Summer rainfall forecast spread in an ensemble initialized with different soil moisture analyses

Eric Anthony Aligo  
*Iowa State University, ealigo@gmail.com*

William A. Gallus Jr.  
*Iowa State University, wgallus@iastate.edu*

Moti Segal  
*Iowa State University, segal@iastate.edu*

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Keywords
Agronomy, mathematical models, perturbation techniques, soil moisture, weather forecasting, Eta model, histogram, rain, correspondence analysis, ensemble forecasting, soil moisture, summer

Disciplines
Agronomy and Crop Sciences | Atmospheric Sciences | Geology

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Summer Rainfall Forecast Spread in an Ensemble Initialized with Different Soil Moisture Analyses

ERIC A. ALIGO AND WILLIAM A. GALLUS JR.

Department of Geological and Atmospheric Sciences, Iowa State University, Ames, Iowa

MOTI SEGAL

Department of Agronomy, Iowa State University, Ames, Iowa

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ABSTRACT

The performance of an ensemble forecasting system initialized using varied soil moisture alone has been evaluated for rainfall forecasts of six warm season convective cases. Ten different soil moisture analyses were used as initial conditions in the ensemble, which used the Weather Research and Forecasting (WRF) Advanced Research WRF (ARW) model at 4-km horizontal grid spacing with explicit rainfall. Soil moisture analyses from the suite of National Weather Service operational models—the Rapid Update Cycle, the North American Model (formerly known as the Eta Model), and the Global Forecasting System—were used to design the 10-member ensemble. For added insight, two other runs with extremely low and high soil moistures were included in this study. Although the sensitivity of simulated 24-h rainfall to soil moisture was occasionally substantial in both weakly forced and strongly forced cases, a U-shaped rank histogram indicated insufficient spread in the 10-member ensemble. This result suggests that ensemble forecast systems using soil moisture perturbations alone might not add enough variability to rainfall forecasts. Perturbations to both atmospheric initial conditions and land surface initial conditions as well as perturbations to other aspects of model physics may increase forecast spread. Correspondence ratio values for the 0.01- and 0.5-in. rainfall thresholds imply some spread in the soil moisture ensemble, but mainly in the weakly forced cases. Relative operating characteristic curves for the 10-member ensemble and for various rainfall thresholds indicate modest skill for all thresholds with the most skill associated with the lowest rainfall threshold, a result typical of warm season events.

1. Introduction

While deterministic forecasts of warm season rainfall have been improving (e.g., Olson et al. 1995), such forecasts of rainfall still show limited skill as determined by standard objective measures (e.g., Doswell et al. 1996; Fritsch and Carbone 2004). Improvements in rainfall forecasts have been achieved through the use of ensemble systems consisting of members with perturbed initial conditions (e.g., Hamill and Colucci 1997) and different physics (e.g., Stensrud et al. 2000; Jankov et al. 2005). Studies have shown, however, that individual members of an ensemble system are often too similar (Hou et al. 2001; Alhamed et al. 2002). These studies indicate that a multiple model and varied initial condition ensemble system represents both model and initial condition errors and provides better spread than a one-model system with different initial conditions alone. In light of these results, National Centers for Environmental Prediction (NCEP) began running the Short-Range Ensemble Forecasting (SREF) system (Du et al. 2004) in April 2001 with varied initial conditions and multiple models, to provide improved real-time probabilistic forecasts of temperature, rainfall, and other important meteorological quantities to weather forecasting offices. The above project yielded an improvement in forecast accuracy of the ensemble mean or median when the number of models in SREF was increased from one to three.

Sutton et al. (2006) showed that warm season rainfall forecasts are sensitively dependent on the soil moisture
initial condition uncertainty. Therefore, they suggested adding initial soil moisture perturbations to the existing suite of atmospheric perturbations and/or different models to improve ensemble spread. They used the Weather Research and Forecasting (WRF) Advanced Research WRF (ARW) model to show that two initial soil moisture analyses were different enough to produce 24-h rainfall differences at 5-km horizontal grid spacing that were similar to those obtained by using two different convective schemes at 20-km horizontal grid spacing. Sutton et al. (2004) used version three of the fifth-generation National Center for Atmospheric Research–Pennsylvania State University (NCAR–PSU) Mesoscale Model (MM5V3) to generate six sets of five-member ensembles, with three sets using the Grell convective scheme and three using the Kain–Fritsch II convective scheme. For each convective scheme, one ensemble was initialized with a set of atmospheric perturbations, the second ensemble with a set of soil moisture perturbations, and the third with a set of combined atmospheric and soil moisture perturbations. Based on one case study, the variability added by the soil moisture perturbations was similar to that added by the atmospheric perturbations. These results reflect the sensitivity of dynamical and thermodynamical forcing of convection to soil moisture. The effects of soil moisture on rainfall can be associated with dynamical modifications of atmospheric systems as well as the generation of mesoscale circulations through the formation of spatial sensible heat flux gradients (Pielke 2001). On the other hand, various studies (e.g., Clark and Arritt 1995; Findell and Eltahir 2003), while considering thermodynamical forcing only, found that soil moisture, by affecting the partitioning of the surface sensible and latent heat fluxes, may play an important role in the development of convection.

