Effects of Surface Heat and Moisture Exchange on ARW-WRF Warm-Season Precipitation Forecasts over the Central United States

S. B. Trier, M. A. Lemone, F. Chen, and K. W. Manning

National Center for Atmospheric Research, Boulder, Colorado

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ABSTRACT

The evolution of the daytime planetary boundary layer (PBL) and its association with warm-season precipitation is strongly impacted by land–atmosphere heat and moisture exchange (hereafter surface exchange). However, substantial uncertainty exists in the parameterization of the surface exchange in numerical weather prediction (NWP) models. In the current study, the authors examine 0–24-h convection-permitting forecasts with different surface exchange strengths for a 6-day period during the International H2O Project (IHOP_2002). Results indicate sensitivity in the timing of simulated afternoon convection initiation and subsequent precipitation amounts to variations in surface exchange strength. Convection initiation in simulations with weak surface exchange was delayed by 2–3 h compared to simulations with strong surface exchange, and area-averaged total precipitation amounts were less by up to a factor of 2. Over the western high plains (105°–100°W longitude), where deep convection is locally generated, simulations using a formulation for surface exchange that varied with the vegetation category (height) produced area-averaged diurnal cycles of forecasted precipitation amounts in better agreement with observations than simulations that used the current Advanced Research Weather Research and Forecasting Model (ARW-WRF) formulation. Parcel theory is used to diagnose mechanisms by which differences in surface exchange influence convection initiation in individual case studies. The more rapid initiation in simulations with strong surface exchange results from a more rapid removal of negative buoyancy beneath the level of free convection, which arises primarily from greater PBL warming.

1. Introduction

Land surface conditions including soil moisture and green vegetation fraction can impact deep convective precipitation (e.g., Pielke 2001). This results from their effect on the daytime sensible and latent heat fluxes, which influences local conditional instability (e.g., Betts and Ball 1995; James et al. 2009) and mesoscale circulations arising from surface heterogeneity (e.g., Pielke and Segal 1986; Lanicci et al. 1987; Segal and Arritt 1992).

Recent simulations (e.g., Trier et al. 2004; Holt et al. 2006) with numerical weather prediction (NWP) models have found sensitivities in convection initiation (CI) and quantitative precipitation forecasts (QPFs) related to these effects of the land surface. However, a major source of uncertainty is the strength of the bulk aerodynamic coefficients for heat and moisture calculated in surface layer parameterizations of such models (e.g., Chen et al. 1997). In this study, we examine the role of the related surface exchange strength on convection initiation and short-range (e.g., 0–24 h) QPFs.

Overall effects of land–atmosphere coupling on warm-season precipitation have also been widely explored in atmospheric general circulation models. There, land–atmosphere coupling on seasonal time scales has been established as an important factor determining predictability in certain regions (e.g., Koster et al. 2004, 2006). Similar studies for 0–24-h forecasts are less common, which may be partly related to the relatively poor predictability of convective precipitation on these shorter time scales (Fritsch and Carbone 2004).
It has been difficult to objectively demonstrate that high-resolution NWP models do a better job of predicting convective precipitation than the coarser operational models do. However, comparative studies using both enhanced convection-permitting grids with \( \Delta x \leq 4 \) km and coarser resolutions that rely upon cumulus parameterizations (Done et al. 2004; Kain et al. 2006; Weisman et al. 2008) have discussed how improvements in the realism of convection initiation and the mode of subsequent convection organization with explicit models provides value-added benefits to weather forecasters. This motivates us to use a convection-permitting model to study impacts of uncertainties in the surface exchange on convection initiation and subsequent precipitation in short-range forecasts.

We examine multiple 0–24-h forecasts for a 6-day “retrospective” period during the International \( \text{H}_2\text{O} \) Project (IHOP_2002) field campaign (Weckwerth et al. 2004), where deep convection was particularly active over the Great Plains of the United States (section 3). Our study is part of a broader, consolidated effort at the National Center for Atmospheric Research (NCAR) to improve short-term explicit precipitation prediction (STEP); through examining different components of the Advanced Research Weather Research and Forecasting Model (ARW-WRF; Skamarock et al. 2005) for this retrospective period. Past studies of warm-season precipitation have shown particular sensitivity to land surface processes over the southern plains region on longer time scales (e.g., Koster et al. 2004). Thus, we anticipate the representation of the surface exchange could impact short-range forecasts in this region as well.

The organization of the paper is as follows. In section 2, we review how the land–atmosphere exchange is handled in ARW-WRF. Section 3 provides an overview of our 6-day retrospective period and its contrasting precipitation events along with a description of the model and experiment design. The sensitivity of the simulated surface fluxes, planetary boundary layer (PBL), and precipitation to the strength of the parameterized surface exchange is examined and compared with observations in section 4. We emphasize how the strength of the surface exchange can influence convection initiation and precipitation forecasts in selected individual cases with different synoptic situations, and examine mechanisms by which this occurs, in section 5.

2. Surface exchange processes in the ARW-WRF–Noah model

The strength of the surface exchange is an important factor in the daytime growth and thermodynamic destabilization of the PBL, which often leads to the development of deep convection. The Noah land surface model (LSM; Ek et al. 2003) provides lower-boundary conditions for the PBL scheme in ARW-WRF, which depend on the surface fluxes of heat \( H \) and moisture \( LE \), defined in the bulk transfer formulas,

\[
H = \rho_c D C_H U(T_s - T - \gamma \Delta z) \quad \text{(1a)}
\]

\[
LE = \rho_d L_v C_H U(q_s - q) \quad \text{(1b)}
\]

In the above equations, \( \rho_cD \) and \( \rho_d \) are, respectively, the density of dry and moist air; \( C_H \) is the specific heat for air at constant pressure; \( L_v \) is the latent heat of vaporization; \( U, T, \) and \( q \) are the mean wind speed, temperature, and specific humidity at the first model level, respectively; \( \gamma \Delta z \) is an adiabatic correction to the temperature; \( T_s \) and \( q_s \) are the temperature and specific humidity at the surface (whose level is that of the roughness length for heat and moisture \( z_{0h} \)); and \( C_H \) is the bulk aerodynamic coefficient for heat (1a) and moisture (1b). To avoid singularities in convectively unstable situations \( (\alpha T/\beta z > \gamma) \), we use the Beljaars (1995) correction, as described in Janjic (1996b) and references therein.

