an increase in the number of rainfall events and the total volume of rainfall (bottom row of Table 2); in 1988, there is a 21% increase in the number of events and a 67% increase in the volume, and in 1993 the increases are 9% and 51%, respectively. The majority of the increased frequency and intensity occurs when increasing soil moisture from 10% to 50% of saturation. This is consistent with the above findings, which favor a stronger soil moisture–rainfall feedback under dry conditions as opposed to wet.

In 1988, other than a small decrease in the 0.1–0.5 mm day^{-1} range, an increase in the number of rainfall events with increasing soil moisture occurs in all of the intervals. Furthermore, there is tendency for the percentage of total rainfall to decrease in the lower intervals ($<$4 mm day^{-1}) and to increase in the moderate and high intervals ($>$4 mm day^{-1}). Hence, the majority of the increase in volume can be explained by an increase in the number of large events where most of the total rainfall occurs. The 1993 simulations display a somewhat different pattern than 1988 in that there is a significant decrease in the number of small rainfall events ($<$1 mm day^{-1}). However, similar to 1988, there is a tendency for an increase in the number of rainfall events in the remaining intervals. Also somewhat similar to 1988, the percentage of rainfall occurring in the low to low-moderate rainfall intervals ($<$4 mm day^{-1}) tends to decrease with increasing soil saturation and the percentage tends to increase in the higher intervals ($>$8 mm day^{-1}) with little change in the moderate range. This shift in distribution of rainfall to larger events from increasing soil moisture results in the significant increase in overall rainfall.

Overall, increasing soil moisture over the upper Midwest tends to result in a moderate increase in the number of rainfall events and a substantial increase in the magnitude of the rainfall events. In both 1988 and 1993, there tends to be a decrease or little change in the number of small rainfall events and an increase in the number of moderate to large events when increasing soil saturation. The overall volume of rainfall increases because of a shift in the frequency distribution toward larger events much like what is shown in Eltahir and Pal (1996). There they suggest that an increase (decrease) in wet bulb temperature over the Amazon results an increase (decrease) in the frequency and magnitude of rainfall.

2) MODEL RESPONSE TO THE TIMING OF THE SOIL MOISTURE ANOMALY

As mentioned in section 1, Findell and Eltahir (1997) found, using the ISWS soil moisture data, that the feedback between soil moisture and subsequent rainfall is strongest in June and July and weakest in winter months. Hence, we should expect the June and July simulations

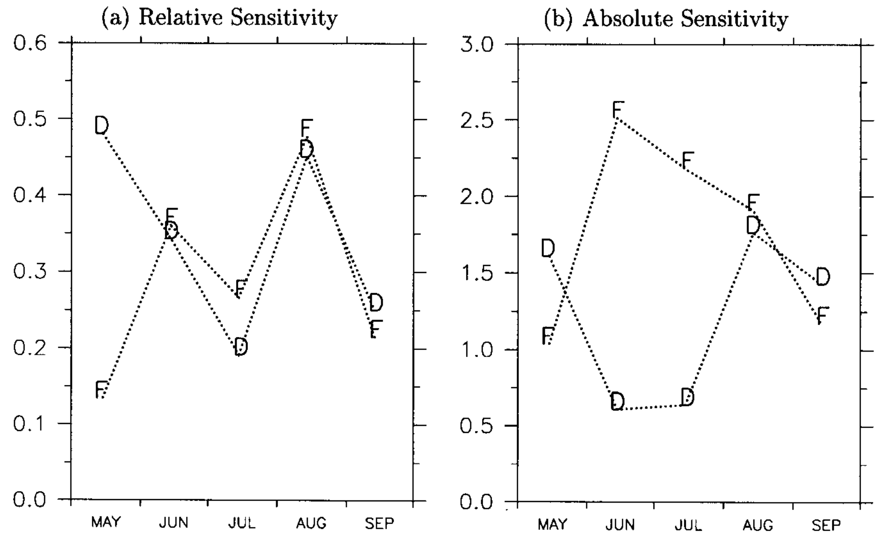

FIG. 9. Relative and absolute sensitivities of rainfall to initial soil moisture. Relative sensitivity is a measure of the relative impact of initial soil moisture on future precipitation expressed as a percentage. Absolute sensitivity is defined as the slope of the best fit line of future precipitation to initial soil moisture for each simulation month. The D denotes the drought year (1988) and the F denotes flood year (1993).