The present paper evaluates summer rainfall forecast spread in weakly forced and strongly forced events in an ensemble initialized with different soil moisture analyses by providing pertinent illustrative features and quantitative analyses. The impact of the uncertainties in the National Weather Service (NWS) Rapid Update Cycle (RUC), North American Model (NAM; formerly known as the Eta Model), and Global Forecasting System (GFS) soil moisture analyses on forecasted rainfall is also examined. As such, the paper provides an extension and complementary insight to the study by Sutton et al. (2006). Section 2 describes the WRF configuration adopted in the study and the method for generating the different soil moisture analyses. Section 3 describes the results, and section 4 provides a summary and conclusions.

2. Data and methodology

The WRF ARW model (Skamarock et al. 2005) was run without the use of a convective parameterization over an approximately 1000 km × 1000 km domain centered over Iowa (see Fig. 4 for a map of the domain) with the following configuration: 4-km horizontal grid spacing, 31 vertical levels, WRF single‐moment six‐class microphysics (Hong and Lim 2006), Yonsei University planetary boundary layer scheme (Hong and Pan 1996; Hong and Dudhia 2004), Monin–Obukhov surface layer scheme (Janjić 2001), along with the Rapid Radiative Transfer Model (Mlawer et al. 1997), and Dudhia (1989) longwave and shortwave radiation schemes, respectively. The RUC land surface model (RUC LSM; Benjamin et al. 2004) was used in all model runs.

Six cases were evaluated, all initialized at 1200 UTC and integrated over a 24-h period using the NWS RUC for the atmospheric initial and lateral boundary conditions. With the limited domain in the present study, lateral boundary condition errors may have adverse impacts (Warner et al. 1997), but the short integration time should somewhat lessen these impacts on the systems of interest. The selected cases were strictly confined to those with mainly clear skies through early afternoon and no rainfall until mid- or late afternoon, such that the impact of soil moisture on convection might be maximized. There were three weakly forced events (15–16 June 2003, 20–21 June, and 24–25 July 2005) and three strongly forced events (9–10 June 2003, 17–18 July, and 9–10 August 2005). The weakly forced events were defined to be those associated with weak mid- to upper-level winds (500-hPa winds less than 15 m s\(^{-1}\)) and either weak frontal systems (near-surface cross‐frontal horizontal temperature gradients less than 5°C over 100 km), no frontal systems at all, or events with outflow boundaries in the region near the convection. Strongly forced events had 500-hPa winds of at least 20 m s\(^{-1}\) and/or near-surface cross‐frontal horizontal temperature gradients of at least 7°C over 100 km.

Soil moisture analyses from the suite of NWS operational models on a variety of grids (20-km RUC, 40-km NAM, and 1° GFS) available through the National Oceanic and Atmospheric Administration Global Systems Division of the Earth System Research Laboratory were used to initialize three of the ensemble members, while additional soil moisture analyses were constructed to initialize seven more ensemble members. The seven additional members were constructed based on the assumption that soil moisture errors are distributed normally with the mean and standard deviation.
determined from the three NWS models. As such, the 10-member ensemble consisted of the three NWS model soil moisture analyses in addition to soil moisture analyses representing the mean, 25th percentile, 75th percentile, and the ±1 and ±2 standard deviations (sd) of the operational dataset (hereafter these runs are denoted, respectively, as RUC, NAM, GFS, mean, 25th, 75th, ±1_sd, and ±2_sd). For added insight, there were two other runs: one initialized with an extremely dry soil (hereafter ED; volumetric soil moisture set to the residual values, i.e., minimum possible volumetric soil moisture content) and the other, an extremely wet soil (hereafter EW; volumetric soil moisture values set to the saturation point). Hence, the results presented in this paper are based on a total of 72 simulations. Figure 1 illustrates the 0–10-cm volumetric soil moisture values from each ensemble member plotted on a standard normal curve at a single grid point for one of the cases discussed in more detail later (note that the plot features will vary with soil depth and with each grid point). The 20-km RUC volumetric soil moisture analyses are defined at the surface and 5, 20, 40, 160, and 300 cm below the surface, unlike the NAM and GFS analyses, which are defined for the 0–10-, 10–40-, 40–100-, and 100–300-cm layers below the surface. To be consistent with the NAM and GFS, the volumetric soil moisture from the RUC levels was interpolated to the NAM and GFS layers. Also, because the GFS volumetric soil moisture analyses from the 2003 cases are defined only for the 0–10- and 10–200-cm layers, they were linearly interpolated to the four layers mentioned above. It should be noted that soil moisture values lower than the residual value that arose during the construction of the different soil moisture analyses were assigned the residual value.

