Sensitivity of Northern Great Plains Convection Forecasts to Upstream and Downstream Forecast Errors

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(Manuscript received 12 September 2016, in final form 27 February 2017)

ABSTRACT

The role of earlier forecast errors on subsequent convection forecasts is evaluated for a northern Great Plains severe convective event on 11–12 June 2013 during the Mesoscale Predictability Experiment (MPEX) by applying the ensemble-based sensitivity technique to Weather Research and Forecasting (WRF) Model ensemble forecasts with explicit convection. This case was characterized by two distinct modes of convection located 150 km apart in western Nebraska and South Dakota, which formed on either side of an axis of high, lower-tropospheric equivalent potential temperature \( u_e \). Convection forecasts over both regions are found to be sensitive to the position of this \( u_e \) axis. The convection in Nebraska is sensitive to the position of the western edge of the \( u_e \) axis near an upstream dryline, which modulates the pre-convective \( u_e \) prior to the diurnal maximum. In contrast, the convection in South Dakota is sensitive to the position of the eastern edge of the \( u_e \) axis near a cold front, which also modulates the preconvective \( u_e \) in that location. The position of the \( u_e \) axis is modulated by the positions of both upstream and downstream mid- to upper-tropospheric potential vorticity anomalies, and can be traced backward in time to the initial conditions. Dropsondes sampling the region prior to convective initiation indicate that ensemble members with better representations of upstream conditions in sensitive regions are associated with better convective forecasts over Nebraska.

1. Introduction

The predictability of mesoscale convection remains a significant forecasting challenge, mainly due to rapid error growth from nonlinear processes on the convective scale (e.g., Park 1999; Hohenegger and Schär 2007). Lorenz (1969) argued that small-scale initial condition uncertainty can grow upscale and hence limit mesoscale predictability (e.g., Zhang et al. 2003, 2006; Hohenegger and Schär 2007). On the other hand, mesoscale features associated with the larger-scale flow (e.g., fronts, external forcing from orography, thermal contrasts) may be characterized by enhanced predictability (Anthes et al. 1985). Nevertheless, small perturbations in the larger-scale flow can also amplify and propagate downscale, which can adversely impact mesoscale forecasts (e.g., Nuss and Miller 2001; Bei and Zhang 2007; Durran et al. 2013; Durran and Gingrich 2014; Durran and Weyn 2016). Therefore, the skill of convective forecasts could benefit from understanding the sensitivity of convective forecasts to forecast fields at earlier times. Sensitivity analysis is one method to evaluate the processes responsible for a particular forecast outcome or metric. Most critically, this method can elucidate how changing a state variable at one particular time impacts subsequent forecast metrics. One approach to sensitivity analysis involves using the adjoint of a forecast model (e.g., Hall et al. 1982; Errico and Vukicevic 1992; Rabier et al. 1992; Errico 1997), which has been applied to synoptic scales (e.g., Langland et al. 1995) and convective scales (e.g., Errico et al. 2003). However, the use of adjoint sensitivity for convection is limited due to the difficulty in coding an adjoint model and due to the...
scheme’s implicit assumption of linear error growth, which likely does not accurately represent nonlinear processes (i.e., convection).

In contrast, ensemble-based sensitivity analysis is an alternative technique that applies linear statistics about a set of nonlinear forecast trajectories to evaluate the relationship between forecast variables and initial conditions (Ancell and Hakim 2007; Hakim and Torn 2008). Applying this method to southern Great Plains convection has indicated that ensemble convective forecasts are sensitive to the position of upstream upper-tropospheric potential vorticity (PV) features and lower-tropospheric thermodynamic boundaries that either trigger convection or modify the preconvective environment (Bednarczyk and Ancell 2015; Torn and Romine 2015; Hill et al. 2016). However, in regions where severe weather is climatologically less frequent, such as the northern Great Plains, convection forecasts may be more sensitive to initial condition fields, especially lower-tropospheric moisture or temperature, due to small differences in the position of synoptic features (e.g., Brooks et al. 1993; Crook 1996). In the northern Great Plains, convection often initiates from drylines (e.g., Campbell et al. 2014; Bergmaier and Geerts 2015) and moves in phase with the diurnal cycle (e.g., Easterling and Robinson 1985). Moreover, severe convection in the region usually occurs when an upstream midtropospheric short-wave trough is present (Doswell 1980), which suggests that synoptic features modulate convection and that uncertainty of these features could be responsible for uncertainty in subsequent convection. Overall, convection in the northern Great Plains is not as extensively documented as in the southern Great Plains, which provides an opportunity to evaluate whether convection forecasts in the northern Great Plains show similar forecast sensitivity to upstream synoptic features as in more climatologically favorable regions for convection (i.e., the southern Great Plains).

This study focuses on the role of earlier forecast errors on subsequent convection forecasts by applying ensemble sensitivity analysis to convection that initiated during the evening of 11 June 2013, which corresponded with intensive observing period (IOP) 16 of the Mesoscale Predictability Experiment (MPEX; Weisman et al. 2015). Applying this method to southern Great Plains convection has indicated that ensemble convective forecasts are sensitive to the position of upstream upper-tropospheric potential vorticity (PV) features and lower-tropospheric thermodynamic boundaries that either trigger convection or modify the preconvective environment (Bednarczyk and Ancell 2015; Torn and Romine 2015; Hill et al. 2016). However, in regions where severe weather is climatologically less frequent, such as the northern Great Plains, convection forecasts may be more sensitive to initial condition fields, especially lower-tropospheric moisture or temperature, due to small differences in the position of synoptic features (e.g., Brooks et al. 1993; Crook 1996). In the northern Great Plains, convection often initiates from drylines (e.g., Campbell et al. 2014; Bergmaier and Geerts 2015) and moves in phase with the diurnal cycle (e.g., Easterling and Robinson 1985). Moreover, severe convection in the region usually occurs when an upstream midtropospheric short-wave trough is present (Doswell 1980), which suggests that synoptic features modulate convection and that uncertainty of these features could be responsible for uncertainty in subsequent convection. Overall, convection in the northern Great Plains is not as extensively documented as in the southern Great Plains, which provides an opportunity to evaluate whether convection forecasts in the northern Great Plains show similar forecast sensitivity to upstream synoptic features as in more climatologically favorable regions for convection (i.e., the southern Great Plains).

