Evaluation of the Noah Land Surface Model Using Data from a Fair-Weather IHOP_2002 Day with Heterogeneous Surface Fluxes

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ABSTRACT
Sources of differences between observations and simulations for a case study using the Noah land surface model–based High-Resolution Land Data Assimilation System (HRLDAS) are examined for sensible and latent heat fluxes \( H \) and \( LE \), respectively; surface temperature \( T_s \); and vertical temperature difference \( T_0 - T_s \), where \( T_0 \) is at 2 m. The observational data were collected on 29 May 2002, using the University of Wyoming King Air and four surface towers placed along a sparsely vegetated 60-km north–south flight track in the Oklahoma Panhandle. This day had nearly clear skies and a strong north–south soil-moisture gradient, with wet soils and widespread puddles at the south end of the track and drier soils to the north. Relative amplitudes of \( H \) and \( LE \) horizontal variation were estimated by taking the slope of the least squares best-fit straight line \( \Delta LE/\Delta H \) on plots of time-averaged \( LE \) as a function of time-averaged \( H \) for values along the track. It is argued that observed \( H \) and \( LE \) values departing significantly from their slope line are not associated with surface processes and, hence, need not be replicated by HRLDAS. Reasonable agreement between HRLDAS results and observed data was found only after adjusting the coefficient \( C \) in the Zilitinkevich equation relating the roughness lengths for momentum and heat in HRLDAS from its default value of 0.1 to a new value of 0.5. Using \( C = 0.5 \) and adjusting soil moisture to match the observed near-surface values increased horizontal variability in the right sense, raising \( LE \) and lowering \( H \) over the moist south end. However, both the magnitude of \( H \) and the amplitude of its horizontal variability relative to \( LE \) remained too large; adjustment of the green vegetation fraction had only a minor effect. With \( C = 0.5 \), model-input green vegetation fraction, and our best-estimate soil moisture, \( H, LE, \Delta LE/\Delta H, \) and \( T_0 - T_s \) were all close to observed values. The remaining inconsistency between model and observations—too high a value of \( H \) and too low a value of \( LE \) over the wet southern end of the track—could be due to HRLDAS ignoring the effect of open water. Neglecting the effect of moist soils on the albedo could also have contributed.

1. Introduction
This paper is the third of a series that addresses the horizontal variability of sensible and latent heat fluxes, \( H \) and \( LE \), and their representation in land surface models, this time in the form of the Noah land surface model–based High-Resolution Land Data Assimilation System (HRLDAS; Chen et al. 2007). In the first two papers (LeMone et al. 2003, 2007b), we examined horizontal variability along the amply watered and densely vegetated eastern track in Fig. 1, using data from the International \( H_2O \) Project (IHOP_2002; Weckwerth et al. 2004; LeMone et al. 2007a) and the 1997 Cooperative Atmosphere–Surface Exchange Study (CASES-97; LeMone et al. 2000). Here we focus on the more sparsely vegetated IHOP_2002 western track, located in the Oklahoma Panhandle.

IHOP_2002 was organized to improve prediction of warm-season precipitation; the surface and boundary layer component looked at how surface processes affect the planetary boundary layer (PBL) and subsequent
development of precipitating convection (LeMone et al. 2007a). It is now well documented that surface processes have significant impact on the origin and development of warm-season convective precipitation (e.g., Anthes 1984; Segal et al. 1988; Chen et al. 1996; Holt et al. 2006). This occurs through affecting local PBL profiles (Findell and Eltahir 2003a,b; Trier et al. 2004; Pielke et al. 1997), or through differential surface heating, which leads to mesoscale circulations that favor the development of convection in their upwelling branches (e.g., Chen et al. 2001b; Trier et al. 2004).

The version of HRLDAS used here is the “default” version described in Chen et al. (2007), except that the land-use data are based on Moderate Resolution Imaging Spectroradiometer (MODIS) data from the Department of Geography at Boston University (see online at www-modis.bu.edu/landcover/index.html). HRLDAS is an offline version of the Noah land surface model (LSM; Mahrt and Ek 1984; Pan and Mahrt 1987; Noilhan and Planton 1989; Chen et al. 1996; Chen and Dudhia 2001a; Ek et al. 2003) developed for mesoscale numerical weather prediction (NWP) models. The Noah LSM and its predecessors have been extensively tested in uncoupled mode (e.g., Chen et al. 1996, 2003; Koren et al. 1999; Slater et al. 2001; Sridhar et al. 2002; Mitchell et al. 2004; Chen 2005) and in coupled mode with atmospheric models (e.g., Chen et al. 1997, 2007; Yucel et al. 1998; Chen and Dudhia 2001b; Ek et al. 2003) for both warm and cold seasons. Chen et al. (2001a) demonstrated that replacing a “bucket model” with the Noah LSM improved 24–48-h precipitation forecasts in the National Centers for Environmental Prediction (NCEP) Eta Model as much as doubling the model horizontal resolution. Today, Noah is used in operational NWP models at NCEP and the U.S. Air Force Weather Agency.