The seven ensemble members constructed from the three NWS models were assigned soil moisture perturbations uniformly across the model domain rather than being random and spatially varied. Consequently, the group of ensemble members in our study represents a vast range in the surface latent/sensible heat flux forcing, and likely provides better potential for ensemble spread due to the broader-scale thermodynamic effects on the convective boundary layer than with randomly assigned soil moisture analyses. Such differences in the surface flux forcing between ensemble members can also noticeably modulate the dynamical forcing of fronts (Koch et al. 1995) and the low-level jet (McCulloch 1988) and, therefore, convection. Additionally, as can be seen later (Fig. 3), uniformly drying or moistening the soil may generate a spatial nonuniform change in the latent heat flux and correspondingly in the sensible heat flux. This can enhance or weaken existing local thermal circulations or develop new circulations where previously none existed. Nevertheless, the method adopted in the present study to create the different soil moisture analyses is not necessarily the one that would be appropriate for an operational ensemble system.

The skill of the ensemble forecast system was evaluated using relative operating characteristic (ROC) curves (Mason and Graham 1999). If the area under a ROC curve is less than 0.5, the forecast does not have skill, while an area of 0.7 is said to represent the lower limit of a useful forecast system (Buizza et al. 1999). The skill of individual members was evaluated using the equitable threat score (ETS; Schaefer 1990) and bias, where

$$ETS = \frac{CFA - CHA}{F + O - CFA - CHA}, \quad (1)$$

$$CHA = O \frac{F}{V}, \quad (2)$$

and

$$bias = \frac{F}{O}. \quad (3)$$

In (1), (2), and (3), each variable indicates the number of grid points at which (i) rainfall was correctly forecasted to exceed the specified threshold (CFA), (ii) rainfall was forecasted to exceed the threshold (F), (iii) rainfall was observed to exceed the threshold (O), and (iv) a correct forecast would occur by chance (CHA), where V is the total number of evaluated grid points.
The NCEP 4-km gridded stage IV multisensor data (Baldwin and Mitchell 1997) were used for verification. The spread of the ensemble rainfall forecasts was evaluated through the use of a rank histogram (Hamill 2001) and the correspondence ratio (CR; Stensrud and Wandishin 2000). A rank histogram illustrates the rank of the observed rainfall amounts with respect to the forecast rainfall amounts from all ensemble members. A U-shaped rank histogram means the observed rainfall amounts are lower and higher than the amounts from any of the ensemble members and indicates the ensemble system exhibits insufficient spread (vice versa for a bell-shaped rank histogram). The CR is defined as the ratio of the area of intersection ($I$) of all individual rainfall field values to the area of union ($U$) of the same rainfall field values,

$$CR = \frac{I}{U},$$

where $I$ and $U$ are defined using threshold values of rainfall. The CR can provide information about the spatial divergence of ensemble members. In the extreme case, $CR = 0$ means that not all of the ensemble members overlap, implying the least spatial agreement among members, whereas $CR = 1$ means all ensemble members completely overlap, implying the most spatial agreement among members.

3. Results

In the following, illustrative features of ensemble rainfall spread are provided in addition to a quantita-
tive skill and spread evaluation. Detailed discussion of the initial volumetric soil moisture and rainfall difference features primarily uses the 9–10 June 2003 case because it had the most widespread 24-h rainfall differences among all six cases.

a. An evaluation of variations in rainfall due to changes in soil moisture for the 9–10 June 2003 case

1) Synoptic overview

During the evening hours of 9 June 2003, a surface cold front and associated 500-hPa short-wave trough moved into the northern high plains by 0000 UTC 10 June 2003 (Figs. 2a and 2b). The cold front (not plotted in the analysis) extended south and westward from a low pressure system in extreme southern South Dakota and northern Nebraska at 0000 UTC 10 June 2003 with its attendant warm front situated along the Iowa–Nebraska border. Convection initiated east of a dryline in central Nebraska shortly after 0000 UTC and developed into a bow echo. The bow echo then moved into southeastern Nebraska, southwestern Iowa, and northwestern Missouri between 0500 and 0600 UTC. The cold front then swept across Nebraska overnight and extended from southwestern Iowa to northern Iowa by 1200 UTC 10 June 2003 (Fig. 2c), with the associated short-wave trough in the eastern part of the simulation domain (Fig. 2d). A detailed analysis of this Bow Echo and Mesoscale Convective Vortex Experiment case can be found in Wheatley et al. (2006).