This case was characterized by the typical synoptic pattern of northern Great Plains severe convection, consisting of a strong dryline and midtropospheric trough over the Rocky Mountains (Bergmaier and Geerts 2015). In addition, this case exhibited significant forecast errors in convective timing and coverage within the 24–36-h real-time MPEX ensemble guidance, which serves as motivation for investigating the dependency of these forecast errors on initial condition uncertainty. In particular, the focus of this study is on the development of two distinct modes of convection occurring within 150 km; therefore, it is possible that the two modes of convection are either sensitive to different features or sensitive to identical features, but via different dynamical mechanisms. As a result, this study evaluates the dynamical mechanisms that contribute to the forecast uncertainty in these two convective modes by comparing the role of synoptic and subsynoptic features on triggering convection or on modifying the preconvective environment.

The remainder of this paper proceeds as follows. Section 2 describes the model system and the application of ensemble sensitivity analysis. Section 3 provides the synoptic overview of the case study, while sections 4 and 5 describe the forecast sensitivity of two modes of convection to earlier forecast fields. Finally, a summary and conclusions are given in section 6.

2. Methodology

a. Model configuration

The 50-member ensemble forecasts are generated using version 3.3.1 of the Advanced Research version of the Weather Research and Forecasting (WRF) Model (Skamarock et al. 2008). The forecasting system for this study is the same as that used in Schwartz et al. (2015) and is summarized in Table 1. The nested computational domain spans the continental United States and adjacent areas [see Fig. 1 of Schwartz et al. (2015)] with a horizontal grid spacing of 15 km in the outer domain and 3 km in the nested domain. Both domains have 40 vertical levels and a 50-hPa top.

b. Ensemble generation and experimental design

Ensemble initial conditions for the 15-km domain are generated via the Data Assimilation Research Testbed software (DART; Anderson et al. 2009). More details regarding the full list of observation types assimilated, along with their observation error sources, are in Table 3 of Romine et al. (2013). The 3-km domain is initialized by downscaling the 15-km analysis and uses the same physics schemes, except that the 3-km domain has the cumulus parameterization turned off. Ensemble lateral boundary conditions for the data assimilation system are generated from NCEP Global Forecast System output every 3 h valid at the appropriate time with perturbations taken from the WRF three-dimensional variational data assimilation (Barker et al. 2012) system via the fixed covariance perturbation technique of Torn et al. (2006). This study focuses on forecasts initialized at 1200 UTC 10 June 2013 (i.e., about 33 h prior to convective initiation); however, a similar analysis was performed on forecasts initialized 12 h (0000 UTC 11 June) and 24 h (1200 UTC 11 June) afterward.
c. Forecast sensitivity

The role of forecast errors at earlier lead times on subsequent convective forecasts is evaluated by using the ensemble-based sensitivity technique (Ancell and Hakim 2007; Torn and Hakim 2008). For an ensemble of size $M$, the sensitivity of a forecast metric $J$ to a model state variable $x_i$ at some earlier lead time is computed via

$$\frac{\partial J}{\partial x_i} = \frac{\text{cov}(J, x_i)}{\text{var}(x_i)},$$

where $J$ and $x_i$ are $1 \times M$ ensemble estimates of the forecast metric and $i$th state variable, respectively; $\text{cov}$ denotes the covariance, and $\text{var}$ denotes the variance. For ease of comparing various fields, the values of $x_i$ are

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FIG. 1. WRF-DART analysis of (left) 0–1-km equivalent potential temperature (shading; K) and 0–1-km winds (vectors; m s$^{-1}$) valid for (a) 1200 UTC 11 Jun and (c) 0000 UTC 12 Jun 2013; and (right) 400-hPa heights (contours; dam), winds (vectors; m s$^{-1}$), and vorticity (shading; 10$^{-5}$ s$^{-1}$) for (b) 1200 UTC 11 Jun and (d) 0000 UTC 12 Jun 2013. The black filled circles and stars represent the locations of the dropsondes obtained during IOP 16 of MPEX used for Fig. 14; the stars represent dropsondes located near the dryline sensitive region. The labels “NE,” “SD,” “WY,” and “CO” represent the U.S. states of Nebraska, South Dakota, Wyoming, and Colorado, respectively.
normalized by its ensemble standard deviation; therefore, all sensitivities will have units of the forecast metric per standard deviation of the state variable. Sensitivity values are considered statistically significant if the absolute value of $\frac{\partial J}{\partial x_i}$ is greater than the 95% confidence interval (e.g., Torn and Hakim 2008). Throughout this paper, $J$ is computed from the 3-km domain, while $x_i$ is taken from the 15-km domain. This choice helps reduce the effect of spurious noise from using a small ensemble to define the linear relationship, while still identifying relevant mesoscale sensitivity fields.

For this study, the forecast metric $J$ is the 1-h maximum vertical kinetic energy $(1/2)w_{\text{max}}^2$, (hereafter referred to as MVKE), averaged over a geographic area. This metric, which is used here as a proxy for convection, is employed because it measures updraft strength, and can help isolate convective precipitation regions. Moreover, averaging over a geographic area ensures the metric measures the

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1 The term $w_{\text{max}}$ denotes the WRF diagnostic variable $W_{\text{UP\_MAX}}$, which is the maximum upward vertical velocity in the column.
extent of convection more so than whether or not convection happens to occur over a small location. The selected metric is similar to the energy norms used in previous predictability studies (e.g., Buizza and Palmer 1995; Barkmeijer et al. 2001; Yamaguchi et al. 2009, 2011; Yamaguchi and Majumdar 2010). For completeness, other forecast metrics related to convection [e.g., accumulated precipitation; Torn and Romine (2015), and column maximum simulated radar reflectivity; Bednarczyk and Ancell (2015)] were evaluated over the same metric region; the results were qualitatively similar (not shown).

d. Forecast composite differencing

In addition to linear regression, the convective forecasts are further analyzed by computing the normalized difference between the 10 members that predict the largest MVKE (hereafter the “strong” members) and the 10 members that predict the lowest MVKE (hereafter the “weak” members) via

$$\Delta x_i = \frac{x_{\text{strong}}^i - x_{\text{weak}}^i}{\sigma_{x_i}},$$

where $x_{\text{strong}}^i$ ($x_{\text{weak}}^i$) represents the mean of the $i$th state variable for the strong (weak) ensemble members and $\sigma_{x_i}$ represents the ensemble standard deviation of $x_i$ computed from all members. The statistical significance of the composite differences is assessed by applying bootstrap resampling without replacement, as described in Torn and Romine (2015).