to show a greater sensitivity to soil moisture than the May, August, and September simulations. Figure 9 displays the relative and absolute sensitivities of future rainfall to initial soil saturation as a function of month. The relative sensitivity is a measure of the relative impact of initial soil moisture on future rainfall. It expresses the percent change in rainfall per percent change in soil saturation. The absolute sensitivity is defined as the slope of the best-fit line to the soil moisture–rainfall data shown in Fig. 7a (not including the control integration). It is a measure of the change in rainfall between bone dry intial conditions and fully saturated.

The relative sensitivity of future rainfall to initial soil moisture (Fig. 9a) varies from approximately 0.1% to 0.5% change in rainfall per percent change in soil saturation. It does not show a clear pattern with the time of year. This suggests that the timing of the soil moisture anomaly had little impact on the relative change of rainfall during May, June, July, August, and September during 1988 and 1993.

The absolute sensitivities (Fig. 9b) lie between 0.6 and 2.5 mm day^{-1} per unit change in soil saturation. On average, the absolute sensitivity is 1.2 mm day^{-1} for 1988 and 1.8 mm day^{-1} for 1993. The difference in rainfall between the control simulations for 1988 and 1993 is approximately 1.8 mm day^{-1} (see Fig. 7). Although this difference is similar to the absoulute sensitivity, it is unrealistic for soil moisture in the Midwest to vary from bone dry conditions to fully saturated. Thus, in the context of these simulations, soil moisture alone could not have caused the extremes observed in 1988 and 1993. However, Fig. 8 indicates that the majority of the soil moisture–rainfall sensitivity is exhibited in the 25%–75% soil saturation regime. Soil moisture over the Midwest typically falls within these bounds. This suggests that soil moisture played an important role in the persistence and maintainence of the extreme events.

Consistent with the findings of Findell and Eltahir (1997), the 1993 simulations exhibit the greatest soil moisture–rainfall sensitivity in absolute terms during June and July; the 1988 simulations do not exhibit this consistency. In terms of relative sensitivity, neither year displays an obvious pattern with the time of year. Note that the the summers of 1988 and 1993 are extreme years. The large-scale conditions may have had an enhanced importance in these events. Additional simulations of years with more normal rainfall conditions are likely to be required to gain a better sense of how the timing of the soil moisture anomaly affects future rainfall.

3) SOIL MOISTURE–RAINFALL PATHWAYS

This section focuses on the pathways through which soil moisture affects future rainfall described in section 2 and Fig. 4. Figures 7, 10, 11, and 12 summarize these results. Again, for brevity, only the averages of the May–September of 1988 and 1993 simulations will be investigated.

It is suggested in section 2 that soil moisture is likely to have a pronounced impact on rainfall in convective regimes as opposed to nonconvective regimes. Figure 7 indicates that convective rainfall for the control simulations tends to be greater than nonconvective rainfall both in July of 1988 and 1993 (bold D and F, respectively). This implies that it is useful to investigate the processes described in section 2.