2) Initial volumetric soil moisture patterns

Soil moisture affects the amount of evapotranspiration; hence, it influences the partitioning of the net available surface thermal energy between the latent and sensible heat fluxes and the related impact on convection. Thus, it is useful to illustrate the interrelationship between the evapotranspiration and the volumetric soil moisture in the RUC LSM, which is adopted in the present study. Figure 3 is a schematic illustration of the transpiration rate function versus volumetric soil moisture for vegetated surfaces. Between the wilting point (defined as the volumetric soil moisture at which soil moisture tension is too high for water to be extracted by the vegetation roots) and the reference point (the maximum volumetric soil moisture in which water in the soil can hold against gravity), the transpiration rate function increases fairly sharply as soil moisture increases. Between the reference point and saturation point, and similarly between the wilting point and the residual point, the transpiration rate function remains nearly constant. A similar pattern as in Fig. 3 is typical also for evaporation from bare soil (Lee and Pielke 1992). Thus, the largest sensitivity of rainfall to changes in soil moisture occurs when the soil moisture is in the window between the wilting point and reference point. Considering the possible volumetric soil moisture values indicated in Fig. 3, this window might be relatively narrow. Overall, based on the characteristics of Fig. 3, it is suggested that the sensitivity of rainfall to variations in soil moisture in this study depends on the values of the volumetric soil moisture and the magnitude of the variations.

The 0–10-cm soil layer initial volumetric soil moisture for the 9–10 June 2003 case was generally between 0.25 and 0.30 in the mean run with values as high as 0.33 in southeastern Minnesota and as low as 0.17 in northern Nebraska (Fig. 4a). In the +2_sd run (Fig. 4b), volumetric soil moisture was generally between 0.35 and 0.40 with values as high as 0.44 in southeastern Minnesota and southeastern Iowa (in some areas these values were at or near the saturated values prescribed in the EW run) and values as low as 0.24 in extreme southern South Dakota. The 0–10-cm soil layer initial volumetric soil moisture differences between the +2_sd and −2_sd (denoted hereafter as Δ2_sd) runs were generally between 0.2 and 0.3 (Fig. 4c). Differences between the +1_sd and −1_sd (denoted hereafter as Δ1_sd) runs were about half of those in Δ2_sd (Fig. 4d). Following Fig. 3, the above differences imply possibly large variations in daytime evapotranspiration (i.e., the surface latent heat flux) and, thus, also in the surface sensible heat flux. An examination of the simulated
sensible and latent heat fluxes confirmed such patterns (not shown). It should be pointed out that the magnitudes and spatial patterns of the volumetric soil moisture and volumetric soil moisture differences were similar at the next lower soil layer centered at 20 cm in the RUC LSM (not shown). The differences in the operational soil moisture analyses were generally between 0.05 and 0.1 (Figs. 5a–c), only a little smaller in magnitude than the differences in the $\Delta 1_{sd}$ plot. It is worth noting that similar features of volumetric soil moisture differences were common also in the other five studied cases.

3) ILLUSTRATIVE RAINFALL PATTERNS

The 24-h observed rainfall (Fig. 6a) for 9–10 June 2003 reached 60 mm in far northeastern Nebraska and in a narrow rainband extending from eastern Nebraska southeastward into northwestern Missouri. Lighter rainfall amounts were observed in southern and northern Iowa as well as in the northern part of the domain. The mean run and the $+2_{sd}$ run (Figs. 6b and 6c, respectively) both missed the rainband in the southwestern part of the domain with the mean run and the $+2_{sd}$ run predicting up to 80 and 100 mm in central Nebraska.
Iowa, respectively (the $+2_{sd}$ run forecasted rainfall is shown since the $+2_{sd}$ and the $-2_{sd}$ runs are likely to represent the maximum possible contrast in rainfall forecasts within the ensemble). Between 20 and 60 mm of rain was falsely predicted by both simulations in eastern South Dakota and southern Minnesota. Rainfall differences were as high as 70 mm in the EW − ED plot (Fig. 6d), and surprisingly slightly more intense in the $\Delta 2_{sd}$ plot (Fig. 6e) with differences as high as 80 mm. Differences were generally around 20 mm in the $\Delta 1_{sd}$ difference plot (Fig. 6f). Interestingly, the widespread areas of positive differences in central Iowa and negative differences to the south seen in the EW − ED and $\Delta 2_{sd}$ plots were reduced in the $\Delta 1_{sd}$ difference plot.