3. Case overview

The 12 June 2013 northern Great Plains severe convective outbreak exhibited conditions typical of previous severe weather events over the region (e.g., Doswell 1980; Klimowski et al. 2003; Campbell et al. 2014). At 1200 UTC 11 June, a surface cyclone over northeastern Colorado was associated with broad 0–1-km southerly winds over Kansas and Nebraska, which advected higher equivalent potential temperature ($\theta_e$) air poleward, forming a $\theta_e$ ridge over Nebraska and southern South Dakota (Fig. 1a). At the western boundary of this $\theta_e$ ridge near eastern Wyoming is a sharp gradient indicative of a dryline, which divides drier air and westerly winds to the west from moister air and easterly winds to the east. Meanwhile, along the northeastern boundary of this $\theta_e$ ridge is a southwestward-moving cold front present over northern South Dakota, which separates relatively warm and moist air to the south from cooler and drier air to the north. The area within the $\theta_e$ ridge was characterized by up to 900 J kg$^{-1}$ of mixed-layer convective available potential energy (MLCAPE), but was strongly capped given the 600–850 J kg$^{-1}$ of mixed-layer convective inhibition (MLCIN) as indicated by WRF-DART analyses at this time (not shown). Simultaneously, the 400-hPa synoptic-scale pattern was characterized by broad southwesterly flow over the Intermountain West ahead of a long-wave trough off the coast of Washington, and a long-wave ridge over the Great Plains (Fig. 1b). Embedded within this southwesterly flow is a robust short-wave trough, characterized by relative vorticity exceeding $2.0 \times 10^{-4}$ s$^{-1}$ located over Utah, which originated in the eastern Pacific days before.

By 0000 UTC 12 June, the dryline moved east into northwestern Nebraska due to the increased westerly flow downslope of the Rocky Mountains (Fig. 1c), induced by the approaching short-wave trough, which is located over eastern Wyoming at this time (Fig. 1d). In
Advance of the dryline, daytime heating increased the boundary layer temperature, which helped the MLCIN decrease to \(<50\, \text{J kg}^{-1}\), while the MLCAPE increased to \(>2800\, \text{J kg}^{-1}\) over northwestern Nebraska. At the same time, the cold front remained in central South Dakota. To the south of the cold front in western South Dakota, the MLCIN decreased to \(<50\, \text{J kg}^{-1}\) while the MLCAPE increased to \(>2000\, \text{J kg}^{-1}\) (WRF-DART analysis; not shown). The remaining capping inversion in these regions suggests additional forcing was necessary for convective initiation to occur. The additional forcing could have resulted in part from the near-surface confluence along the dryline and cold front, and from the synoptic forcing for ascent ahead of the midtropospheric trough; therefore, forecast errors associated with these features could impact the subsequent prediction of convection.

At 0100 UTC 12 June, convection initiated along the dryline near the Wyoming–Nebraska border and also farther north in eastern Wyoming, such that two hours later the convection in Nebraska split into more discrete supercells, while the convection that originated in northeastern Wyoming moved into western South Dakota.
Dakota and evolved into a convective line (Fig. 2a). At 0500 UTC, the northernmost supercell in Nebraska moved into southern South Dakota, while the convective line in South Dakota shifted eastward and extended farther along the north–south direction (Fig. 2b). Ultimately by 0700 UTC, the convection in South Dakota associated with the convective line and supercell evolved into a larger mesoscale convective system located in central South Dakota (Fig. 2c). Overall, the two initially distinct convective modes in Nebraska and South Dakota resulted in two tornadoes in Nebraska, and numerous wind and hail reports in South Dakota (National Climatic Data Center 2016).

During MPEX, 24–36-h real-time WRF forecasts of these two convective modes were characterized by significant forecast errors; Fig. 2 compares the time evolution of the observed reflectivity and the simulated reflectivity for two representative ensemble members. Overall, all ensemble forecasts are consistently 4 h too early with simulating convection compared to the observed reflectivity, which suggests model error. One potential error source could be the Mellor–Yamada–Janjic’ (MYJ) planetary boundary layer scheme, which tends to produce near-surface conditions that are too cool and moist while also underpredicting the MLCIN when a substantial capping inversion is present (e.g., Coniglio et al. 2013). Given the strong capping inversion in this case, the MYJ biases could have produced lower lifting condensation levels (LCLs) and MLCIN in the model compared to the observations, which would make it easier for convective initiation to occur earlier. However, among the ensemble there is large variability in the extent of convection over northwestern Nebraska and western South Dakota. In general, the simulated reflectivity shows a north–south-oriented line of convection from western South Dakota into northern Nebraska (Figs. 2d,e), while some members have little convection in northern Nebraska (Figs. 2g,h). By contrast, all members have a northwest–southeast-oriented line of convection in western South Dakota, but there are differences in the east–west position of the line or whether the convection is a continuous line, or broken segments, which amplify over time (Figs. 2f,i). Figure 2 also indicates potential forecast failure over southern Nebraska and northern Kansas; however, analysis of this region is beyond the scope of this work.