In (1a) and (1b), larger \( C_H \) results in larger fluxes for the same vertical differences of \( (T_s - T) \) and \( (q_s - q) \). In the surface layer parameterization, \( C_H \) is approximated to be the same for heat and moisture and is estimated from the Monin–Obukhov similarity theory. An approximate form of Eq. (A4) in Chen et al. (1997) is used,

\[
C_H = \frac{k^2R}{\ln \left( \frac{z}{z_{0m}} \right) - \Psi_m \left( \frac{z}{L} \right) / \left[ \ln \left( \frac{z}{z_{0h}} \right) - \Psi_h \left( \frac{z}{L} \right) \right]} \quad \text{(2)}
\]

where \( k = 0.4 \) is the von Kármán constant, \( R \) is the ratio of the exchange coefficients for momentum and heat under neutral stability (assumed to be unity), and the functions \( \Psi_m \) and \( \Psi_h \) are corrections for the near-surface atmospheric stability \( z/L \), where \( z \) is the geometric height, \( L \) is the Obukhov length, and \( z_{0m} \) and \( z_{0h} \) are, respectively, the roughness lengths for momentum and scalars (e.g., heat and moisture). The \( z_{0m} \) is defined as the height at and below which the mean wind speed becomes zero and is a function of the vegetation category with values of about 0.05–0.10 m for grasslands to about 1 m for forested regions. The \( z_{0h} \), below which vertical transfer is through molecular diffusion and above which mixing by air currents dominates, is typically \( < z_{0m} \) but is less well-known (e.g., Chen et al. 1997; Chen and Zhang 2009).

The radiative skin temperature calculated in the LSM, \( T = T_s \), is used as a lower-boundary condition (at \( z = z_{0w} \)) for the surface layer parameterization in which \( C_H \) is
calculated. Here, \( z_{0r} \) is determined by the Zilitinkevich (1995) equation,

\[
z_{0r} = z_{0m} \exp \left( -kC_{zil} \frac{U_* z_{0m}}{v} \right),
\]

as described by Janjic (1996a). In (3), \( U_* \) is the friction velocity (i.e., square root of the surface stress), \( v \) is the kinematic molecular viscosity of air \((\sim 1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1})\), and \( C_{zil} \) is an empirical coefficient. In the current versions of ARW-WRF, \( C_{zil} \) is assigned a default value of 0.1 based on earlier comparisons of model results and field data (Chen et al. 1997).

Equation (3) relates \( z_{0r} \) and \( z_{0m} \), which are important in determining \( C_H \) [Eq. (2)] and, through Eqs. (1a) and (1b), the strength of the surface fluxes. From Eq. (3), estimates of \( z_{0m} \) are influenced by the appropriateness in choice of \( z_{0m} \), the accuracy of \( U_* \) obtained from the surface layer parameterization, and the specification of \( C_{zil} \). Since surface roughness (and wind drag) is strongly dependent on vegetation height, \( z_{0m} \) in NWP models is often specified as a function of the vegetation category alone. However, when this approach was adopted by Chen and Zhang (2009), \( C_{zil} \) variations of approximately two orders of magnitude (0.01 to 1.0) were needed to explain \( C_H \) variations derived using Eq. (1a) over a variety of vegetation types (including multiple types of grasslands, croplands, forests, and shrubland).

LeMone et al. (2008) compared observations of surface fluxes at three sites along a flight track in the western part of the IHOP region with results obtained from varying inputs (soil moisture, green vegetation fraction, and \( C_{zil} \)) to the Noah LSM-WRF surface-PBL parameterization run offline (uncoupled with the remainder of ARW-WRF). They found that the coupling strength for fluxes, \( \rho c_p C_H U \), was very sensitive to \( C_{zil} \), especially at low values of \( C_{zil} \) (Fig. 1). Their results in Fig. 1 are from an IHOP fair-weather day, where measurements of \( U, \rho, \) and \( c_p \) in Eq. (1a) were roughly constant, so variations along the \( y \) axis occur primarily from changes in \( C_H \).

The above results indicate considerable uncertainty in the bulk aerodynamic coefficients for heat and moisture that influence surface exchange strength. These results further suggest that the sensitivity of convective precipitation forecasts to possible ranges in surface exchange strength can be explored in a particularly simple fashion by varying the empirical parameter \( C_{zil} \) in models that employ Eq. (3) in their surface layer parameterization. This has motivated our design of numerical experiments discussed in the following section.

**3. Experiment Design**

**a. The 1200 UTC 10 June–1200 UTC 16 June 2002 IHOP retrospective period**

The 6-day retrospective was an active precipitation period having diverse precipitation systems with different forcing mechanisms over the IHOP region shown in Fig. 2a. Figure 3 indicates mesoscale convection that organized along quasi-stationary surface boundaries such as drylines and frontal zones (days 1–3) and a particularly large rain event (day 6) associated with a rapidly moving midtropospheric short wave and cold front at the end of the period. There was also a relatively precipitation free day (day 5) following a frontal passage in which the convection was orographically generated (Fig. 3) and limited to the western part of the IHOP region (Fig. 2). Apart from days that had strong synoptic forcing (days 4 and 6), convection typically initiated during the late afternoon with maximum domain-averaged precipitation amounts in the late evening (Fig. 3).

**b. Numerical model**

Our simulations utilize ARW-WRF (version 2) with a single 800 x 750 horizontal domain (Fig. 2) and 3-km horizontal grid spacing. This horizontal resolution captures the salient mesoscale aspects of convection without the need for cumulus parameterization. The vertical grid contains 42 levels that are stretched to provide enhanced resolution within the PBL (where \( \Delta z < 100 \text{ m} \))
and ~1-km spacing at the model top near 50 hPa. All simulations use the Thompson et al. (2008) bulk microphysical parameterization, which predicts cloud water, cloud ice, rain, snow, and graupel hydrometeor species. Other physical parameterizations include the Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) and Dudhia (1989) shortwave radiation schemes.