Also of note is the general tendency for less rainfall in southwestern Iowa and parts of Missouri in the EW and $+2_{sd}$ runs as inferred by the negative rainfall differences in the EW − ED and $\Delta 2_{sd}$ plots. The less rainfall in southwestern Iowa and parts of Missouri in the EW and $+2_{sd}$ runs was the result of convection initiating 1–2 h later and not developing as far south ahead of the dryline in Nebraska as compared with the $-2_{sd}$ and ED runs. By the end of the simulation, an outflow boundary and its associated convection had propagated farther south and east in the ED and $-2_{sd}$ runs than in the EW and $+2_{sd}$ runs. Because the drier soil runs had a drier boundary layer, enhanced evaporative cooling could explain the more rapid movement of the outflow boundary.

When considering runs using the initial volumetric soil moisture from the operational model (RUC, NAM, and GFS) analyses (Fig. 7), rainfall difference features for GFS − NAM (Fig. 7a) resembled those seen in the $\Delta 1_{sd}$ plot, with rainfall differences generally around or less than 10 mm and negative differences north of positive differences in southern Iowa. For NAM − RUC (Fig. 7b) and GFS − RUC (Fig. 7c), the rainfall difference features resembled those seen in the EW − ED and $\Delta 2_{sd}$ plots with rainfall differences generally exceeding 10 mm and with positive differences north of negative differences in southern Iowa. These results are consistent with the RUC having much lower values of domain-averaged initial volumetric soil moisture than the GFS and NAM (not shown).

It is worth pointing out that 24-h rainfall difference features were generally similar in all six cases. The greatest rainfall difference was 100 mm and occurred in

Fig. 5. The 0–10-cm soil layer initial volumetric soil moisture differences for the 9–10 Jun 2003 case for (a) NAM–RUC, (b) GFS–NAM, and (c) GFS–RUC. Contour levels are −0.2, −0.1, −0.05, 0.05, 0.1, and 0.2.
The 24-h rainfall (mm) for 9–10 Jun 2003 from (a) observations, (b) the mean run, (c) the $+2_{sd}$ run, and for (d) EW–ED, (e) $\Delta_{2_{sd}}$, and (f) $\Delta_{1_{sd}}$. The contour interval in Figs. 6a–c is 20 mm. The contour levels in Figs. 6d–f are $-80$, $-60$, $-40$, $-20$, $-10$, $10$, $20$, $40$, and $60$ mm.

Fig. 6.
the 9–10 August 2005 case between both the +2_sd and −2_sd runs and the NAM and RUC runs.

Figure 8 presents a time series from 0000 to 1200 UTC of the observed rain rate and forecasted rain rates from all model runs at a grid point in the region of negative rainfall differences indicated by the closed circle in Fig. 6d (with volumetric soil moisture values specified in Fig. 1). Following Fig. 8, the peak observed and 25th percentile simulated rain rates were both around 23 mm h⁻¹. The largest rain-rate difference occurred between the ED and 75th percentile runs, with the drier soil simulated peak rain rate of over 40 mm h⁻¹ comparing with less than 10 mm h⁻¹ in the wetter soil run. Seven of the 12 runs forecasted the heaviest rain rate between 0700 and 0800 UTC, 1 h after the other five runs forecasted their peak rain rate and 3 h after the observed peak rain rate. In general, the rain rates peaked earliest in the driest soil runs (ED, −2_sd, RUC, −1_sd, and 25th percentile), reflecting the support that the increased sensible heat flux had on earlier initiation of convection. From the run with the lowest initial volumetric soil moisture to the run with the highest initial volumetric soil moisture there was an overall downward trend in the peak rain rate at this specific location but the transition was not smooth. Similar features were present (not shown) also at three other selected grid points (indicated by triangles in Fig. 6d) in the region of positive and negative rainfall differences. At these three grid points, peak rain-rate differences among individual runs varied between 21 and 26 mm, while the timing of the peak simulated rainfall amounts differed by 1–2 h.

b. An evaluation of variations in rainfall due to changes in soil moisture among different cases

1) SENSITIVITY OF AFTERNOON AND NIGHTTIME RAINFALL TO SOIL MOISTURE

An evaluation of the sensitivity of afternoon and nighttime rainfall to an extreme soil moisture variation can provide additional insight into the potential ensemble rainfall spread. In the following, such an evaluation is presented for two cases.