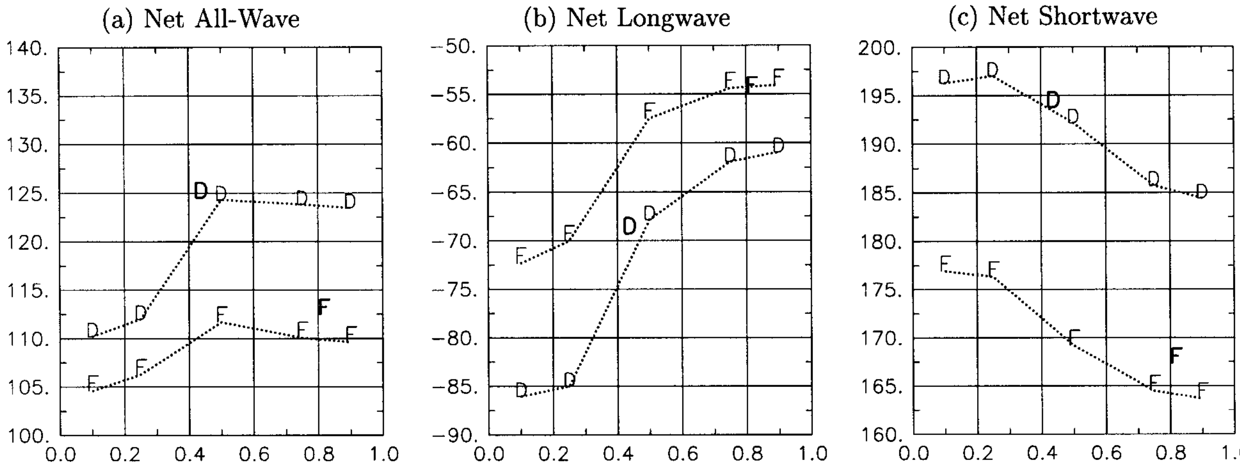

FIG. 10. Simulated monthly surface radiation fields (W m^{-2}) for the 1988 and 1993 simulations as a function of initial soil saturation: (a) net radiation, (b) net longwave radiation, and (c) net solar radiation. Each data point represents the average of the May, Jun, Jul, Aug, and Sep simulations over the Midwest region outlined in Fig. 5 given the initial soil moisture. The D denotes the drought year (1988) and the F denotes flood year (1993). The boldface D and F denote the control simulations for each year.

In section 2a, it is suggested that anomalously high soil moisture should result in an increase in net surface radiation due to an increase in net longwave and shortwave radiation (barring any cloud feedback). Figure 10 displays the surface radiation summary for the July 1988 and July 1993 simulations over the affected flood and drought region. Consistent with the sensitivities of convective and total rainfall to soil moisture (see Fig. 7), the simulations display a positive feedback between net surface radiation and soil moisture. This feedback is considerably stronger in 1988 than 1993. Like the rainfall sensitivity, the net surface radiation sensitivity is strongest in the 25%–50% initial soil moisture range. However, when the initial soil saturations exceeds 50%, the net surface radiation remains nearly constant. Upon further inspection, it is evident that the net surface longwave radiation sensitivity is the dominant factor responsible for the net radiation–soil moisture feedback. This suggests that the surface cooling and/or the greenhouse effect for water vapor play an important role in the soil moisture–rainfall feedback. On the other hand, the net surface shortwave radiation displays a strong negative feedback to initial soil moisture. This suggests that the decrease in net solar radiation associated with the increase in cloudiness dominates the increase associated with the decrease in albedo from the wetter soils and greener denser vegetation (not shown). BATS does not account for the relationship between vegetation albedo and soil moisture. Although we do not expect this vegetation feedback to be nearly as strong as the cloud feedback shown in these simulations, its inclusion should result in an increased sensitivity between net surface radiation and soil moisture.

In section 2b, we indicate that the sum of the surface latent and sensible heat fluxes should balance the net surface radiation at long timescales. Figure 11 displays a summary of the surface heat fluxes versus initial soil moisture for the affected flood and drought region. This figure indicates that the sum of the latent and sensible heat fluxes display a significant sensitivity to initial soil moisture and nearly balance the net surface radiation (within 5 W m^{2}). (Note that the range on the *y* axis of Fig. 11 is considerably larger than that of Fig. 10.) There is a small tendency for the ground heat flux to decrease with increasing soil moisture. This decrease is responsible for the slightly higher latent plus sensible heat flux sensitivity than net radiation. This mechanism, which is not discussed in section 2, results in an additional positive feedback between soil moisture and rainfall.

The response of the sum of the latent and sensible heat fluxes to soil moisture is determined by competing factors; the latent heat flux tends to increase with increasing soil moisture while the sensible heat flux tends to decrease. The simulations presented here indicate that the change in latent heat flux with changes in initial soil moisture tends to outweigh the corresponding change in sensible heat flux. In addition, associated with the decrease in sensible heat flux with increasing soil moisture is a decrease in PBL height (not shown). Thus, not only does the heat flux into the PBL increase, the heat flux added per unit depth of PBL increases even more significantly.