FIG. 2. (a) Map of USGS 24-category land use over the model domain for simulations described in section 3. Land use categories that do not occur over the simulation domain are marked with an asterisk in the legend at right. The white inner rectangle denotes the IHOP region of interest in the current study. This is the region for which area averages in Fig. 3 are computed. (b) Values of $C_{zil}$ (see text) for the simulations where it is a function of the vegetation types in (a) through Eq. (4). Both S2 and S9 are the locations of IHOP surface flux stations where corresponding model output is compared in Fig. 5.
The PBL parameterization (Janjic 1990, 1994, 2001) used in our primary simulations, referred to hereafter as the Mellor–Yamada–Janjic (MYJ) PBL scheme, predicts turbulent kinetic energy (TKE) and governs vertical mixing between model layers. Local forcing of TKE is provided by shear production, buoyancy production, and dissipation terms. Horizontal mixing is determined using a Smagorinsky first-order closure discussed in Section 4.1.3 of Skamarock et al. (2005).

The initial conditions for ARW-WRF are obtained from the National Centers for Environmental Prediction (NCEP) Environmental Data Assimilation System (EDAS) analyses, which have a horizontal grid spacing of ~40 km. Lateral boundary conditions with a 3-h frequency are generated from corresponding operational Eta Model for the same times.

This atmospheric model is coupled with the Noah LSM (Ek et al. 2003). The LSM has a single vegetation canopy layer and predicts volumetric soil moisture and temperature in four soil layers. The depths of the soil layers are sequentially 0.1, 0.3, 0.6, and 1.0 m. The root zone is contained in the upper 1 m (top-three layers).

The initial land surface conditions are supplied by the NCAR high-resolution land surface data assimilation system (HRLDAS). HRLDAS (Chen et al. 2007) is run offline but on the same 3-km horizontal grid as the ARW-WRF simulations for an 18-month spinup period prior to each forecast. This land surface initialization uses a variety of observed and analyzed conditions including the following: 1) the National Weather Service (NWS) Office of Hydrology Stage 4 rainfall data on a 4-km national grid (Fulton et al. 1998); 2) 0.5° hourly downward solar radiation derived from Geostationary Operational Environmental Satellite-8 and -9 (GOES-8 and GOES-9) as described by Pinker et al. (2002); 3) near-surface atmospheric temperature, humidity, wind, downward longwave radiation, and surface pressure from 3-hourly EDAS analyses; 4) 1-km horizontal resolution U.S. Geological Survey (USGS) 24-category land use and 1-km horizontal resolution state soil geographic soil texture maps; and 5) 0.15° monthly satellite-derived green vegetation fraction based on 5-yr averages (Gutman and Ignatov 1997).

c. Simulations

We analyze sets of simulations designed to examine the effect of the strength of the surface heat–moisture exchange on daytime PBL evolution, convection initiation, and 0–24-h QPF over the IHOP region. A set of three experiments (Table 1) use constant values of $C_{zil}$ and span a range of values consistent with results from empirical

<table>
<thead>
<tr>
<th>$C_{zil}$ Parameter Value</th>
<th>PBL scheme</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strong surface exchange</td>
<td>0.01</td>
<td>MYJ</td>
</tr>
<tr>
<td>Weak surface exchange</td>
<td>1.0</td>
<td>MYJ</td>
</tr>
<tr>
<td>WRF default</td>
<td>0.1</td>
<td>MYJ</td>
</tr>
<tr>
<td>Variable surface exchange</td>
<td>Function of vegetation type according to Eq. (4)</td>
<td>MYJ</td>
</tr>
<tr>
<td>Strong surface exchange</td>
<td>0.01</td>
<td>YSU</td>
</tr>
<tr>
<td>Weak surface exchange</td>
<td>1.0</td>
<td>YSU</td>
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Day 5 Early Afternoon Surface Conditions

(a) 19 UTC 14 Jun (t = 7 h) Soil Moisture for Variable CZIL Run

(b) 19 UTC 14 Jun (t = 7 h) Sensible Heat Flux for Variable CZIL Run

(c) 19 UTC 14 Jun (t = 7 h) Latent Heat Flux for Variable CZIL Run

FIG. 4. (a) Volumetric soil moisture in the top 0.1-m layer, (b) surface sensible heat flux, and (c) surface latent heat flux at 1900 UTC (1300–1400 LDT) 14 Jun 2002 for the simulation in which $C_{zil}$ is based on vegetation type (section 3c). The IHOP surface flux stations S2 and S9 for which simulated and observed fluxes are presented in Fig. 5 are annotated as in Fig. 2. The W and E partitioned rectangles denote subdomains for area averages presented in subsequent figures and discussed in the text.

studies (Chen et al. 1997; Chen and Zhang 2009). These include simulations with $C_{zil} = 0.01$ and $C_{zil} = 1.0$, which are respectively referred to as the strong surface exchange and weak surface exchange runs. Simulations with the standard $C_{zil}$ value used in recent versions of ARW-WRF of 0.1 are referred to as the WRF default runs. We analyze a fourth set of simulations where $C_{zil}$ varies across the domain as a function of momentum roughness length,

$$C_{zil} = 10^{-4.0z_{0m}},$$

based on empirical relationships between vegetation types and $C_H$ discussed in Chen and Zhang (2009). These simulations are referred to as the variable surface exchange runs (Table 1). Over most of the IHOP region, the variable $C_{zil}$ lies between the WRF default value of 0.1 and the weak exchange value of 1.0 (Fig. 2b). These relatively large $C_{zil}$ values are consistent with the relatively small roughness lengths of the dominant grassland, cropland, and shrubland vegetation types (Fig. 2a). In urban areas and in some forested regions near the edges of the IHOP subdomain (Fig. 2a), including the Ozark Mountains and the eastern edge of the Rocky Mountains, $C_{zil}$ values are approximately at or less than the strong exchange value of 0.01 (Fig. 2b).

It should be noted, however, that even in the constant $C_{zil}$ runs, $C_H$ varies spatially, primarily through its dependence on $z_{0m}$ (2), which is a function of vegetation category. These interdomain variations of $C_H$ for the constant $C_{zil}$ runs are still much less than those that occur in simulations in which $C_{zil}$ is allowed to vary according to (4). Each of the four sets of simulations with different specifications of $C_{zil}$ (and thus $C_H$) comprise 24-h forecasts initialized at 1200 UTC for each of the six individual days of the retrospective period (section 3a). To explore possible sensitivities to forecast length and initialization time, 12–36-h forecasts initialized at 0000 UTC were compared to their 0–24-h counterparts (i.e., same valid times) initialized at 1200 UTC.