Figure 9 presents the 1800–0000 UTC rainfall for 9–10 June 2003 in the ED (Fig. 9a) and EW (Fig. 9b) runs, and the 0600–1200 UTC rainfall in the ED (Fig. 9c) and EW (Fig. 9d) runs. Between 1800 and 0000 UTC, the ED run produced up to 13 mm of rain in

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Fig. 7. The 24-h rainfall differences (contoured every 10 mm) for the 9–10 Jun 2003 case for (a) GFS – NAM, (b) NAM – RUC, and (c) GFS – RUC.
northeastern Nebraska, likely associated with a warm front, where the EW run had no rain at all. From 0600 to 1200 UTC, the EW run ended up producing slightly more intense rain in most of the domain. It should be noted that these 6-h rainfall difference features of more afternoon frontal rainfall in the ED runs and more intense late-night rainfall in the EW runs were weakest in this case, but much more pronounced in the four other frontal/outflow boundary cases. For example, Figs. 10a–d, valid at the same hours as in Figs. 9a–d, but for 20–21 June 2005, show that the ED run produced much more rainfall in the afternoon as compared with the EW run, whereas the EW run produced more intense rainfall during the late-night period. The synoptic environment for this case included a deep 500-hPa West Coast trough and a broad ridge from the central plains to the East Coast along with a front in central Minnesota that remained stationary throughout the 24-h integration (not shown). Strong convection in southern South Dakota at 1200 UTC 20 June 2005 propagated eastward into central Minnesota and then southward into central Iowa by 0000 UTC 21 June 2005. Finally, it is worth noting that similar 6-h rainfall features were evident also in the +2_sd runs; however, they were less noticeable in the +1_sd runs (not shown) in the above two cases and two additional frontal/outflow boundary cases.

The above rainfall features of more afternoon frontal rainfall in the ED runs and more intense late-night rainfall in the EW runs occurred in five of six cases, all of which had observed convective systems forced by fronts or outflow boundaries. Increased surface sensible heat flux over dry surfaces increases destabilization of the lower atmosphere, thus enhancing frontal uplift (e.g., Koch et al. 1995) and promoting earlier convective initiation. In contrast, higher daytime latent heat fluxes in the wetter soil runs, although delaying the initiation of convection, contribute to higher low-level moisture and to increased rainfall in those runs during the late-night period when convection matures.

2) DOMAIN-AVERAGED RAINFALL FOR ALL SIMULATED CASES

Figure 11 presents the domain-averaged rainfall for all six cases and the spread among all simulations. The EW and +2_sd runs were often very similar to the +2_sd run, forecasting more domain-averaged rainfall than the EW run in three of six cases (9–10 and 15–16 June 2003, and 24–25 July 2005). The ED and −2_sd runs were also similar, with the −2_sd run forecasting less domain-averaged rainfall than the ED run in one case (20–21 June 2005). Initial volumetric soil moisture values in the +2_sd and EW runs were close to or at the soil saturation point in many areas. Hence, the forecasted rainfall from these runs should be similar since, as implied by Fig. 3, almost no difference exists in the surface latent heat fluxes between these two runs. Similarly, the volumetric soil moisture totals in the −2_sd
and ED runs were both close to or at the soil residual point, where the surface latent heat fluxes are effectively zero.

Among just the three runs using operational soil moisture analyses, the NAM produced the most rainfall in three cases, and the least rainfall in three cases as well. The GFS produced the most rainfall in three cases and the least rainfall in one case. The RUC model never produced the most rainfall, but produced the least rainfall in two cases. The RUC never produced the most rainfall possibly because this model had the lowest domain-averaged initial volumetric soil moisture in five of six cases as compared with the NAM and GFS.

Domain-averaged rainfall differences among individual ensemble members ranged from 9% in the 9–10 June 2003 case to 50% in the 15–16 June 2003 cases, but it must be noted that the domain-averaged rainfall in the 15–16 June 2003 case was less than 0.35 mm, a factor of 5 smaller than the case with the next lowest domain-averaged rainfall. Note, the greatest domain-averaged rainfall difference (absolute value) between any two ensemble members (not including the ED and EW runs) was largest for the cases that had the largest domain-averaged rainfall. Among the three wettest cases, one of them was a weakly forced case (20–21 June 2005) and had a peak ensemble member rainfall difference (1.1 mm) equal to or slightly larger than those in the 9–10 June 2003 and 9–10 August 2005 strongly forced cases.

c. Analyses of skill and ensemble spread

Substantial rainfall differences between the ensemble members were evident in some locations in all six cases. Here, we present quantification of the ensemble and model skill in addition to the ensemble spread.

The skill of the ensemble was evaluated using ROC curves, while the skill of individual members was evalu-
Figure 12 illustrates the ROC curves for various rainfall thresholds (stated in inches; 1 in. = 25.4 mm) and shows all curves above the no-skill diagonal, indicating some skill, with the area under the curve, computed using the trapezoidal method, the largest (0.74) for the lowest rainfall threshold and the smallest (0.58) for the highest rainfall threshold—results common for warm season rainfall (e.g., Gallus and Segal 2004). According to Hamill and Juras (2007), variations in rainfall climatology within a model domain can result in biased ROC values if computed once for the entire domain. They recommend calculating the probability of false detection (POFD) and probability of detection (POD) independently for subdomains within the larger model domain that have the same rainfall climatological probability. Weighted-average values of the POFD, POD, and ROC are then obtained from the subdomains. We performed a bulk test and computed the POFD and POD separately for each of four equally sized subdomains. These subdomains accounted for, to a first approximation, the observed east–west and north–south variations in the rainfall climatological probability in the simulated domain. Average values of the POFD, POD, and ROC were then obtained from the four subdomain values of POFD and POD. The new ROC calculations resulted in only a slightly less skillful ensemble than is shown in Fig. 12.