Given the range of predicted convection within the ensemble, this study and the subsequent sensitivity analysis focuses on the processes that modulate the

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2 Northeast Wyoming and southeast Montana are not sampled well by radars due to the large distances between neighboring NEXRAD WSR-88D stations and mountain blockage (e.g., Salazar et al. 2009).
forecast differences in MVKE over the black and red boxes of Figs. 2d and 2g [hereafter denoted the Nebraska MVKE (NEKE) and the South Dakota MVKE (SDKE), respectively]. These boxes enclose the largest ensemble standard deviation of 1-h MVKE during the time period of 34–36 h, shortly after convective initiation (33 h; 2100 UTC 11 June). The variability in NEKE and SDKE can be illustrated by the time series of hourly MVKE for each ensemble member (Figs. 3a,b). As suggested by Fig. 3a, there are large differences in the timing and magnitude of NEKE, particularly near convective initiation. Given the desire to focus on a time period when most of the members have convection, but not before cold pool dynamics become the main forcing mechanism, this study focuses on NEKE between 2300 UTC 11 June (35 h) and 0000 UTC 12 June (36 h). Shifting the metric boxes east–west or increasing the time window yields qualitatively similar sensitivity results [not shown, but similar as in Bednarczyk and Ancell (2015)]. Comparatively, the time series of SDKE shows primary variability in the timing of convection between 2200 (34 h) and 2300 UTC 11 June (35 h; Fig. 3b). In essence, the NEKE and SDKE metrics represent the areal extent of convective updrafts within the specified region over the specified period, focusing on regions with large ensemble standard deviation in updraft strength (i.e., 1-h MVKE). Repeating the sensitivity calculation for largest gridpoint vertical kinetic energy in both boxes also yields similar results (not shown). For the NEKE metric, the members that appear to produce a more accurate convective forecast are those that have less convection (i.e., weak members), regardless of the universal timing error. By contrast, the combination of timing errors, simulated convective line differences, and insufficient radar coverage make it difficult to say which members are more accurate for SDKE. Therefore, the SDKE metric is presented to identify the meso- to synoptic-scale features to which convection within the SDKE box is sensitive.

4. NEKE forecast sensitivities

One hypothesis for the differences in convection within these two regions is that the preconvective thermodynamic environment is more favorable for convective initiation in the strong members. This hypothesis is evaluated for NEKE by comparing the thermodynamic profiles prior to convective initiation for the members

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**Fig. 6.** Sensitivity of the 35–36-h area-averaged vertical kinetic energy over the black box shown in Fig. 2 to the (a) 12-, (b) 18-, (c) 24-, and (d) 30-h forecast of 0–1-km equivalent potential temperature (shading; m² s⁻²) initialized at 1200 UTC 10 Jun 2013. Black stippled regions indicate where the sensitivity is statistically significant at the 95% confidence level. Contours are the ensemble-mean equivalent potential temperature (K). The thick black line in (d) represents the location for the vertical cross section in Figs. 8 and 9b.
that produce more robust convection versus those that do not using Eq. (2) (Fig. 4; at 1800 UTC 11 June). Between 700 and 850 hPa, the strong members are characterized by up to 1.3 standard deviation (3.3 K) greater $u_e$ (Fig. 4a). In this layer, the $u_e$ difference is principally due to water vapor differences, as the strong members have up to 1.2 standard deviation (1.6 g kg$^{-1}$) more water vapor mixing ratio (Fig. 4b), but also have up to 0.9 standard deviation (1.4 K) lower temperatures than the weak members (Fig. 4c). The more favorable environment for convection is statistically significant starting at 1500 UTC as the strong members are characterized by statistically higher 0–1-km water vapor mixing ratio (Fig. 5a) and, by 1800 UTC, statistically higher 0–1-km $u_e$ (Fig. 5b), which result in lower surface-based LCLs (i.e., higher LCL pressures; Fig. 5c), thus making it easier for convective initiation to occur. Additionally, at 1800 UTC the strong members are characterized by >1.3 standard deviation (0.015 m s$^{-1}$) greater vertical velocity between the surface and 500 hPa (Fig. 4d), which is associated with up to 1.3 standard deviation greater lower-tropospheric convergence ahead of the ensemble-mean dryline (not shown). The increased upward vertical motion characterizing the strong members would be expected to cool and moisten at the top of the boundary layer, partially reducing the capping inversion over the region. Overall, the pre-convective lower-tropospheric moisture and vertical velocity differences could have been modulated by the advection of moisture or dynamical forcing for ascent associated with a synoptic or mesoscale feature, such as the aforementioned upstream midtropospheric trough or dryline; therefore, this hypothesis is tested by computing the sensitivity of NEKE and the preconvective environment to the position and amplitude of these mesoscale and synoptic features.

The lower-tropospheric $u_e$ differences suggest that NEKE may be particularly sensitive to lower-tropospheric boundaries (e.g., Figs. 1a,c); therefore, Fig. 6 shows the sensitivity of NEKE to the 0–1-km $u_e$ at various lead times. While there are small incoherent sensitive regions over Wyoming, South Dakota, and Nebraska at 12 h (0000 UTC 11 June; Fig. 6a) and 18 h (0600 UTC; Fig. 6b), by 24 h (1200 UTC) the largest sensitivity values to $u_e$ are located along the western and eastern sides of the amplifying $u_e$ ridge in Wyoming and South Dakota, respectively (Fig. 6c). The western (eastern) side of this $u_e$ ridge, which denotes the position of the dryline (cold front), is characterized by positive (negative) sensitivity values, suggesting that increasing the $u_e$ on the west side...
or decreasing the $\theta_e$ on the east side of the $\theta_e$ axis is associated with larger NEKE. The combination of these two sensitive regions suggests that shifting the $\theta_e$ axis to the southwest at this lead time is associated with an increase in convection over the NEKE box later that day (6.4 m$^2$ s$^{-2}$ per standard deviation change in $\theta_e$). In addition, the sensitivity of NEKE to lower-tropospheric zonal and meridional wind is consistent with the position variability of the $\theta_e$ gradients (Fig. 7). For example, there is a large region of negative sensitivity to zonal wind over
eastern Wyoming and western Nebraska behind the dryline (denoted by the ensemble-mean wind shift) and an elongated region of positive sensitivity to meridional wind in eastern Wyoming along the dryline, both suggesting that shifting the dryline to the west, which in turn should result in more southeasterly winds over eastern Wyoming, is associated with increased NEKE (>8.0 m$^2$s$^{-2}$ per standard deviation; Figs. 7a,c). The sensitivity of NEKE to the dryline position is also supported by the wind shift from easterly to westerly winds for the weak members, suggesting that the dryline passed this location at 1500 UTC (Fig. 5d). For the 30-h forecasts (1800 UTC), the sensitive regions remain along the boundaries, as the dryline is now located along the Wyoming–Nebraska border to the west of NEKE and the cold front is positioned in central South Dakota (Figs. 6d). Furthermore, the zonal and meridional wind sensitivity patterns move eastward following the movement of the dryline. Meanwhile, there exists a positive (negative) sensitivity to the zonal (meridional) winds on the east side of a surface anticyclone over Minnesota, which indicates that increasing the winds on this side of the anticyclone, which in turn might be expected to shift the cold front southward, is associated with increased NEKE (Figs. 7b,d). Overall, these results indicate sensitivity to the position of both upstream and downstream features.