The effects of surface exchange strength on the evolution of the daytime PBL and subsequent precipitation can be influenced by the choice of model PBL parameterization. We explore this sensitivity by performing simulations that use the Yonsei University (YSU) PBL scheme but are

1900 UTC (1300–1400 LDT) 14 Jun 2002 for the simulation in which $C_{zil}$ is based on vegetation type (section 3c). The IHOP surface flux stations S2 and S9 for which simulated and observed fluxes are presented in Fig. 5 are annotated as in Fig. 2. The W and E partitioned rectangles denote subdomains for area averages presented in subsequent figures and discussed in the text.
otherwise identical to the strong ($C_{zil} = 0.01$) and weak ($C_{zil} = 1.0$) coupling runs described above (Table 1). In contrast to the MYJ PBL scheme, the YSU scheme (Noh et al. 2003; Hong et al. 2006) allows nonlocal vertical mixing. Comparisons are made with the MYJ simulations for day 5 (1200 UTC 14 June–1200 UTC 15 June). On this day, afternoon cloudiness was less widespread than on other days, which affords a cleaner comparison of surface exchange effects on the afternoon PBL and subsequent precipitation. The general lack of clouds over the IHOP region on this day is reflected in the widespread strong early afternoon surface fluxes (Figs. 4b and 4c).

4. Sensitivity to surface exchange strength and comparison with observations

a. Comparison of simulated surface fluxes and PBL with local IHOP measurements

The simulated and observed fluxes at selected IHOP surface flux stations on day 5 (Fig. 5) represent the transition from predominately sensible $H$ to latent $LE$ fluxes from west to east across the IHOP region (cf. Figs. 4b and 4c). Though representativeness issues can complicate model comparisons with individual observation sites, the selection of stations with similar observed and model land use types (grasslands) and cloudless conditions may mitigate such difficulties to some degree. A model comparison with station S2 (Fig. 5a) suggests a positive bias in the strength of simulated $H$ in the western IHOP region, with values from the weak surface exchange run most closely matching observations. In contrast, the observed $LE$ lies in the middle of the range of simulated $LE$ at both the western and eastern edges of the region (Figs. 5b and 5d). Here, the variable surface exchange run agrees remarkably well with the observations at each of these stations, which span a wide range of soil wetness in the simulations (Fig. 4a).

The much greater total surface flux $H + LE$ in the strong surface exchange run than in the weak surface exchange run (Fig. 5) implies substantial differences in the surface energy budget, $R_{net} = H + LE + G$, where $R_{net}$ is the net radiation gain (including incoming and
reflected shortwave and outgoing longwave), and \( G \) is the flux into the ground. For example, at S2 midday \( H + LE \) is \( \sim 300 \text{ W m}^{-2} \) greater in the strong surface exchange run than in the weak surface exchange run (Figs. 5a and 5b), with \( \sim 150 \text{ W m}^{-2} \) less \( G \), which contributes to a lower skin temperature (\( \Delta T_s \sim -20 \text{ K} \)) and smaller outgoing longwave radiation that increases \( R_{\text{net}} \) by \( \sim 150 \text{ W m}^{-2} \). Together, the differences in \( G \) and \( R_{\text{net}} \) approximately balance those in \( H + LE \).

The westernmost station (S2 in Fig. 4) approximately coincides with the IHOP Homestead sounding site (Weckwerth et al. 2004) and thereby allows us to examine the impact of local surface fluxes on the afternoon clear convective boundary layer and evaluate how well this interaction is simulated at this location. More comprehensive studies of the observed PBL evolution on this day are found in Couvreux et al. (2009) and Bennett et al. (2010).

Figure 6 presents observations and the simulated PBL structure at our extremes of surface exchange strength (\( C_{zil} = 0.01 \) and \( C_{zil} = 1.0 \)) for day 5. The PBL thermal and moisture structures for the other simulations vary relatively smoothly between those of the simulations using our extremes, particularly for potential temperature (not shown). Although the weak surface exchange run (\( C_{zil} = 1.0 \) with MYJ PBL) produces fluxes that closely match the observations (Fig. 6c), the associated PBL is \( \sim 500 \text{ m} \) too shallow and \( \sim 3 \text{ K} \) too cool (Fig. 6a), whereas the strong surface exchange run (\( C_{zil} = 0.01 \) with MYJ PBL) has a PBL depth and potential temperature similar to the observations (Fig. 6a) despite much stronger than observed \( H \) (Fig. 6a). These comparisons suggest that the vertical mixing in the MYJ PBL scheme may not be aggressive enough at this particular location.

A simulation with strong surface exchange and the YSU PBL scheme (\( C_{zil} = 0.01 \) with YSU PBL) produces a warmer and deeper PBL than with MYJ (Fig. 6a) despite slightly smaller \( H \) (Fig. 6c). Here, the too warm and too deep YSU PBL is more consistent with the too large simulated \( H \) (Figs. 6a and 6c) than is the better represented PBL using MYJ. Although the differences in potential temperature among runs with different PBL schemes can be significant, these differences are much
smaller than those between runs for which \( C_{zil} \) is equal to 0.01 and 1.0 (Fig. 6a). This is not the case for the PBL moisture, where the choice of PBL scheme makes a larger difference than for potential temperature, particularly when the surface exchange is strong (Fig. 6b). Acting alone, the larger \( LE \) associated with stronger surface exchange (Fig. 6d) promotes greater PBL moisture. However, because of the very dry conditions above the PBL at this location (Fig. 6b), particularly deep vertical mixing occurs with strong surface exchange for the more aggressive YSU PBL scheme, leading to the driest PBL of all four simulations (Fig. 6b). This deeper and drier simulated daytime PBL using the YSU versus MYJ PBL scheme is consistent with results over the western high plains from previous studies (Weisman et al. 2008).

b. Regional comparison of PBL variables

Figure 7 presents a comparison of PBL variables in the primary MYJ simulations with the Rapid Update Cycle (RUC) model (Benjamin et al. 2004) analyses for the full diurnal cycle averaged over the 6-day retrospective period within the broader IHOP subdomain regions shown.
in Fig. 4. Here, we select RUC analyses as a proxy for observations since they both assimilate more observations at synoptic times than do the corresponding EDAS analyses used to initialize the ARW-WRF simulations (section 3b), and they are considered more independent from these simulations. The mean diurnal cycles of potential temperature and water vapor mixing ratio (Figs. 7a–d) are interpolated from the simulation and RUC analyses grids to 100 m AGL. This height is above daytime superadiabatic surface layers so that conditions are more representative of the PBL.