Table 1 shows the ETS and bias scores among all ensemble members as being similar, indicating that, potentially, the 10 members could make a good ensemble. Table 1 also shows a low bias for lighter rainfall thresholds and a high bias for heavier thresholds, similar to what was found in Jankov et al. (2007) and Shaw (2004) for runs with explicit rainfall, a feature uncommon for model runs using convective parameterizations (e.g., Gallus et al. 2005).

Figure 13 illustrates U-shaped rank histograms in both the weakly forced and strongly forced cases suggesting a lack of spread, but with slightly more spread in the weakly forced cases. Jankov et al. (2006) found...
much better spread (flat rank histograms) with their mixed physics and varied initial condition ensemble. The values of CR were 0.34 for the 0.01-in. rainfall threshold (Fig. 14) and 0.13 for the 0.5-in. rainfall threshold in the weakly forced cases. Gallus and Bresch (2006) compared output from only two model configurations, differing in dynamic core, physical schemes, and initial conditions, and obtained 24-h-averaged CR values as low as 0.31 for the 0.01-in. rainfall threshold and 0.066 for the 0.5-in. threshold. Even when the physics alone were changed, they obtained CR values equal to or less than the values obtained in the present study. According to the definition of CR, values of CR should decrease as the number of ensemble members increases. Thus, based on the results above, the larger 10-member ensemble in the present study lacks spread. The CR values were 40%–50% higher in the strongly forced cases, indicating that the weakly forced cases were the most spatially divergent. One might expect the weakly forced events to have more spread because without a strong lifting mechanism weakly forced events are governed by thermodynamic forcing, low-level wind/moisture convergence features, and possibly by local circulations induced by spatial sensible heat flux variations. As stated previously, these convection-supporting mechanisms are sensitive to the soil moisture and its spatial distribution. In contrast, rainfall in strongly forced events is likely to be more focused in the vicinity of fronts and/or upper-level troughs. It should be noted that spread typically grows with time and might be larger if the model was integrated beyond 24 h; however, the lack of perturbed lateral boundary conditions in this study would slow or limit this growth of spread (Nutter et al. 2004).

4. Summary and conclusions

In recent years, studies have used both different initial conditions and physics to improve the ensemble spread and forecast accuracy of rainfall predictions.
Sutton et al. (2006) suggested perturbing the land surface state in addition to perturbing the atmosphere in order to increase the spread in the ensemble members. In this study, as a first step, we evaluated the impact of perturbations in volumetric soil moisture while the initial atmospheric conditions were not perturbed. We used 10 different soil moisture analyses in an ensemble and evaluated the spread in three strongly forced and three weakly forced cases. The cases chosen were ones where the timing of rainfall should maximize the sensitivity to soil moisture.

Subjectively, 24-h rainfall differences were noticeable (>10 mm) over widespread areas when different soil moisture analyses were used to initialize the WRF ARW model. This study illustrated, in detail, one case where it was found that at individual grid points the timing and magnitude of the peak rainfall rate differed among the ensemble members. In four of five cases with frontal-/outflow-induced convection, the ED runs had more frontal rainfall during the 1800–0000 UTC time period as compared with the EW runs, while more intense rainfall occurred by the 0600–1200 UTC period in the EW runs (similar features were evident also in the $\pm 2_{\text{sd}}$ and to a lesser degree in the $\pm 1_{\text{sd}}$ runs). This result agrees with previous findings that dry surfaces better promote daytime frontal uplift as compared with wet surfaces. In the EW run, higher levels of