In the first two papers of this series, we used data for the moist grasslands and croplands of southeast Kansas to introduce and interpret the slope $\Delta \text{LE}/\Delta H$, where $H$ and LE are time-averaged values over different sites in an area with horizontally uniform downwelling radiation. In LeMone et al. (2003), we used the surface energy budget and data from CASES-97 to show that $\Delta \text{LE}/\Delta H$ should be negative and near $-1$ for clear-sky days, with positive values indicating an insufficient sample or horizontal variability too small to define a slope. In LeMone et al. (2007b), data and land surface model runs for CASES-97 and IHOP_2002 were used to show that $\Delta \text{LE}/\Delta H$ tended to be ill-defined right after widespread rainfall, with the slopes evolving with time from around $-1.5$ to $-2.5$ when they were first definable to shallower values a few days later. Correspondingly, there was a general tendency for shallower slopes with drier soils.

This paper’s original objective was to demonstrate the use of horizontal heterogeneity of surface fluxes and surface temperature as well as their means and evolution to evaluate the ability to HRLDAS to replicate the horizontal variation of fluxes during IHOP_2002. We were surprised to find that Noah–HRLDAS did not work as well as expected for 29 May 2002, a day characterized by significant horizontal heterogeneity (Kang et al. 2007).

Hence, this paper’s objectives changed from a model–observation comparison for a number of days to exploring the reasons HRLDAS failed to replicate the observations of $H$, LE, and $T_s$ along the IHOP_2002 western track on 29 May 2002, suggesting some rem-
Section 2 describes the data and the HRLDAS modeling system. The control run using default HRLDAS values is described in section 3. Using results from the control run to evaluate various terms in the model equations, we show why \( H \) and LE produced by HRLDAS should be anticorrelated along the flight track, and then discuss the Zilitinkevich coefficient. Results of the sensitivity studies are described in section 4. Results and further aspects of the analysis are discussed in section 5, and conclusions are summarized in section 6.

2. Observations and HRLDAS

a. Data collection and selection of case study

The observations on this day are from surface flux towers and five ~60-km flight legs flown at ~65 m AGL by the University of Wyoming King Air along the IHOP_2002 western track (Fig. 1). In contrast to the eastern track (LeMone et al. 2007b), this region had much less rainfall and vegetation. Furthermore, it has more heterogeneous soil types and land cover. The four surface flux sites along the western track [sites 1–3 maintained by NCAR, site 10 by the University of Colorado (CU)] were located on different types of surface—bare ground (site 1), ungrazed Conservation Reserve Program grassland (site 2), sagebrush (site 3), and heavily grazed grass (site 10)—allowing a good sample of horizontal variability from the surface as well as from the air.

Aircraft \( H \) and LE estimates are based on vertical velocity determined from the Rosemount 858AJ/1332 differential gust-probe system and a Honeywell Laserf SM inertial navigation system, with temperature from a reverse-flow thermometer developed at the University of Wyoming. Water vapor fluctuations are sampled using a Li-Cor-6262 gas analyzer and a Lyman-\( \alpha \) instrument from NCAR. The Li-Cor measures infrared absorption by water vapor; while the Lyman-\( \alpha \) measures absorption of ultraviolet radiation. The Li-Cor is standard on the King Air, was used for LeMone et al. (2003) and LeMone et al. (2007b), and provides stable measurements, while the Lyman-\( \alpha \) has faster response but needs periodic calibration using a reference water vapor density; hence, we show results from both here. Surface \( H \) and LE are based on sonic temperature and vertical velocity from sonic anemometers, and mixing ratios derived from Campbell Scientific KH20 Krypton hygrometer water vapor densities.

The radiometric surface temperature \( T_s \) is sensed by a Heiman KT-19.85 radiometer on the King Air; and by Everest 4000.4GL sensors at the surface flux sites. Kang et al. (2007) indicate reasonable consistency between the aircraft and surface measurements, although L. Oolman (University of Wyoming 2007, personal communication) suggests there is a slightly cool daytime bias (~1 K) in \( T_s \) due to pitting on the lens incorporating some effect of the air temperature at flight level. For comparison with HRLDAS values, we estimated the difference between \( T_s \) and 2-m temperature \( T_2 \) by subtracting the best-fit straight line to \( T_2 \) from the surface sites (average of 1815 and 1845 UTC) from the best-fit quadratic for the aircraft \( T_s \) for the flight leg centered at 1830 UTC. The aircraft normalized differential vegetation index (NDVI) is from Exotech radiometer data.

The soil and vegetation data used here were collected in several different ways. Recorded values of soil moisture at 5 cm at the NCAR sites and the profile at the CU site were measured using Campbell Scientific CS-615 probes; profiles at the NCAR sites were based on Decagon ECH2O probes, installed and processed in partnership with the Hydrologic Science Team in the Department of Bioengineering of Oregon State University (OSU). Manual measurements of soil moisture were made using a MESA Systems TRIME FM-3 sensor on 29 and 30 May. On the same visits, NDVI measurements were made using a Cropscan MSR5 probe. We used soil temperatures from Campbell Scientific CS-107 probes at the NCAR sites, and REBS STP-I probes at the CU site. Soil heat flux was measured using REBS HFT-3 flux plates at all four sites. Soil characterization (i.e., mineral content, texture, and bulk density) was done for sites 1–3 by the OSU Hydrologic Science Team and for site 10 by Alfieri at CU. Further details are available from LeMone et al. (2007a).