Over the western subdomain (Fig. 4), the magnitude of the diurnal cycle of potential temperature (Fig. 7a) and water vapor (Fig. 7c) in the WRF default and strong surface exchange simulations compare best with those of the RUC analyses. The much weaker diurnal cycle in the weak surface exchange runs (Figs. 7a and 7c) is consistent

**Fig. 8.** Comparisons of Stage 4 precipitation observations with simulated area-averaged hourly precipitation rates over the (a) western and (b) eastern subdomains depicted in Fig. 4 for 6-day averages of 0–24-h forecasts initialized at 1200 UTC. The vertical lines indicate approximate average noon and midnight LDT over the different averaging regions.

**Fig. 9.** Comparisons of Stage 4 precipitation observations with simulated area-averaged hourly precipitation rates over the entire IHOP region depicted by the solid rectangles in Fig. 4 for 6-day averages of (a) 0–24-h forecasts initialized at 1200 UTC and (b) 12–36-h forecasts initialized at 0000 UTC but valid for the same times as those in (a). (c) Equitable threat scores for simulated 3-h precipitation amounts calculated over the same IHOP region and averaged for the six 0–24-h (12–36-h) forecasts initialized at 1200 (0000) UTC. The vertical lines indicate approximate average noon and midnight LDT over the IHOP region.
with the shallower daytime PBL (Fig. 7e). The shallower PBL and its lesser vertical mixing in the weak surface exchange simulations contributed to a significantly cooler average afternoon PBL over this broad region than in observations (Fig. 7a), as found for the Homestead site on day 5 (Fig. 6a).

The cooler average PBL and its shallower depth in weaker surface exchange runs also occur over the eastern subdomain (Figs. 7b and 7f). However, in this region where $LE > H$ (Figs. 4b and 4c), the differences among simulations are less pronounced than over the west. One important difference between the two regions is that the strong surface exchange runs have a drier afternoon PBL than weak surface exchange runs in the west (Fig. 7c), whereas the opposite is true in the east (Fig. 7d). We attribute this regional difference in water vapor evolution among simulations to stronger vertical mixing of dry air into the PBL due to large sensible heat flux differences in the west (e.g., Fig. 5a), whereas large latent heat flux differences (e.g., Figs. 5d and 5f) dominate in the east.

c. Regional comparison of simulated and observed precipitation

Over the western subdomain, afternoon and evening 6-day-average precipitation rates are largest for the simulations with stronger surface exchange (Fig. 8a). The average onset of precipitation in the strong surface exchange runs also occurs $\sim 2$ h earlier than for the weak surface exchange runs (Fig. 8a), consistent with the more rapid growth of the daytime PBL (Fig. 7e). Average precipitation rates from the variable surface exchange runs most closely match those from Stage 4 precipitation observations (Fig. 8a), consistent with the best agreement of simulated to average RUC-analyzed 100-m moisture values around the time of afternoon convection initiation at $t = 9–12$ h (Fig. 7c).

Average precipitation rates over the eastern subdomain (Fig. 8b) are larger than over the western subdomain.

![Fig. 11. RUC analysis of 850-hPa winds, temperature (dashed contours, with 2.5°C contour interval), dewpoint (°C, scale at right), and geopotential height (solid contours, with 30-m contour interval). The rectangle denotes the IHOP analysis region.](image)

![Fig. 12. Equitable threat scores for simulated 3-h precipitation amounts calculated over the rectangular IHOP region in Figs. 11 and 13 for 0-24-h forecasts initialized at 1200 UTC 11 Jun 2002. The vertical lines denote average local daylight times over the region as in Fig. 9.](image)
Over the eastern subdomain, smaller average afternoon and evening precipitation rates in the weak surface exchange runs are similar to the western subdomain, however, the delayed onset of daytime precipitation relative to the strong surface exchange runs is less evident.

The pronounced afternoon minimum in observed precipitation in the eastern subdomain was not replicated by any of the sets of 0–24-h simulations (Fig. 8b). This shortcoming of the simulations is likely influenced by model spinup issues because the final stages of nocturnal convection that commonly occur over this latitude band near sunrise (e.g., Carbone et al. 2002) cannot be well simulated using a 1200 UTC initialization. Previous studies using convection-resolving versions of ARW-WRF have had success in simulating this daytime minimum in central plains precipitation when the model is run continuously over multiple diurnal cycles (e.g., Trier et al. 2006, 2010).

Differences in the 6-day-average simulated diurnal cycle of precipitation over the entire IHOP region (Fig. 2) are evident in a comparison between the 0–24- (Fig. 9a) and 12–36-h forecasts valid at the same times (Fig. 9a). For instance, the 12–36-h forecasts have greater postsunrise morning precipitation and a smaller evening maximum than do the 0–24-h forecasts. The 12–36-h forecasts have an afternoon precipitation minimum also found in the observations (though much stronger in the model) that was missed in the 0–24-h forecasts, presumably because of the previously noted model spinup issues in the 1200 UTC initializations.

Unlike for forecast length and initialization time, the surface exchange strength does not fundamentally alter the simulated diurnal cycle of precipitation (Figs. 9a and 9b). However, differences related to surface exchange strength occur in both the 0000 (12–36-h forecasts) and 1200 UTC (0–24-h forecasts) initializations. These include greater late afternoon–evening area-averaged precipitation (by up to 40%–100%) and an earlier onset (by up to 1–3 h) for strong compared to weak surface exchange (Figs. 9a and 9b).

Although some attributes of the 12–36-h forecasts, including the area-averaged afternoon precipitation minimum, may appear more realistic than their counterparts from the 0–24-h forecasts, their average skill as measured by the equitable threat score (ETS; Rogers et al. 1996) for 3-h precipitation totals is considerably less (Fig. 9c). This likely reflects the greater difficulty in precisely forecasting
the location of precipitation at longer lead times. In both the 0–24- and 12–36-h forecasts, despite substantial differences in area-averaged precipitation rates, the 6-day-average ETS differences among simulations of different surface exchange strengths are modest (Fig. 9c). However, individual cases from the 0–24-h forecasts discussed in the next section reveal larger forecast skill differences among these simulations under specific circumstances.

5. Case studies

The previous section highlights sensitivity of average precipitation rates to differences in the evolution of the daytime PBL for sets of simulations with different surface exchange strengths. In the current section, we examine in more detail how the surface exchange strength influences the timing of local deep CI and subsequent evolution of precipitation for three cases with different synoptic forcing.