<table>
<thead>
<tr>
<th>Run</th>
<th>0.01</th>
<th>0.10</th>
<th>0.25</th>
<th>0.50</th>
<th>0.75</th>
<th>1.00</th>
</tr>
</thead>
<tbody>
<tr>
<td>$+2_{\text{sd}}$</td>
<td>0.222 (0.7)</td>
<td>0.190 (0.7)</td>
<td>0.144 (0.8)</td>
<td>0.098 (1.1)</td>
<td>0.061 (1.7)</td>
<td>0.034 (2.7)</td>
</tr>
<tr>
<td>75th</td>
<td>0.218 (0.7)</td>
<td>0.184 (0.7)</td>
<td>0.143 (0.8)</td>
<td>0.099 (1.0)</td>
<td>0.057 (1.5)</td>
<td>0.025 (2.3)</td>
</tr>
<tr>
<td>$+1_{\text{sd}}$</td>
<td>0.208 (0.7)</td>
<td>0.175 (0.7)</td>
<td>0.140 (0.8)</td>
<td>0.101 (1.0)</td>
<td>0.057 (1.5)</td>
<td>0.026 (2.4)</td>
</tr>
<tr>
<td>Mean</td>
<td>0.206 (0.7)</td>
<td>0.182 (0.7)</td>
<td>0.143 (0.7)</td>
<td>0.099 (0.9)</td>
<td>0.055 (1.3)</td>
<td>0.021 (2.1)</td>
</tr>
<tr>
<td>$-1_{\text{sd}}$</td>
<td>0.219 (0.6)</td>
<td>0.175 (0.6)</td>
<td>0.135 (0.7)</td>
<td>0.094 (0.8)</td>
<td>0.052 (1.3)</td>
<td>0.016 (2.0)</td>
</tr>
<tr>
<td>25th</td>
<td>0.212 (0.6)</td>
<td>0.179 (0.6)</td>
<td>0.138 (0.7)</td>
<td>0.097 (0.9)</td>
<td>0.053 (1.3)</td>
<td>0.021 (2.1)</td>
</tr>
<tr>
<td>$-2_{\text{sd}}$</td>
<td>0.223 (0.6)</td>
<td>0.179 (0.6)</td>
<td>0.148 (0.6)</td>
<td>0.113 (0.8)</td>
<td>0.059 (1.1)</td>
<td>0.018 (1.7)</td>
</tr>
<tr>
<td>RUC</td>
<td>0.213 (0.6)</td>
<td>0.176 (0.6)</td>
<td>0.136 (0.7)</td>
<td>0.103 (0.9)</td>
<td>0.058 (1.2)</td>
<td>0.021 (2.0)</td>
</tr>
<tr>
<td>NAM</td>
<td>0.209 (0.7)</td>
<td>0.177 (0.7)</td>
<td>0.134 (0.7)</td>
<td>0.096 (1.0)</td>
<td>0.051 (1.4)</td>
<td>0.023 (2.2)</td>
</tr>
<tr>
<td>GFS</td>
<td>0.213 (0.7)</td>
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<td>0.145 (0.8)</td>
<td>0.107 (1.0)</td>
<td>0.058 (1.5)</td>
<td>0.025 (2.3)</td>
</tr>
</tbody>
</table>

**Fig. 13.** Rank histograms for 24-h rainfall for the 10-member ensemble averaged over the (a) strongly and (b) weakly forced cases.

**Fig. 14.** Correspondence ratio for the 0.01- and 0.5-in. rainfall thresholds for the 10-member ensemble averaged separately over the weakly and strongly forced cases.
low-level moisture may enhance the amounts and/or areal coverage of rainfall later in the night. Among runs initialized with volumetric soil moisture from the operational models, the NAM had either the most or least domain-averaged rainfall, while the RUC never had the most rainfall. The RUC had the lowest domain-averaged initial volumetric soil moisture in five of six cases as compared with the NAM and GFS, and this might explain why the RUC never had the most rainfall.

The ETS and bias values for different rainfall thresholds, averaged over all six cases, were nearly the same for each ensemble member suggesting that the 10 members, potentially, could make a good ensemble if the ensemble spread is large enough. The ensemble spread, however, was found to be insufficient based on a U-shaped rank histogram in both the weakly forced and strongly forced cases, although there was more spread in the weakly forced cases. Also supporting this finding, values of CR were lowest in the weakly forced cases. The slight difference in spread is attributed to the fact that weakly forced events lack strong lifting mechanisms and are sensitive to soil moisture and its spatial variability. In contrast, rainfall in strongly forced events is likely to be more focused in the vicinity of fronts and/or upper-level troughs. ROC curves indicated the ensemble forecast system did have modest skill for all rainfall thresholds with the largest area under the ROC curve occurring with the lightest rainfall threshold, a result common for warm season rainfall.

In conclusion, the precipitation amount within convective systems can be strongly sensitive to soil moisture perturbations, but the perturbations, if only applied to soil moisture, might not add enough variability to rainfall forecasts over the entire domain. The variability might be greatly increased by perturbing both atmospheric initial conditions and land surface initial conditions together as well as perturbing other aspects of the model physics (a subject to be addressed in future work). The results in this study, however, may be influenced by the design of the model, including the specific selection of the various model physical schemes (particularly the land surface module). Additionally, the method used to create the soil moisture analyses is not necessarily one that would be the most appropriate for an operational ensemble system. Further tests should be performed to determine if the use of different soil moisture analyses might be a helpful component in a larger ensemble system using mixed models, parameterizations, and initial atmospheric conditions.

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