We selected 29 May 2002 for this study for two reasons. First, the measured fluxes were heterogeneous, consistent with a strong north–south gradient in soil moisture; and second, we wanted to understand why HRLDAS did not do better in replicating the observations. A strong rain event 2 days previous deposited ~80 mm of rain at site 1 near the south end of the track, and ~10–30 mm at the remaining sites. The heterogeneous fluxes on this day and their impact on boundary
layer depth are summarized in Kang et al. (2007). From Strassberg et al. (2008), the mean wind is from 194° at 4.8 m s\(^{-1}\) at 65 m and the Monin–Obukhov length is \(-25\) m. According to Kang et al., the PBL depth varied from about \(1\) km on the south end of the track to \(1.3\) km on the north end.

**b. Flux calculation**

Aircraft fluxes for each flight leg were found from covariances after subtracting out the linear trends for each variable and shifting the temperature and mixing ratio time series to account for time lags relative to the vertical velocity, which were induced by spatial separation on the aircraft and differences in sensor time response. The covariances were averaged in two ways: in overlapping \(4\)-km running averages at \(1\)-km intervals for looking at horizontal variability, and in \(1\)-km bins for examining the effect of horizontal filtering on \(\Delta LE/\Delta H\). The data processing is identical to that described in LeMone et al. (2007b), except that the individual legs were truncated so that the averages for all flight legs are at the same geographic points, eliminating the need to interpolate before averaging.

Surface fluxes, processed by NCAR for sites 1–3 and CU for site 10, are relative to half-hour averages. Details can be found in LeMone et al. (2007a) or at the NCAR Earth Observing Laboratory (EOL) Web site. (The processed aircraft and surface data are available at www.rap.ucar.edu/research/land/observations/ihop.php.) Observations along the flight track are summarized in Table 1.

**c. HRLDAS**

The HRLDAS model (Chen et al. 2007) is based on the Noah LSM (Mahrt and Ek 1984; Pan and Mahrt 1987; Noilhan and Planton 1989; Chen et al. 1996; Chen and Dudhia 2001a; Ek et al. 2003). We use 18 months of input data to obtain soil-moisture and temperature profiles that are physically realistic and consistent with past weather and with the physics package of its parent Weather Research and Forecasting (WRF) Model. The HRLDAS physics are as described in Chen et al. (2007), but we alter the Zilitinkevich coefficient for some runs. This will be discussed further after HRLDAS physics are introduced in section 3. We run HRLDAS on a \(3\)-km grid.

Input parameters for HRLDAS consist of static and time-varying surface databases and historical meteorological data. The static databases include a 19-category land surface characterization based on \(2001\) MODIS data from the Department of Geography at Boston University (www-modis.bu.edu/landcover/index.html), and the \(1\)-km resolution State Soil Geographic (STATSGO) soil data. Time-varying fields include the \(0.15^\circ\) monthly green vegetation fraction based on satellite-based climatological data collected between 1985 and 1990 (Gutman and Ignatov 1998). These are currently standard for HRLDAS (Chen et al. 2007), except for the MODIS land-cover data, which are used because of changes in land use along the western track. As in the case of HRLDAS with the default vegetation table, albedo and emissivity are defined according to vegetation type and are kept constant except for alteration for snow cover.

The meteorological data (i.e., temperature, humidity, wind, pressure, and downwelling longwave radiation) are from the \(3\)-hourly NCEP Eta Data Assimilation System (EDAS), while downwelling shortwave radiation is from \(0.5^\circ\) Geostationary Operational Environmental Satellite-8 and -9 (GOES-8 and -9) data (Pinker et al. 2002), and rainfall is from \(4\)-km hourly National Oceanic and Atmospheric Administration (NOAA) National Weather Service (NWS) Office of Hydrology stage IV rainfall (Fulton et al. 1998).

For 29 May, the input meteorological data are consistent with a near-cloudless fair-weather day, with almost no horizontal variability in downwelling radiation. The input downwelling radiation is \(\sim 50\) W m\(^{-2}\) higher than the sites 1–3 average, almost completely due to
differences in the longwave component. A half-hour\(^1\) shift in the input downwelling solar radiation partially offsets this bias during the averaging period 1700–2100 UTC, but causes an artificially early sunset and leads to underestimates of \(H\) at 0000 UTC. Because of the latter bias, \(H\) values at 0000 UTC were recomputed for selected runs by averaging \(H\) at 2300 and 0000 UTC to obtain a model \(H\) at 2330 UTC for comparison to the observed 0000 UTC value. This corrects the timing of the solar radiation but shifts the other input data, so the true \(H\) estimates are likely between the original and “shifted” values. Estimated albedos and emissivities are close to input values. Input temperatures at 2 m are 1–2 K higher than observations between 1700 and 2100 UTC. Since there is not exact agreement, we will use the difference \(T_s - T_0\) as well as \(T_s\) in evaluating HRLDAS. The western track is distant from the Next Generation Weather Radar (NEXRAD) radars, lead-

\(^1\) The radiation data are documented as applying to 15 min past the hour, but comparisons with observations suggested they applied to about 30 min past the hour. For these simulations, the radiation data were assumed to be on the hour.
ing to errors in the HRLDAS input rainfall data. Figure 6 of Chen et al. (2007) shows a significant underestimate of the 26–27 May rain event at site 1, and overestimates at sites 2 and 3; with input rainfall totals from 10 May in agreement for site 1, and overestimates at sites 2 and 3.