As alluded to in section 4b(1), the response of the latent heat flux to increases in soil moisture is nonlinear. The highest sensitivity occurs in between 25% and 50% of saturation, and the lowest sensitivities occur at the lower and higher soil saturations. The lack of sensitivity at the lower soil moisture values is primarily a result of transpiration ceasing due to the soil saturation falling below the wilting point. The lack of sensitivity at the higher soil moisture values is likely to be a result of evapotranspiration reaching the potential evaporation rate. Consequently, the highest sensitivities are seen in

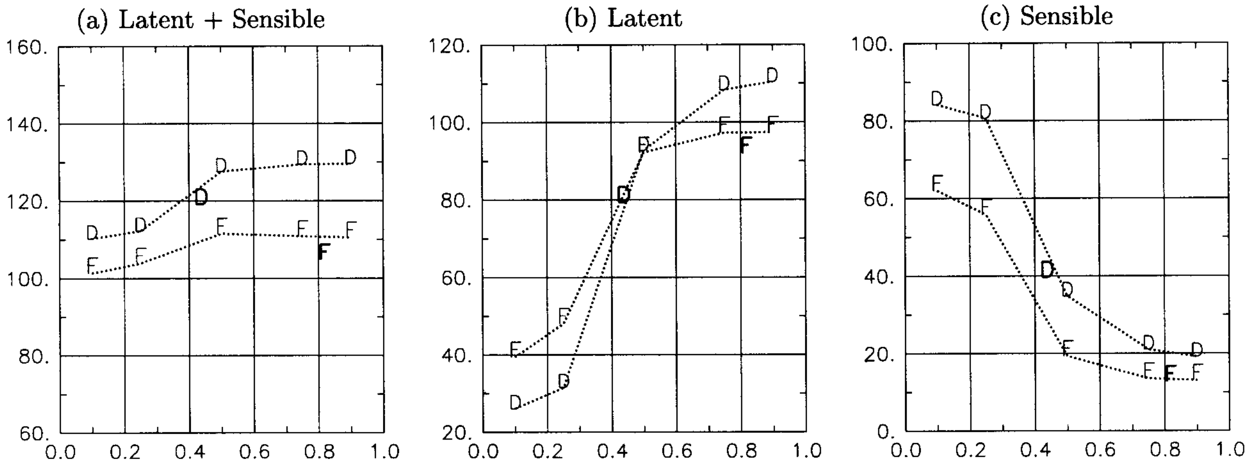

FIG. 11. Simulated monthly surface heat flux fields (W m^{-2}) for the 1988 and 1993 simulations as a function of initial soil saturation: (a) sensible + latent heat flux, (b) latent heat flux, and (c) sensible heat flux. Each data point represents the average of the May, Jun, Jul, Aug, and Sep simulations over the Midwest region outlined in Fig. 5 given the initial soil moisture. The D denotes the drought year (1988) and the F denotes flood year (1993). The boldface D and F denote the control simulations for each year.

between these extremes. This feature is carried over to the other surface fields presented in this paper.

Figure 12 displays soil moisture's impact on the surface moist static energy, temperature, and water vapor mixing ratio. As is the case with the sum of the latent and sensible heat fluxes, the magnitude of the surface MSE is determined by competing factors of temperature and humidity; anomalously high soil moistures are typically associated with lower near-surface temperatures and higher near-surface humidities. It is evident from Fig. 12 that soil moisture has a pronounced impact on both the low-level mixing ratio and temperature. Again like the sum of the latent and sensible heat fluxes, the moisture component outweighs the temperature component. In other words, the increase in mixing ratio with increasing soil moisture outweighs the decrease in temperature yielding an overall increase in MSE. Note that the MSE tends to be more sensitive to soil moisture than the sum of the sensible and latent heat fluxes. The increased sensitivity is likely to be a result of changes in the PBL height and in the amount of entrainment of low MSE air from above the PBL. Thus, increases in soil moisture tend to increase the flux of high MSE air into the PBL from below (increase in the sum of the latent and sensible heat fluxes), increase the MSE per unit mass of air (shallower PBL depth), and reduce the amount of entrained air of low MSE from above the PBL (decrease in the PBL growth rate). The combination of these additive processes is largely responsible for the increase in convective activity.

5. Summary of results and conclusions

In this study, we investigate the physical pathways and mechanisms responsible for the soil moisture–rainfall feedback using a modified version of NCAR's RegCM. The extreme drought of 1988 and the extreme flood of 1993 are used as representative events. Several questions are addressed: Are the persistent patterns in extreme hydrologic events maintained by external forcings or by internal mechanisms involving soil moisture? What role does the magnitude and direction of the soil moisture anomaly play? How does the timing of the soil moisture anomaly impact the strength of the feedback? What are the pathways responsible for the soil moisture–rainfall feedback?