The sensitivity of NEKE to the position of the dryline is further supported by taking a vertical cross section along the west–east black line indicated in Fig. 6d and computing the sensitivity of NEKE to $\theta_e$, meridional wind, and zonal wind at various pressure levels. At 1200 UTC (24 h), NEKE is most sensitive to lower-tropospheric fields near the dryline, located at 42.5°N, 105.4°W, with positive sensitivity to $u_e$ (Fig. 8a) and meridional wind (Fig. 8c), while negative sensitivity to zonal wind (Fig. 8e), indicating that shifting the dryline to the west is associated with increased convection. By 1800 UTC, the ensemble-mean dryline moves eastward to 42.6°N, 103.0°W, yet the NEKE sensitivities remain large along the dryline (Figs. 8b,d,f). In addition, the sensitivity to midtropospheric zonal wind, meridional wind, and $\theta_e$, which is associated with the approaching short-wave trough, becomes as large as the aforementioned lower-tropospheric sensitivity. Moreover, the trough is also characterized by

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**Fig. 7.** First principal component of the 1200 UTC 11 Jun 330–340-K PV field over the upstream trough regressed onto (a) the 1200 UTC 11 Jun 330–340-K PV field over the trough (shading; PVU, where 1 PVU = $10^{-6}$ K kg$^{-1}$ m$^2$s$^{-1}$), (b) the 1200 UTC 11 Jun equivalent potential temperature (shading; K) along the cross section indicated in Fig. 6d, and (c) the 1800 UTC 11 Jun vertical profile of water vapor advection averaged over NEKE (red contour; g kg$^{-1}$ day$^{-1}$). The solid black contours in each panel represent the respective ensemble-mean values. Black stippled regions in (b) and the gray shading in (c) indicate where the regression is statistically significant at the 95% confidence level.
positive (negative) sensitivity in the meridional (zonal) wind, which suggests that shifting the midtropospheric trough to the west is associated with increased convection, similar to the lower-tropospheric dryline sensitivity.

Given that NEKE is greater when both the trough and dryline are located farther to the west, it is possible that the positions of these two features are related to one another. However, the amorphous shape of troughs and drylines makes it difficult to define their position through area-averaged quantities and thus quantify their relationship. One way to describe the position variability of these features and hence quantify the position covariability is to compute the principal component (PC) of empirical orthogonal functions (EOFs) of fields that describe these features, similar to Chang et al. (2013). Here the position of the trough is quantified through the PC of the first EOF of the 1200 UTC 330–340-K Ertel PV field surrounding the trough (Fig. 9a).3 The 330–340-K layer is selected because the trough’s PV anomaly is maximized within this layer. The EOF pattern is characterized by a positive–negative dipole centered on the trough itself, which indicates the southwest–northeast position variability of the trough, such that positive (negative) PCs denote a more southwest (northeast) trough. The relationship between the position variability of the trough and the dryline is now evaluated by regressing the PC of this EOF pattern onto a vertical cross section (as indicated in Fig. 6d) of the 1200 UTC $u_e$ field perpendicular to the dryline position (Fig. 9b). The large, statistically significant, positive regression values near $42.5^\circ N, 105.4^\circ W$, which is along the dryline, denote that a more southwest (northeast) trough location implies higher $u_e$ or a more western (eastern) dryline. One possible dynamical mechanism linking the trough and dryline is that the horizontal advection of dry air and subsequent vertical mixing associated with the trough allow the dry air to mix down to the surface (e.g., Hane et al. 2001), which can increase the amount of vertical mixing in the boundary layer and propel the dryline east for larger distances than in quiescent conditions (e.g., McCarthy and Koch 1982; Hane et al. 1993). Another way to support the connection between the trough and the dryline is to assess the trough’s role in the low-level moisture advection over NEKE; this

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3 The EOF patterns shown in this study are computed by subtracting grid points of the field for each ensemble member from the ensemble mean and normalizing by the standard deviation, and then regressing the first PC onto the respective field; resulting in EOF patterns with the same units as the respective field.
is tested by regressing the first PC of the 1200 UTC 11 June trough position onto a vertical profile of the 1800 UTC water vapor advection over NEKE (Fig. 9c). The positive statistically significant regression values over the lower troposphere (e.g., 800 hPa to the surface) suggest that a more westward trough is associated with increased water vapor advection over NEKE; in essence implying that the trough modulates the moisture content over NEKE by controlling water vapor advection (i.e., the dryline position) into NEKE. Overall, these results suggest that the position variability of the upstream trough can explain the position variability of the dryline.