3. Results: The control run

a. Comparing HRLDAS and observations

Figures 2–4 compare observations for 29 May 2002 on the IHOP_2002 western track with results from our
control HRLDAS run for the variables it supplies to its parent NWP model, namely $H$, LE, and $T_s$.

Figure 2 shows $H$ and LE from HRLDAS control along with the corresponding observations. Fractional random error for the 4-km averages is 0.12 for $H$ and 0.34 for LE [LeMone et al. (2007a), based on Mann and Lenschow (1994)]. In the figure, the observed northward increase in $H$ (decrease in LE) is consistent with the heavy rainfall at the south end of the track 2 days previous. HRLDAS $H$ is higher and LE is mostly lower than what is observed, and the horizontal variability is not captured.

The differences between the model and observations appear to be real. The slight differences between surface and aircraft flux observations reflect true changes in the vertical. The small differences suggest that adjusting the aircraft data to surface values would bring the aircraft data closer to HRLDAS results, but not enough. To confirm this, we produced linear mean flux profiles interpolated to 1830 UTC from aircraft data and used the profiles to adjust the flight-level $H$ and LE curves to surface values assuming a constant change with height. This procedure increased the $H$ curve to 108% of its flight-level value and reduced the LE curve to 81%-94% of its flight-level value, far from enough to explain the discrepancy (not shown). Furthermore, if we shift “observed” aircraft $H$ and LE curves left in Fig. 2 to correspond to their corresponding surface source regions—at most 2–3 km based on Song and Wesely (2003)—this is not enough to force $H$ and LE to modeled values. For comparison, both $H$ and LE in Kang et al. (2007) are similar to Fig. 2 except for a drop in $H$ and a rise in LE at the northern end of the track and smaller averages due to their calculating fluxes relative to 4-km averages, which excludes the contribution of scales >4 km.

In Fig. 3, HRLDAS $H$ is greater not only than the average observed $H$ for most of the daytime hours, but it is also greater than the highest values observed at the surface. In HRLDAS, $H$ becomes more negative (−64 W m$^{-2}$) than any of the observed values at 0000 UTC, likely a result of the half-hour shift in input solar radiation. From section 2c, the “true” model $H$ value at 0000 UTC is probably between −64 and −11 W m$^{-2}$, the value corresponding to the correct input solar radiation at 0000 UTC. HRLDAS LE lies within the range observed and is fairly close to the surface average. Like the modeled fluxes in Fig. 2, HRLDAS $T_s$ (Fig. 4), does not exhibit the observed north–south gradient; and it slightly cooler on average than observed.

b. Application and interpretation of $\Delta$LE/$\Delta H$

1) $\Delta$LE/$\Delta H$ and Its Use in Judging What HRLDAS Should Replicate

On larger scales, observed $H$ and LE are anticorrelated in Fig. 2, with $H$ increasing and LE decreasing toward the north, satisfying the LeMone et al. (2003, 2007b) “anticorrelation” criterion for aircraft data representing surface processes on days with clear skies. Similarly, the observed slopes of the best-fit lines in the plots of LE as a function of $H$ in Fig. 5 support an overall anticorrelation, with the aircraft-based slope $\Delta$LE/$\Delta H$ (from −1.6 to −1.7) only slightly greater than the −1.5 slope for the surface data. This indicates that at the larger scales represented by the slopes, the amplitude of horizontal LE variation is about 1.6 times the amplitude of horizontal $H$ variation. However, there seems almost no relationship between $H$ and LE fluctuations at smaller scales. This is not surprising since the magnitudes of the fluctuations are of the order of the corresponding standard errors in Fig. 2. The slope for the HRLDAS control run is better defined but much shallower.

Figure 6 shows how the observed slopes change when we produce $H$-versus-LE plots using data smoothed with increasing horizontal filtering. LeMone et al. (2003, 2007b) interpret the steepening in slope from 1- to 4-km filter lengths as resulting from removal of the effect of large eddies concentrating both $H$ and LE in their upwelling regions. As we smooth the data further, removal of the remaining smaller-scale $H$ and LE fluctuations increases the correlation coefficient for the corresponding $\Delta$LE/$\Delta H$ line without changing its
Thus, we associate only the larger-scale variations that produce the approximately $H/100^2$ slope with surface processes, dismissing the smaller-scale fluctuations as largely statistical noise.