Critical to the differences in CI onset are differences in the convection inhibition energy (CIN) that conditionally unstable air parcels originating in the daytime PBL must overcome to reach their level of free convection (LFC). The CIN is illustrated schematically by the gray shading in Fig. 10. CIN, however, is undefined at locations where a parcel LFC is absent [i.e., where zero convective available potential energy (CAPE) exists], which can result in discontinuities that pose a drawback to its spatial analysis. Alternatively, we can examine a continuous field

![Diagram showing different surface conditions](image-url)
approximating the CIN by using the maximum temperature deficit ($\Delta T_{\text{min}}$) of the most unstable lifted air parcel of 500-m depth. The most unstable parcels can occur at any level below ~3 km AGL, but for the current application of daytime CI they originate in the PBL.

The location of $\Delta T_{\text{min}}$ for a lifted PBL parcel is indicated by the white line in Fig. 10, and its value is given by the departure of the lifted parcel temperature from the environmental temperature. Note that the relationship between CIN and $\Delta T_{\text{min}}$ is analogous to the relationship between CAPE and the minimum lifted index for air parcels with positive buoyancy. In the forthcoming analysis, we make use of fields of both $\Delta T_{\text{min}}$ and CAPE to illustrate effects of surface exchange on CI.

a. Day 2 (1200 UTC 11 June–1200 UTC 12 June) CI along quasi-stationary boundaries

Day 2 of the retrospective period was characterized by late afternoon and early evening development of mesoscale clusters of deep convection along a surface front in Kansas and more isolated convection along a dryline (Cai et al. 2006; Wakimoto and Murphey 2010) that trailed southwestward into the Texas panhandle and southeast New Mexico. The 1200 UTC 11 June 850-hPa RUC analyses (Fig. 11) show very moist conditions ahead of the baroclinic zone, which moved slowly southeastward within the IHOP region during the day.

The ETS for this case indicates that the weak surface exchange run has the best afternoon and early evening QPF skill (Fig. 12). It is clear from comparing patterns of forecasted 3-h precipitation amounts (Figs. 13a and 13b) with observations (Fig. 13c) that the reduced late afternoon and evening ($t = 12–16$ h) ETS in the strong surface exchange run (Fig. 12) results primarily from overforecasting the spatial extent of the precipitation.

Figure 14 presents midafternoon CAPE and $\Delta T_{\text{min}}$ for the weak and strong surface exchange runs. Other simulations for this day and other cases had features intermediate between these extremes of surface exchange and are not presented. By midafternoon, CAPE is about $500–1000$ J kg$^{-1}$ larger in the strong surface exchange run for regions highlighted by the arrows (Figs. 14a and 14b). Since CAPE in these locations is large ($2000–4000$ J kg$^{-1}$) in both simulations, the differences in $\Delta T_{\text{min}}$ (Figs. 14c and 14d) are likely more critical to differences in the timing of CI and perhaps the subsequent precipitation amounts. In particular, narrow bands of reduced $\Delta T_{\text{min}}$ magnitude appearing in the strong surface exchange run (Fig. 14c) are absent in the weak surface exchange run (Fig. 14d). These bands, which extend along the leading edge of the surface front from Kansas into northwest Oklahoma (northeasternmost two arrows) and within the dryline moisture gradient in southeast New Mexico (southwesternmost arrow), are consistent with more widely forecasted precipitation within these and nearby regions during the next 2–5 h (Figs. 13a and 13b).

The reduction in $\Delta T_{\text{min}}$ along the leading edge of the surface front in the strong surface exchange run results from both $0.5–1.0$ g kg$^{-1}$ greater moisture (cf. Figs. 14a and 14b) and $0–2.5$ K warmer potential temperatures (cf. Figs. 14c and 14d), whereas within the dryline moisture gradients it is due entirely to the $2.5–5$ K warmer conditions (cf. Figs. 14c and 14d).

In contrast to the cold front and dryline zones, there are also broad regions of small $\Delta T_{\text{min}}$ magnitude located in southern Missouri and northern Arkansas in both simulations (Figs. 14c and 14d) where little subsequent precipitation occurs (Figs. 13a and 13b) despite substantial CAPE (Figs. 14a and 14b). This indicates that vanishing $\Delta T_{\text{min}}$ magnitude (and similarly vanishing CIN) is not a sufficient condition for CI. In the absence of significant convergence, factors inhibiting deep convection that are not considered by parcel theory including dry-air entrainment and downward-directed pressure forces within updrafts may be more important (e.g., Trier 2003). In the current case, CI is mostly limited to persistent mesoscale convergence zones. However, the CI is clearly modulated by the strength of the surface–atmosphere heat and moisture exchange.

b. Day 6 (1200 UTC 15 June–1200 UTC 16 June) squall line associated with a mobile short wave

The heaviest area-averaged precipitation event of the retrospective period occurred on the sixth and final day
The large-scale environment in its advance (Fig. 15) consisted of a southeastward-moving midtropospheric short wave along with strong warm advection, which together implied favorable quasigeostrophic forcing for ascent over the western part of the IHOP domain.

The initial formation of a mesoscale convective system (MCS) during the morning (Fig. 16g) is well forecasted by both the weak (Fig. 16a) and strong (Fig. 16d) surface exchange simulations. The success of all simulations in capturing the timing and location of the initial CI is revealed by large ETS for 3–6-h precipitation forecasts of ~0.5 (Fig. 17). The lack of sensitivity to surface exchange strength in the onset of the precipitation event differs from the previously discussed case. This aspect along

Fig. 16. As in Fig. 13, but for the (a),(d),(g) 3–6-, (b),(e),(h) 7–10-, and (c),(f),(i) 15–18-h periods of 0–24-h forecasts of day 6. The annotations 1 and 2 in (e) and (h) highlight precipitation features discussed in the text. The rectangles denote the region over which equitable threat scores and bias of simulated 3-h precipitation amounts are presented for the entire 0–24-h forecast period (Fig. 17) of day 6.
with the onset of precipitation relatively early in the diurnal cycle (prior to 1800 UTC) indicates a more dominant role of forced ascent on CI.

Differences in the forecasted 3-h precipitation amounts do eventually develop where, in contrast to the previous case (Fig. 12), there is greater skill in evening forecasts ($t = 12$–$18$ h) for the stronger surface exchange runs (Fig. 17). Their superiority in this case can be traced to two aspects beginning in the early-to-midafternoon.