The hypothesized connection between the upstream trough and the dryline suggests that the errors in the dryline position are related to variability in the trough position. Therefore, one possible explanation for the lack of coherent sensitivity of NEKE to the near-surface \( \theta_e \) prior to 1200 UTC 11 June is that the forecast errors in this case originate from errors in the trough’s position, which prior to 1200 UTC is not close enough to influence the dryline’s position. This possibility is evaluated by computing the sensitivity of NEKE to the 330–340-K PV at earlier times (Fig. 10). For the 6-h forecast (1800 UTC 10 June) the midtropospheric trough, identified as the ensemble-mean PV maximum over California, is characterized by a broad positive (negative) sensitivity on its western (eastern) side (Fig. 10a), which suggests that shifting the trough to the west (i.e., delaying the arrival of the trough) would be associated with increased convection over the NEKE box later on. Moreover, later lead times exhibit a more focused concentration of positive (negative) sensitivity centered on the trough as it moves to the northeast, translating over the California–Nevada border by 12 h (Fig. 10b), over central Nevada by 18 h (Fig. 10c), and over Utah by 24 h (Fig. 10d). One way to demonstrate the consistency of the sensitivity of NEKE to the trough’s position is a Hovmöller diagram of the sensitivity to the 330–340-K PV computed along the subtropical waveguide using the Martius et al. (2006) method (Fig. 11). The sensitivity dipole associated with the trough is easily traced backward in time to uncertainty in the 0-h trough position just off the California coast. The only time when the sensitivity to the trough position is not statistically significant is between 12 and 15 h when the trough crosses the Sierra Nevada, which likely contaminates the sensitivity. Nonetheless, Fig. 11 implies that that the NEKE forecasts are highly sensitive to the 0-h position errors associated with this trough.

Recall that the preconvective environment for the strong and weak members is characterized by large differences in the near-surface \( \theta_e \) (cf. Fig. 4a), which suggests that this quantity modulates NEKE. As a consequence, the importance of the trough and dryline position on convection forecasts can be further validated by computing the sensitivity of the 0–1-km \( \theta_e \) averaged over the NEKE box just prior to convective initiation (i.e., 2100 UTC 11 June) to the 0–1-km \( \theta_e \) and 330–340-K PV at earlier times (Fig. 12). If the convection is modulated by the trough and dryline position then the near-surface \( \theta_e \) averaged over NEKE should be sensitive to the trough and dryline in similar locations. Indeed, the 2100 UTC 0–1-km \( \theta_e \) is sensitive to the 1200 UTC (24h) positions of both the dryline (Fig. 12a) and the trough (Fig. 12b) in almost identical locations as the sensitivity of NEKE to these features; positive sensitivity on the west side of both the trough over Utah and the \( \theta_e \) axis over Wyoming and negative sensitivity on the east side of both the trough over Wyoming and the \( \theta_e \) axis over South Dakota, which suggests that shifting the trough or \( \theta_e \) axis (i.e., dryline) to the west is associated with greater 0–1-km \( \theta_e \) in the NEKE box by 2100 UTC. As a consequence, it appears that the same processes that modulate NEKE also modulate the lower-tropospheric thermodynamic state prior to convective initiation.

The sensitivity of NEKE to the midtropospheric trough and dryline is not unique to the chosen initialization time; ensemble forecasts initialized closer to the event exhibit significant sensitivity to the same
features. Figure 13 shows the sensitivity of NEKE\textsuperscript{4} to the 0–1-km $\theta_e$, 0–1-km zonal wind, and 330–340-K PV at 1800 UTC 11 June initialized on 0000 UTC 11 June (left column) and 1200 UTC 11 June (right column). For both initialization times, the NEKE forecasts exhibit positive sensitivity along the dryline in eastern Wyoming (Figs. 13a,b) and negative sensitivity to the zonal wind west of the dryline (Figs. 13c,d), suggesting that shifting the dryline westward would lead to increased convection over the NEKE box, though the sensitive area shrinks for the 1200 UTC initialization compared to the 0000 UTC initialization. In addition, a similar west–east positive–negative dipole sensitivity to the trough is evident in the 0000 UTC initialization (Fig. 13e); however, there is negligible sensitivity to the trough in the 1200 UTC initialization (Fig. 13f). The decreased sensitivity to the trough’s position for the 1200 UTC 11 June initialization suggests that errors in the position of this trough have relatively little impact on the lower troposphere for initialization times closer to the event.

In addition to the sensitivity results, another way to quantify the importance of lower-tropospheric forecast errors along the dryline around 1200 UTC on NEKE forecasts 12 h later is to compute the error in each ensemble member’s forecast with respect to dropsonde profiles collected by the G-V aircraft during MPEX IOP 16. There were a total of 33 dropsondes collected during this IOP (denoted “All Drops.”; filled circles and stars in Fig. 1), which included 11 profiles that sampled the sensitive area in eastern Wyoming and western Nebraska between 0928 and 1055 UTC (denoted “Sens. Drops.”; filled stars in Fig. 1). Recall from section 3 that the convection within NEKE is too extensive in the strong members compared to the weak members. In turn, if the NEKE forecast is sensitive to thermodynamic errors in Wyoming and far western Nebraska around 1200 UTC, then the ensemble members with smaller thermodynamic errors in this region should also have less extensive convection (i.e., the weak members). Figure 14 shows the relative improvement, which is defined as the difference between the mean absolute error (MAE) averaged over all of the strong members and the MAE averaged over all of the weak members, divided by the MAE in the ensemble mean, computed with respect to the Sens. Drops and to All Drops.\textsuperscript{5} Here, relative improvement is used to provide a fair comparison between the sensitive and all dropsonde sets, which have different number of observations and error amplitudes. For temperature (Fig. 14a), the MAE for the weak members is 10%–25% lower than the strong members for dropsondes in the sensitive region between 575 and 800 hPa, which is statistically significant at the 95% confidence level.\textsuperscript{6} Recall that the bottom part of this layer corresponds to where the strong members have statistically higher $\theta_e$ prior to convective initiation (cf. Fig. 4a). By contrast, the MAE difference

\textsuperscript{4} NEKE is computed independently for each ensemble initialization.

\textsuperscript{5} The forecast values are obtained by spatially and temporally interpolating the model grids to the location and time when the dropsonde profile was taken.

\textsuperscript{6} Determined using a bootstrap resampling procedure similar to what is used to establish statistical significance in Eq. (2).
between strong and weak members is not statistically significant with respect to all dropsondes. Similar to temperature, the weak members are characterized by generally lower MAE in specific humidity throughout the troposphere, particularly between 625 and 775 hPa for the sensitive dropsondes (Fig. 14b), while the differences with respect to all dropsondes is smaller. Moreover, the zonal and meridional wind error differences are statistically insignificant (not shown). Nevertheless, this result strongly suggests that ensemble members with less convection over the NEKE box (i.e., more accurate convection forecasts compared to the observed event) are obtained by having a better representation of the upstream environment within the sensitive region.