2) HRLDAS: Why $H$ and LE should be anticorrelated

In the foregoing, we argued that the $H$ and LE variation in Figs. 2 and 5 that correspond to surface processes should be anticorrelated. Here, we show why $H$ and LE should also be anticorrelated in HRLDAS for the conditions (nearly uniform radiative forcing) represented on 29 May, with a negative $\Delta H/\Delta LE$ slope.

We start with the surface energy balance neglecting storage in the vegetation canopy:

$$ R_{\text{net}} = H + LE + G_{\text{sfc}}, $$

where $R_{\text{net}}$ is the net radiation, and $G_{\text{sfc}}$ is the heat flux into the soil. If we rearrange (1a) slightly,

$$ H + LE = R_{\text{net}} - G_{\text{sfc}}, $$

we can argue that higher (lower) values of $H$ mean lower (higher) values of LE, provided the available energy $R_{\text{net}} - G_{\text{sfc}}$ does not vary much horizontally. Thus, we expect $H$-versus-LE plots to show a negative slope around $-1$ under these conditions.

To show how $\Delta LE/\Delta H$ deviates from $-1$, we express $H, R_{\text{net}},$ and $G_{\text{sfc}}$ in terms of $T_s - T_0$, where $T_0$ is the air temperature at the lowest model level (i.e., 2 m). Provided $T_s - T_0 \ll T_0$, $R_{\text{net}}$ is given by

$$ R_{\text{net}} = S_\downarrow (1 - \alpha) + \varepsilon L_\downarrow - \varepsilon \sigma T_s^4 $$

$$ \approx S_\downarrow (1 - \alpha) + \varepsilon L_\downarrow - \varepsilon \sigma T_0^4 \left[ 1 + \frac{4(T_s - T_0)}{T_0} \right]. $$

where $\sigma$ is the Stefan–Boltzmann constant ($5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$), $\varepsilon$ is emissivity, $\alpha$ is albedo, $S_\downarrow (1 - \alpha)$ is the net downwelling shortwave radiation, and $L_\downarrow$ is the downwelling longwave radiation [Ek and Mahrt (1991), modified to allow reflected longwave radiation].

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2 The slope value for 4 km in Fig. 6 differs slightly from that in Fig. 5 because the 4-km averages were found directly from the 25-Hz data rather than from the smoothed 1-km data.
Likewise, we replace $T_s - T_1$ with $(T_s - T_0) + (T_0 - T_1)$ to express $G_{sfc}$ as

$$G_{sfc} = K_t \frac{(T_s - T_0)}{\Delta z} e^{-2F_g} + K_r \frac{(T_0 - T_1)}{\Delta z} e^{-2F_g},$$

(3)

where $K_t$ is the soil thermal conductivity, a function of soil composition, soil porosity, and soil moisture, $F_g$ is the green vegetation fraction, $T_1$ is the soil temperature in the first model soil level (0–10 cm), and $\Delta z = 5$ cm is the distance from the surface to the middle of the first model soil level. In this version of HRLDAS, $K_t$ is computed following Peters-Lidard et al. (1998). The coefficient $\exp(-2F_g)$ in (3) describes “insulation” of the ground by overlying vegetation as an analog to the extinction of radiation. In its original form, the exponent as proposed by Choudhury et al. (1987) and applied by Peters-Lidard et al. (1997) was $-\beta\text{LAI}$, where LAI is the leaf area index. This term was changed to $-\beta F_g$ in Ek et al. (2003), with the empirical coefficient $\beta = 2$ based on offline Noah runs.

Finally, the sensible heat flux $H$ is written as

$$H = \rho c_p c_h U (T_s - T_0),$$

(4)

where $\rho$ is the air density, $c_p$ is the specific heat of air at constant pressure, $c_h$ is an eddy exchange coefficient, and $U$ is the wind at the lowest atmospheric model level (i.e., 2 m). In (4), we neglect the small correction (0.02 K) for the adiabatic lapse rate in $T_s - T_0$.

Rewriting (2) and (3) in terms of $H$ using (4):

$$R_{net} = +S_d(1 - \alpha) + \varepsilon L_\downarrow - \varepsilon\sigma T_0^4 - \frac{4\varepsilon\sigma T_0^4}{\rho c_p c_h U} \frac{H}{T_s - T_0},$$

(5)

and

$$G_{sfc} = \frac{K_t(T_s - T_0)}{\Delta z} e^{-2F_g} + \frac{K_r e^{-2F_g}}{\rho c_p c_h U \Delta z} H,$$

(6)

Finally, substituting (5) and (6) in (1b), we obtain the following:

$$\text{LE} + H \left( 1 + \frac{K_t}{\rho c_p c_h U \Delta z} e^{-2F_g} + \frac{4\varepsilon\sigma T_0^4}{\rho c_p c_h U} \right) = S_d(1 - \alpha) + \varepsilon L_\downarrow - \varepsilon\sigma T_0^4 - \frac{K_r}{\Delta z} e^{-2F_g}.$$

(7)

The effects of horizontal variation of $\alpha$ and $\varepsilon$ in (7) are small for the two land-cover classes represented, because their values are so similar: for grassland and cropland, respectively, $\alpha = 0.19$ and 0.20 and $\varepsilon = 0.96$ and 0.985.