The first is the observed development of precipitation within the MCS westward of $100^\circ$W denoted by feature 1 (Fig. 16h) being captured by the strong surface exchange run (Fig. 16e) while it is delayed in the weak surface exchange run (Fig. 16b). In both the weak and strong surface exchange runs, a moist tongue within surface southwesterly flow contributes to large CAPE upstream of the western part of the MCS (Figs. 18a and 18b). However, the potential temperatures are $\sim 2.5$ K warmer with comparable moisture in the strong surface exchange run (northeastern arrows in Fig. 18). These differences in potential temperature contribute to a more rapid reduction in the magnitude of $\Delta T_{\text{min}}$ along the southwest edge of the MCS gust front in the strong surface exchange run, which promotes triggering of new convection corresponding to feature 1 of Fig. 16 that has yet to occur in the weak surface exchange run (cf. Figs. 18c and 18d).

The second difference concerns the observed afternoon development of smaller convective clusters over elevated terrain to the southwest denoted by feature 2 (Fig. 16h), which is captured by the strong surface exchange run (Fig. 16e) but is missed by the weak surface exchange run (Fig. 16b). A large region of small negative buoyancy ($|\Delta T_{\text{min}}| < 1$ K) denoted by the southwesternmost arrow in the strong surface exchange run (Fig. 18c), which is absent in the weak surface exchange run (Fig. 18d), supports growth of this terrain-induced convection (Fig. 18c) as it drifts slowly southeastward. Several hours later the terrain-induced convection merges with the more rapidly southward moving convection along the MCS gust front (not shown), leading to similar evening positions and orientations of the southwestern leading edge of the MCS precipitation shield in the observations (Fig. 16i) and the strong surface exchange run (Fig. 16f). This interaction of these two different components of afternoon convection is not well simulated by the weak surface exchange run, and the result is a slower southward progression and modified orientation of the forecasted evening MCS precipitation (cf. Figs. 16c and 16i).

c. Day 5 (1200 UTC 14 June –1200 UTC 15 June)
orographically forced CI

The precipitation over the IHOP region on day 5 was the lightest and least widespread of any of the retrospective days (Fig. 3) because of unfavorable synoptic forcing associated with a midtropospheric ridge at the west edge of the region (Fig. 19). The precipitation that did occur was confined to a relatively narrow north–south corridor over which southeasterly (upslope) surface flow restored limited amounts of moisture into east-central and southeast Colorado (Figs. 20a and 20b). Unlike the previous two cases, there were no mesoscale boundaries or large-scale forcing to focus convection, so CI was dependent on smaller-scale terrain-induced convergence.

Modest CAPE of $\sim 250$–$1250$ J kg$^{-1}$ was limited to the region of positive horizontal moisture advection, and in contrast to days 2 and 6 was nearly equal in the weak and strong surface exchange runs (Figs. 20a and 20b). The nearly equal CAPE in the current case can be explained by the warmer midafternoon surface potential temperatures in the strong surface exchange run (Figs. 20c

Fig. 17. (a) Equitable threat score and (b) bias for simulated 3-h precipitation amounts calculated over the rectangular IHOP region in Figs. 15 and 16 for 0–24-h forecasts initialized at 1200 UTC 15 Jun 2002. The vertical lines denote average local daylight times over the region as in Fig. 9.
and 20d) being offset by smaller surface water vapor mixing ratios (Figs. 20a and 20b). This was typical of afternoon conditions over the western high plains during the 6-day retrospective period (Figs. 7a and 7c) and was attributed in the previous section to the enhanced vertical mixing of dry air from aloft into the PBL in the strong surface exchange run.

As for days 2 and 6, the differences in the timing of convection triggering can be linked to the more rapid reduction of $\Delta T_{\text{min}}$ magnitude in the strong surface exchange run (Fig. 20c) than in the weak surface exchange run (Fig. 20d), which is related to warmer surface potential temperatures. In addition, the stronger surface exchange results in slightly enhanced upslope flow over southeast Colorado, which could locally enhance surface convergence. By 2100 UTC, small storms occur near terrain features in the strong surface exchange simulation (Fig. 20c), whereas significant negative buoyancy ($\Delta T_{\text{min}} < -1$) is still widespread from the mountains eastward in the weak surface exchange run and inhibits storm development (Fig. 20d). Convection eventually develops in the weak surface exchange run, but the precipitation is less widespread (Fig. 21a) and located west of the precipitation in the strong surface exchange run (Fig. 21c), which has become organized into small convective clusters by evening.

The ETS for the WRF default, variable surface exchange, and strong surface exchange simulations show only modest skill (Fig. 22a), which is not surprising considering the small scale of the precipitation patterns.

**FIG. 18.** As in Fig. 14, but for 2100 UTC (−1600 LDT) 15 Jun 2002.
during this period of unfavorable synoptic forcing (Fig. 19). However, the forecast skill in these simulations represents an improvement over that of the weak surface exchange run, which demonstrates no skill (Fig. 22a). The lesser skill of the weak surface exchange run in the current case is accompanied by a strong bias toward too little precipitation throughout the period (Fig. 22b), which is a consequence of the PBL experiencing too little daytime growth and warming.

For this day we examined the relationship between precipitation and the land–atmosphere exchange for an additional set of simulations that utilized the YSU PBL scheme. Here, we find that the choice of PBL scheme does not fundamentally alter the role of surface exchange strength on precipitation; namely, that the stronger surface exchange results in heavier precipitation that initiates earlier and therefore has a leading edge that advances eastward more rapidly (cf. Figs. 21a and 21c with Figs. 21b and 21d). Precipitation amounts are, however, significantly influenced by the different PBL schemes. In particular, the simulations that use YSU produce less precipitation in high plains locales (Fig. 21), which may be partly related to deeper vertical mixing reducing afternoon PBL moisture in YSU (relative to MYJ) as discussed earlier for upstream afternoon soundings (Fig. 6b).

6. Summary and discussion

In this study, we examine the sensitivity of the daytime PBL and precipitation in a cloud-resolving atmospheric model (ARW-WRF) to the parameterized surface–atmosphere exchange strength for a 6-day convectively active period during IHOP_2002 field phase. The surface exchange strength in the model was influenced by varying the Zilitinkevich (1995) coefficient \( C_{zil} \) in Eq. (3), which is typically set to a domainwide constant value, through a range representative of maximum and minimum derived values (e.g., Chen and Zhang 2009).