5. SDKE forecast sensitivities

The role of synoptic and mesoscale features in modulating the convection in Nebraska is now compared with the role of these features in modulating the more linear convection in South Dakota (cf. Figs. 2 and 3). The South Dakota convection forecast variability may be explained by comparing the thermodynamic profile over the SDKE box between the strong and weak members\(^7\) prior to convective initiation (Fig. 15; 2100 UTC). For

\(^7\) Note that the strong and weak members associated with the SDKE box are different from that of the NEKE box.
this metric, the strong members are characterized by 1.5 standard deviation (6.12 K) greater use between the surface and 850 hPa compared to the weak members (Fig. 15a). Within this layer, the strong members have both higher (standardized difference) water vapor mixing ratio (Fig. 15b) and potential temperature (Fig. 15c). As a result, the strong members are characterized by 343 J kg\(^{-1}\) more MLCAPE (not shown). Additionally, the strong members are characterized by up to 2.0 standard deviation (0.05 m s\(^{-1}\)) larger vertical velocity throughout a deep layer from 250 to 850 hPa (Fig. 15d). Unlike the NEKE box, this metric area is closer to the path of the midtropospheric trough, which is likely to produce deeper forcing for ascent over the area. The forcing for ascent, especially in the lower troposphere, would be expected to reduce the capping inversion over South Dakota, such that the strong members are characterized by 30 J kg\(^{-1}\) less MLCIN and 40 hPa higher LCL pressures (i.e., LCLs closer to the ground; not shown). Overall, the importance of both lower-tropospheric vertical motion and use on SDKE forecasts is evaluated by computing a multivariable regression analysis with the 30-h 700–850-hPa vertical motion and 0–1-km use as predictors and SDKE as the predictand. The 0.435 correlation coefficient between the predictors and predictand, in combination with the 0.114 correlation coefficient computed between the predictors, confirms that these two predictors explain a significant amount of variance, yet are relatively independent of one another. As a result of these calculations and the environmental profile differences, this subsection will focus on the sensitivity of SDKE to earlier forecast times and relate these sensitivities to lower-tropospheric use and vertical velocity within this region.

Similar to NEKE, the position of the use axis is hypothesized to modulate the preconvective lower-tropospheric use differences. In support of this hypothesis, Fig. 16a indicates that the SDKE forecast is sensitive to the 24-h 0–1-km use along the western (eastern) boundary of the use ridge in Wyoming (South Dakota), denoting the position of the dryline (cold front). Moreover, the use ridge is straddled from west to east by a negative–positive dipole sensitivity, reversed of the use ridge’s dipole signature for the NEKE box, which indicates that shifting the use axis to the east is associated with an increase in convection over the SDKE box later that day. In support of the position variability of the use axis, SDKE is sensitive to the 0–1-km zonal (meridional) wind maximized over eastern Wyoming on the west side of the use axis, which also indicates that shifting the dryline eastward is associated with increased SDKE (4.8 m\(^2\) s\(^{-2}\) per standard deviation: Figs. 16c,e). In addition, SDKE exhibits a negative (positive) sensitivity to zonal (meridional) wind on the east side of a surface anticyclone in North Dakota, which suggests that decreasing the winds on the east side of the anticyclone yields an eastward shift of the use axis (i.e., cold front). Moreover, these same patterns are present six hours later, following the southwestward and eastward movement of the cold front and dryline, respectively (Figs. 16b, 16d, and 16f). At this time, the cold front is near the edge of the SDKE box, which suggests that the cold front
modulates the lower-tropospheric moisture and potential temperature over SDKE, such that shifting the cold front to the northeast would result in more moisture and warmer temperatures, potentially making the SDKE region more supportive of convection.

The SDKE forecasts are also characterized by sensitivity to the location of mid- to upper-tropospheric features, as shown in the 330–340-K PV field (Fig. 17). In comparison to NEKE, which is characterized by the largest sensitivity to the upstream trough, SDKE exhibits less sensitivity to this feature. Moreover, from 12 to 24 h the upstream trough is straddled from west to east by a negative–positive dipole sensitivity (Figs. 17b–d), reversed of the trough’s dipole signature for the NEKE box, which indicates that shifting this trough to the east is associated with an increase in SDKE. In contrast, SDKE is highly sensitive to the structure of the northern waveguide (i.e., PV gradient). Beginning at 6 h, a region of negative sensitivity (up to $4.8 \text{ m}^2\text{s}^{-2}$ per standard deviation) is present along the midlatitude waveguide extending from northern Washington to Minnesota, which suggests that shifting the jet to the north would be associated with greater convection in the SDKE box (Fig. 17a). Over subsequent times, the sensitive region moves eastward to the west side of a long-wave trough over Minnesota and Wisconsin (e.g.,

![Fig. 15](image-url)
Figs. 17b–d), which indicates that shifting the trough to the north is associated with an increase in SDKE (>6.0 m$^2$s$^{-2}$ per standard deviation). Overall, these results suggest sensitivity to the downstream forecast state, such that shifting the northern waveguide more poleward is associated with increased convection over the metric box in South Dakota.

The sensitivity of SDKE to the position of both an upstream and downstream trough can be explained by the influence of each feature on the preconvective environment. The upstream trough is hypothesized to modulate the forcing for vertical motion (cf. Fig. 15d), which can modulate the SDKE-region LCL and cap strength, and hence convective initiation. Therefore, the relationship between the upstream trough and lower-tropospheric vertical motion is assessed by computing the sensitivity of the 2100 UTC 700–850-hPa vertical velocity over the SDKE box to the 330–340-K PV at earlier forecast times (Figs. 18a,b). At 24 h, the vertical motion is sensitive to the position of the upstream trough, as evidenced by the negative–positive dipole straddling the trough over Idaho, but also to the structure of the broad PV anomaly as shown by the positive sensitive region over Colorado and Wyoming and negative region over California and Nevada (Fig. 18a). This pattern, which persists to 30 h (Fig. 18b), indicates that shifting this trough to the northeast is associated with greater lower-tropospheric vertical motion. Similarly,
the downstream trough is also straddled by a negative–positive dipole sensitivity; however, this trough remains far from the SDKE region and thus likely does not directly influence the lower-tropospheric vertical motion over the SDKE region. Instead, the downstream trough’s dipole sensitivity pattern likely reflects that members with large SDKE have a more progressive synoptic pattern. Moreover, the downstream trough is hypothesized to modulate the lower-tropospheric $u_e$ over SDKE; therefore, the sensitivity of the 2100 UTC 0–1-km $u_e$ over SDKE to the 330–340-K PV field at 24 h (Fig. 18c) and 30 h (Fig. 18d) is computed. These two lead times reveal that the preconvective lower-tropospheric $u_e$ is sensitive to the downstream trough, but not the upstream trough, and indicate that shifting the downstream trough to the east is associated with greater $u_e$ over SDKE. As a consequence, it appears that the downstream trough modulates the preconvective $u_e$ over the SDKE box, while the upstream trough modulates the preconvective vertical motion.