To get some idea about the behavior of the terms in (7), we use the results for the control run at 1800 UTC. To simplify, we calculate $K_r$ for each soil type by inserting the appropriate input variables for each soil type into the Peters-Lidard et al. (1998) equations and then find the best-fit least squares straight line ($|R| = 0.99$).
for $K_t$ as a function of volumetric soil moisture $\Theta_1$ in the range 0.15–0.35 to obtain the following:

$$K_t = a_i + b_i \Theta_1,$$

(8)

where $\Theta_1$ is the average for the top 10 cm of the soil and $i$ designates the soil type. In (7), $\Theta_1 = 0.23 \pm 0.03$, $F_g = 0.37 \pm 0.004$, $T_0 = 302.84 \pm 0.49$ K, $T_1 = 298.4 \pm 0.84$ K, and $\rho c_p U = 51.26 \pm 1.41$ W m$^{-2}$ K$^{-1}$. Using $a = 0.98$, $4 \varepsilon a T_0^3 = 6.17 \pm 0.03$ W m$^{-2}$ making the radiation term $4 \varepsilon a T_0^3/\rho c_p U = 0.12 \pm 0.003$. Using $\Delta z = 0.05$ m and (8) to approximate $K_t$, the soil term in the $H$ coefficient on the left-hand side of (7) averages 0.21 ± 0.05, leading to an $H$ coefficient of $1 + 0.21 + 0.12 = 1.33 \pm 0.05$. Even though the individual numbers change along the track, the changes are such that the resulting $H$ coefficient does not vary much. On the right-hand side, the radiation terms total to 735 ± 2 W m$^{-2}$, and the soil term is $-47 \pm 6$ W m$^{-2}$, adding up to 687 ± 8 W m$^{-2}$.

The $H$ coefficient in (7) is positive definite. This implies a negative $\Delta H/\Delta LE$ slope if (i) the $H$ coefficient and the right-hand side of (7) do not vary much horizontally, and (ii) LE and $H$ vary horizontally significantly more than the right-hand side. Under these conditions, which appear to be satisfied on 29 May, HRLDAS supports the observational evidence for $H$ and LE being anticorrelated. Indeed, the $H$ coefficient in (7) could be related to $\Delta LE/\Delta H$. We will return to this topic.

c. The Zilitinkevich coefficient $C$

The one HRLDAS parameter modified in this study was the coefficient $C$ in the Zilitinkevich (1995)$^3$ equation used in the surface layer parameterization to compute the “scaling height” for heat and water vapor $z_{oh}$ from the roughness height $z_{om}$, which in HRLDAS is provided from a lookup table as a function of landcover type. From Chen et al. (1997),

$$z_{oh} = z_{om} \exp \left(-k C \frac{u_a z_{om}}{v}\right),$$

(9)

where $u_a$ is the friction velocity, $k = 0.4$ is the von Kármán constant, $v$ is the kinematic molecular viscosity of air ($\sim 1.6 \times 10^{-5}$ m$^2$ s$^{-1}$), and $C$ is an empirical coefficient set to 0.1 based on comparing model results and field data (Chen et al. 1997). In (9), $u_a z_{om}/v$ is the roughness Reynolds number. In the Monin–Obukhov equations, $z_{om}$ is the height at which the average wind goes to zero, and $z_{oh}$ is the height at which the air temperature $T_{air} = T_s$. Scalar transports at $z < z_{om}$ are assumed to be by molecular processes. It follows that (9) applies strictly for bare soil (e.g., see Zeng and Dickinson 1998); the documented complications

$^3$This expression is similar, except for an exponent of 0.45 instead of a square root, to one in Zilitinkevich (1970), which, according to Zeng and Dickinson (1998) derives from a combination of dimensional analysis and heat transfer experiments.
FIG. 7. Green vegetation fraction $F_g$ scenarios used as input into Noah–HRLDAS sensitivity runs (Tables 2–5). Not shown: extreme or Ex ($F_g = 0.01$) and KA average ($F_g = 0.25$) scenarios.

FIG. 8. Level-1 (0–10 cm) soil-moisture $\Theta_1$ scenarios for runs in Tables 2–5. Not shown, $\Theta_1 = 0.25$. Except for runs 20–22, other soil layers retained values from the control run.
related to the vertical distribution of temperature in vegetation of finite height and the influence of sunlight, wind, and so on (e.g., Garratt 1992; Brutsaert and Sugita 1996; Sun and Mahrt 1995; Sun 1999) are ignored. Thus, the height and density of the vegetation are felt through $z_0$, LAI, and $F_g$. Moreover, Monin–Obukhov similarity (e.g., Paulson 1970) is assumed to apply down to $z_0$. Based on measured $u*$ values, the default value of $C/Hz_{10,05}$ in HRLDAS is equivalent to $z_0/H_{10,02}$ on 29 May, with larger values for smaller $z_0$ (a range of 0.03–0.12 m is used).