These numerical experiments established sensitivity of both the timing of deep convective precipitation and area-averaged precipitation amounts to prescribed \( C_{zil} \) values. Simulations were compared with IHOP observations of the fluxes and daytime fair-weather PBL and more widespread PBL and precipitation properties determined from gridded model-based analyses and Stage 4 precipitation analyses for the entire 6-day retrospective. Here, the observations for the most part intersected the broad range of possible responses from the simulations with different surface exchange strengths.

The surface exchange strength does not fundamentally alter the general location of mesoscale precipitation systems and overall characteristics of their forecasted diurnal cycle, which contrasts with other sensitivities explored in the study including model initialization time. However, both 6-day averages and a more detailed examination of individual cases revealed that simulations with strong surface exchange (\( C_{zil} = 0.01 \)) systematically produced precipitation that both initiated up to several hours earlier and had greater amounts than in corresponding simulations with weak surface exchange (\( C_{zil} = 1.0 \)). The quicker onset and larger precipitation amounts in the strong surface exchange runs were linked to more rapid growth and warming of the daytime PBL owing to enhanced surface sensible heat flux. The simulations that used \( C_{zil} = 0.1 \), the default value in the current versions of ARW-WRF, produced both a quicker onset of precipitation and larger overall amounts than observed in 6-day averages, suggesting that the surface exchange may be somewhat too strong.

These model sensitivities of precipitation to surface exchange strength appear comparable or even greater than those in previous studies of land surface–atmosphere interaction where different land surface models were used (e.g., Trier et al. 2008) and initial land surface conditions including the specificity of soil wetness (e.g., Trier et al. 2008) and green vegetation fraction (e.g., James et al. 2009) were varied. Examination of a particular case in the current study that lacked large-scale forcing (day 5) suggests that effects of the surface exchange strength on precipitation could also be comparable or greater than those associated with the choice of PBL scheme. Thus, being able to properly account for uncertainties in the surface exchange strength could be potentially beneficial for some forecasting applications.

Precipitation sensitivity to surface exchange strength was greatest over the high plains part of the IHOP region located west of \( \sim 100^\circ \text{W} \) longitude. This may be due in part to drier soils in these locations having a greater influence...
on enhancements of the sensible heat flux $H$ differences among simulations. For dry soils, the temperature difference $T_s - T$ in (1a) is larger than for wetter soils. When $C_{zH}$ is increased, the bulk aerodynamic coefficient for heat and moisture $C_H$ is reduced, which reduces $H$. This increases $T_s$ and hence $T_s - T$ partially compensate the reduction in $C_H$. For larger initial values of $T_s - T$ (dry soils), the fractional impact of this compensation is less, which allows greater $H$ changes.

Concepts from the parcel theory of conditional instability were applied to three diverse precipitation events during the retrospective period to illustrate in greater detail the role of surface exchange strength in daytime convection initiation. The more rapid convection initiation in the strong surface exchange simulations is primarily due to the earlier removal of negative buoyancy for conditionally unstable PBL air parcels. Although the stronger surface exchange can enhance CAPE and reduce the negative buoyancy through input of both heat and moisture into the PBL, it is the associated potential temperature increases that appear most crucial to these convection initiation timing differences. This was most evident over the western part of the IHOP region, where deeper vertical mixing results in greater PBL drying during the day than in weak surface exchange runs despite increased moisture input from the surface. Forecasted precipitation differences among simulations with different surface exchange strength can occur from the onset of convective triggering (e.g., days 2 and 5), while in cases with strong large-scale forcing they may not develop until much later as MCSs mature (e.g., day 6).

![Fig. 20. As in Fig. 14, but for 2100 UTC (−1600 LDT) 14 Jun 2002.](image)
For 6-day averages over the western part of the IHOP region, precipitation forecasts using variable $C_{zil}$, which depended on vegetation type (heights) through assigned momentum roughness lengths (e.g., Chen and Zhang 2009), most closely followed the Stage 4 observations both in timing of convection initiation and overall amounts. Thus, allowing the surface exchange strength to vary based on properties of the vegetation indicates the potential promise for more realistic operational forecasts of precipitation in this convection-triggering region.

Fig. 21. Three-hour precipitation amounts for simulations (a)–(d) with different combinations of PBL scheme (columns) and surface exchange strength (rows) during $t = 12–15$ h of the day 4 forecasts, and (e) observed (Stage 4) precipitation amounts for the same period. The rectangles denote the region over which equitable threat scores of simulated 3-h precipitation amounts are presented for the entire 0–24-h forecast period (Fig. 22) of day 5. The cross symbol indicates the location of the Homestead sounding site from which observed and simulated PBL profiles from earlier in the afternoon are presented in Figs. 6a,b.
Farther east (i.e., east of 100°W), the advantage of using variable $C_z$ was less evident. We speculate that one possible reason for these geographical differences is that a sizeable fraction of the convection over the central plains originates from upstream and are perhaps more likely to be strongly influenced by cumulative errors from other model parameterizations. Clearly, the systematic variability of surface exchange effects on precipitation over the IHOP region alone indicates the need for additional studies that investigate the applicability of the current results for both different locations and for other seasons.

A major impediment to the accurate representation of surface exchange in operational models is the uncertainty in the roughness length for heat and moisture $z_0$. Allowing $C_z$ to vary in Eq. (3), as was done in the current study, is an expedient way to strongly influence $z_0$ and thereby examine the sensitivity of precipitation to a broad range of surface exchange strengths. However, other factors including how the roughness length for momentum $z_{0m}$ is specified and the accuracy of the friction velocity $u_*$ calculation also influence $z_0$ and the surface exchange. More research is needed to discern how best to determine $z_0$ in operational models using Eq. (3) or other techniques (e.g., Brutsaert 1982).

As future research stimulates improvements in the parameterization of surface exchange, additional work will likely be required to optimally translate such improvements into increased QPF skill. This is because errors in surface exchange combine with other sources of model error to influence PBL and precipitation forecasts. Frameworks that allow for assessment of the performance of combinations of multiple components of a modeling system on forecasts (e.g., Santanello et al. 2009) may be helpful in this regard.

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REFERENCES


Chen, F., and Y. Zhang, 2009: On the coupling strength between the land surface and the atmosphere: From viewpoint of surface


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