To address these questions, we perform several numerical experiments. The soil saturation in each simulation is initialized with one of six different distributions of soil saturation ranging from 10% to 90% including an observed (control) distribution. Month-long simulations are performed during May, June, July, August, and September for 1988 and 1993. The following are the four main conclusions of this study:

1) Increases in initial soil moisture are shown to result in an increase in future rainfall over the Midwest. Soil moisture's impact on both the energy and water budgets proves to be crucial in determining the strength of the soil moisture–rainfall feedback.
2) The simulations indicate that there is an asymmetric response in the soil moisture–rainfall feedback due to the existence of multiple evapotranspiration regimes. The asymmetry is weighted such that the soil moisture rainfall–feedback is more responsive to a negative soil moisture perturbation than to a positive perturbation. This suggests that the soil moisture–rainfall feedback favors droughts in comparison with floods over the Midwest.
3) The simulations indicate that the soil moisture–rainfall feedback remains strong when the model is initialized at observed extremes in soil saturation. Based on these model results, one would conclude that soil moisture did play a significant role in main-

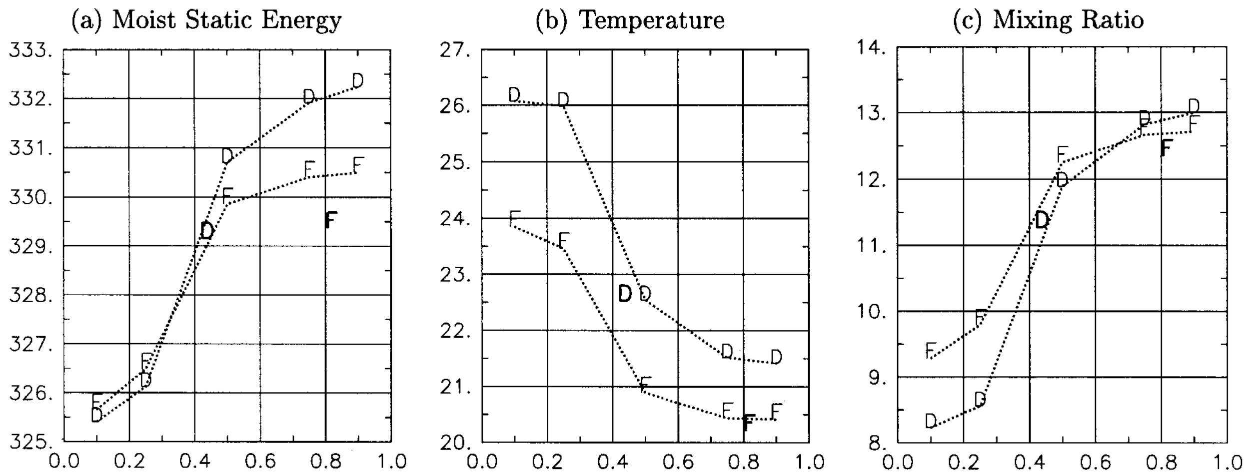

FIG. 12. Simulated monthly moist static energy (KJ kg^{-1}), temperature (°C), and water vapor mixing ratio (g kg^{-1}) for the 1988 and 1993 simulations as a function of initial soil saturation: (a) moist static energy, (b) temperature, and (c) mixing ratio. Each data point represents the average of the May, Jun, Jul, Aug, and Sep simulations over the Midwest region outlined in Fig. 5 given the initial soil moisture. The D denotes the drought year (1988) and the F denotes flood year (1993). The boldface D and F denote the control simulations for each year.

taining the persistence patterns of the drought of 1988 and the flood of 1993. However, the simulations suggest that the initiation of the events is likely to be a result of large-scale circulation anomalies.

4) During the late spring and summer, the strength of the soil moisture–rainfall feedback displays little dependence on the timing of the soil moisture anomaly. This suggests that knowledge of the soil moisture conditions during any of these months can improve the predictability of rainfall.

The choice of domain and convective parameterization is likely to have played a small but significant role in the outcome of the above conclusions. Future efforts are directed toward addressing these issues.

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