4. Results: Sensitivity studies

a. Introduction

In all, 22 HRLDAS runs (Tables 2–5) were conducted to explore the sensitivity of HRLDAS $H$, LE, $T_s$, $T_g$, available energy $H + LE$, $G_{dir}$, and $\Delta LE/\Delta H$ to different input parameters and different values of $C$ in (9); and to improve HRLDAS agreement with observations. For each run, modified values were introduced at 0000 UTC 29 May. The input parameters modified include $F_g$ (Fig. 7), $\Theta_1$ (Fig. 8), $\Theta_2$ ($\Theta$ for 10–40 cm), soil type (Fig. 9), and $z_{om}$. Land use (Fig. 9) and $\Theta$ for the third and fourth soil layers (60 and 100 cm thick, respectively) were left at their control values.

b. Effects of soil moisture and green vegetation fraction

Given the sparse vegetation along the western track, variability in soil moisture should strongly influence horizontal variability in $H$, LE, and $T_s$; this conjecture is supported by Alfieri et al. (2007). On 29 May, the decrease (increase) in LE ($H$) from south to north along the western track is consistent with the observed moister soils to the south (Fig. 10). In the figure, there is an overall northward decrease, with the lowest values at site 3, the highest values at site 2, and $\Theta_1$ lying within the scatter except for site 3. At site 1, however, only the NCAR 5-cm soil moisture is for 29 May: because the soil was too wet to walk on, the site crews had to wait until 30 May to take manual measurements and bring the malfunctioning soil-profiling instruments back online. The soggy soil is consistent with over 80 mm of rainfall there 2 days previous (Fig. 11). From the air, the
southern end of the track appeared to have saturated soils, with puddles up to 100 m in diameter (Fig. 12); indeed, one road looked impassible. We speculate that the HRLDAS soil moisture is too low along the southern part of the track for two reasons. First, the stage IV radar rainfall used for HRLDAS initialization underestimated the 26–27 May rain event by almost a factor of 2 (Fig. 6 in Chen et al. 2007). Second, the “extra” water in the Noah model is not stored in puddles that can either evaporate or soak into the soil; rather it is simply taken out of the model domain as “surface runoff.”

The lack of horizontal variability in $H$, $LE$, and $\Theta_1$ and the too-shallow $\Delta LE/\Delta H$ slope in the HRLDAS control run (Figs. 2–6), combined with the foregoing evidence for a soaked south end of the track prompted the high soil moisture values at the south end of the track for the “extreme,” “dry north,” “wet north,” and “wetter north” scenarios in Fig. 8. The first two listed scenarios were based on the NCAR sites 1–3, while the wet north profiles gave equal weighting to site 3 and the CU site 10.

Since 2002 was a drought year, we also suspected that $F_g$ used in the model was too high. This prompted a family of scenarios (Fig. 7) including the extreme, with almost no vegetation, the “high north” scenario, which allowed for less vegetation at the southern end of the track, and three scenarios (“KA step,” “KA smooth,” and KA average $F_g = 0.25$) with $F_g$ based on King Air NDVI measurements, using the formula in Gutman and Ignatov (1998):

$$ F_g = \frac{NDVI - NDVI_{min}}{NDVI_{max} - NDVI_{min}}, $$

where $NDVI_{max} = 0.52$ and $NDVI_{min} = 0.04$.

From Fig. 13, changes in $\Theta_1$ and $F_g$ from their control values increase horizontal variation in both $H$ and $LE$, but $H$ remains too large and $LE$ too small, while the available energy $H + LE$ remains larger than observed values (Tables 1 and 2). Horizontal variability is induced primarily by imposed $\Theta_1$ variability. This is illustrated by 1) significant horizontal variability in runs 2 and 8, for which $F_g = \text{constant}$; 2) the similarity of runs 8–10 with the same $\Theta_1$ but differing $F_g$; and 3) the fact that changing $F_g$ from 0.38 (run 10) to 0.01 (run 2) while keeping $\Theta_1 \sim 0.4$ at the southernmost point changes $H$ and $LE$ there by only $\sim50 \text{ W m}^{-2}$. The run-to-run variation for runs 2–10 is greatest to the north because of a larger range in input soil moisture (Fig. 8).
The effect of sand at the HRLDAS grid point at \( \sim 36.83^\circ N \) takes the form of excursions in \( H \) and \( LE \) that are reduced for lower soil-moisture values. All of these strategies steepen the \( \Delta LE/\Delta H \) slopes relative to the control run (Table 2). Eliminating points with \( \Theta_1 < 0.1 \), where \( LE \sim 0 \), steepens slopes from around \(-1\) to around \(-1.1\) (not shown), but values remain too shallow relative to observations.

The \( T_s \) simulations in Fig. 14 are more successful, with the runs with the greatest decrease soil moisture toward the north coming closest to reproducing the observed gradient. There is a slight low bias that increases toward the north.