One possible explanation for how errors associated with the downstream PV gradient are connected to SDKE is that the position of the cold front in South Dakota is modulated by the winds associated with the PV differences in the sensitive region, via “action at a distance” (e.g., Hoskins et al. 1985). This hypothesis can be tested by applying statistical PV inversion (e.g., Hakim and Torn 2008) to the difference in 330–340-K PV between the strong and weak members in order to derive the perturbation lower-tropospheric winds and $\theta_e$, similar to what is done in Torn and Romine (2015). Figure 19 shows the 24-h 330–340-K PV anomaly, defined as the difference between the strong and weak members, and the domain over which the inversion is applied, along with the inverted 850-hPa winds and $\theta_e$. A negative PV anomaly such as this should be associated with anticyclonic perturbation winds. To the southwest of this PV anomaly, the inverted winds are southerly and normal to the cold front extending from North Dakota into eastern South Dakota, which also coincides with positive perturbation $\theta_e$ (Fig. 19), suggesting that the negative PV anomaly is associated with a northern shift of the cold front. Therefore, this result further suggests that the errors associated with the position of the downstream PV anomaly modulate the position of the cold front and lower-tropospheric $\theta_e$ over South Dakota.

6. Summary and conclusions

This study assesses the role of variability in the location of upstream and downstream features on ensuing forecasts of severe convection over the northern Great

Fig. 17. Sensitivity of the 34–35-h area-averaged vertical kinetic energy over the red metric box in Fig. 2 to the (a) 6-, (b) 12-, (c) 18-, and (d) 24-h 330–340-K PV (shading; m$^2$ s$^{-2}$) initialized at 1200 UTC 10 Jun 2013. The contours are the ensemble-mean 330–340-K PV at the valid time (PVU). Black stippled regions indicate where the sensitivity is statistically significant at the 95% confidence level.
Plains on 11–12 June 2013. This case study utilized a 50-member ensemble of convection-permitting WRF forecasts, which were characterized by large spread in convection over Nebraska and South Dakota. The relationship between convection over two nearby regions and earlier forecast fields is computed using the sensitivity of area-averaged maximum vertical kinetic energy (a proxy for convection) to forecast fields at earlier times using the ensemble-based sensitivity technique. Moreover, sensitivity analysis is employed to analyze the mechanisms influencing some members to generate erroneous convection compared to the observed event.

The forecast sensitivities suggest that the convection in both regions is modulated by the positions of lower-tropospheric thermodynamic boundaries at earlier lead times, which in turn modulate the preconvective lower-tropospheric $\theta_e$; such that aligning the $\theta_e$ axis with the metric region gives rise to more convection. Over Nebraska, the convection is sensitive to the position of the dryline in eastern Wyoming, such that the ensemble members that produce more convection in Nebraska are associated with a more western dryline than the members that produce less convection. The more western dryline is coincident with stronger easterly and southerly flow ahead of the dryline, leading to higher moisture advection in Nebraska and a more favorable thermodynamic environment for convective initiation at the diurnal maximum. Similarly, the convection in South Dakota is sensitive to the position of the cold front in central South Dakota and the aforementioned dryline, such that the members that produce more convection are associated with a more northern cold front and eastern dryline, which allows for greater $\theta_e$ in the region before convective initiation.

The positions of the dryline and cold front are modulated by mid- to upper-tropospheric synoptic features both upstream and downstream of the convection. The position of the dryline is modulated by an upstream midtropospheric trough, such that a more western trough is associated with a more western dryline and subsequently greater convection in Nebraska. By contrast, the position of the cold front is modulated by a downstream upper-tropospheric trough over Minnesota, such that a more northern trough results in an anticyclonic...
circulation perturbation in the lower troposphere that is associated with a more northern cold front and subsequently greater convection in South Dakota. Overall, the role of upstream PV errors in subsequent convection forecasts over the northern Great Plains is consistent with previous studies, particularly over the convectively active southern Great Plains (e.g., Roebber et al. 2002; Hanley et al. 2013; Torn and Romine 2015); however, this study also highlights the importance of downstream PV features.

This study emphasizes that the role of PV features on convection forecasts is not limited to the convective triggering location; rather they can modify the preconvective thermodynamic environment by altering the location of synoptic forcing for ascent or by modulating the position of lower-tropospheric boundaries near the region of convective initiation. Downstream PV features in particular are more likely to impact the preconvective environment when either they pass over the preconvective region or when they modulate nearby lower-tropospheric thermodynamic boundaries that approach the preconvective region. As indicated by the dropsonde comparisons, better representation of features impacting the preconvective thermodynamic environment can produce better convection forecasts, and ensemble-based sensitivity analysis appears to be able to identify sensitive regions for sampling.

Acknowledgments. We would like to acknowledge high-performance computing support from Yellowstone (ark:/85065/d7wd3xhc) provided by NCAR’s Computational and Information Systems Laboratory, sponsored by the National Science Foundation. We also thank three anonymous reviewers for their insightful comments on an earlier version of this manuscript. This work was supported by National Science Foundation Grant 1239787 and NOAA Grant NA14OAR4830172.

REFERENCES
Barkmeijer, J., R. Buizza, T. N. Palmer, K. Puri, and J. F. Mahlouf, 2001: Tropical singular vectors computed with linearized...