Figures 15 and 16 illustrate the large sensitivity to soil moisture in HRLDAS. In the “zigzag” run (run 12 in Table 4), the soil moisture was subjectively modified so
that peaks and valleys roughly corresponded to the horizontal variation in observed LE (cf. Figs. 2 and 8). Figure 15 (bottom) shows a reasonable match of modeled to observed LE along parts of the track (if we allow for slight dislocation of peaks), but at a price: both \( H \) and \( T_s \) have absurdly large horizontal fluctuations corresponding to the \( H_9004 \) excursions. From the foregoing, the artificially large \( H \) amplitude follows from (7) and is consistent with the too-shallow \( \Delta LE/\Delta H \) slope for run 12 in Table 4.

c. Soil-type sensitivity

Figure 17 shows grassland \( H \) and LE values for run 11 (\( F_g = \Theta_1 = 0.25, \) Table 4) to illustrate the role of soil type in HRLDAS. Aside from the sand point, the effect is fairly small. For \( G_{sfc} \), the magnitudes increase with \( K_n \), with the highest value for sand, and the lowest value for silty clay loam. Consistent with its high \( K_n \), the sandy soil is coolest. The remaining variability is related to horizontal \( T_0 \) variability and the influence of the soil at lower levels.

d. Varying \( C \) in the Zilitinkevich equation [(9)]

From the foregoing, even large changes in \( \Theta_1 \) and \( F_g \), or setting the soil type to sand could not force good agreement between HRLDAS and observations along the whole track. Likewise, from Tables 1–4, the \( \Delta LE/\Delta H \) value (–1.3) closest to observations is for the highly artificial all-sand run in (9) with \( F_g = 0.01 \). Similarly, the leg-averaged coefficient \( \rho c_p c_h U \) in (4) had surprisingly small variation (leg averages 45–52 W m\(^{-2}\) K\(^{-1}\)) for the first 12 runs, despite differences in soil moisture, soil type, and vegetation, which produced large variations in leg-averaged \( T_s - T_0 \) (Tables 2–4). In contrast, observed \( \rho c_p c_h U \) values found from (4) using surface data for the 1700–2100 UTC period averaged about 16.4 ± 2.6 W m\(^{-2}\) K\(^{-1}\) for the surface flux sites—roughly a factor of 3 lower than produced by HRLDAS. Chen et al. (1997) note that the exchange coefficient \( c_h \) is sensitive to (9). Thus, an additional 10 runs were designed to test the effects of \( C \) on \( \rho c_p c_h U \), the fluxes, \( \Delta LE/\Delta H \), and \( T_s \).

Figure 18 shows that the impact of \( C \) on \( \rho c_p c_h U \) and \( \Delta LE/\Delta H \) is significant. In both cases, the best match to observed values is at \( C = 0.5 \). At least for the southern Great Plains during the summer, other researchers have also found that increasing \( C \) from its default value of 0.1 improved results. When Gutmann and Small (2007, manuscript submitted to Water Resour. Res., hereinafter GS) varied \( C \) in the Noah model using IHOP_2002 surface flux site data for dry, sunny days as input, predicted \( T_s \) matched observed values best for \( C \) values between 0.17 and 0.99; although simply using
$C = 0.5$ did not improve $H$ values. Trier et al. (2004) found using $C = 1$ in HRLDAS coupled to WRF produced boundary layer depths, $T_s$ and mixing ratios more similar to those observed than $C = 0.1$, for the June 1998 convection-initiation case they simulated. For comparison, Zilitinkevich (1995) recommends $C = 2$, and using (9) in Zeng and Dickinson (1998) and assuming their exponent 0.45 for $u_z z_{0m}/v$ is not significantly different from 0.5 in our (9) yields $C = 0.33$, but they caution the relationship strictly applies to bare ground.

Figure 19 examines the effect of $C$ for runs with control soil types: $\Theta_s$ and $F_g$. Increasing $C$ decreases $H$, making leg-averaged values closer to observations. The change in LE is small but new values are farther from observations. These changes are accompanied by a significant increase in $T_s$. Consistent with the $T_s$ increase, $H$ remains positive at 0000 UTC for $C = 0.5$ and $C = 1$.
(Table 5). Leg-averaged $T_s - T_0$ values for $C = 0.5$ (run 17, Table 5) and $C = 0.1$ (run 1, Table 2) bracket the observed values (Table 2). The direction and relative size of the $H$ and LE changes are consistent with the trends in Mitchell et al. (2004) when they changed $C$ from 0.2 to 0.05.

From Fig. 20 and altering $\Theta_1$ to better reflect observations and keeping $C = 0.5$, $H$, LE, and $T_s$ become more consistent with observations, with the closest match for the soil moisture pattern we consider closest to observations [wetter north (WWN) $\Theta_1 = 0.4$ to 0.22, run 19]. From Table 5, the higher WWN soil-moisture values lead to a slightly negative $H$ at 0000 UTC ($-25 \text{ W m}^{-2}$), slightly lower than observed ($3 \text{ W m}^{-2}$), but not as negative as those for the $C = 0.1$ runs. Shifting the input data to account for the solar radiation time offset, the 0000 UTC $H$ value for run 19 increases to $+9 \text{ W m}^{-2}$, closer to both observations and the “corrected” value for the control run. Values of $\Delta \text{LE}/\Delta H$ are similar to those observed (Tables 1 and 5; Fig. 5). However, $T_s$ and $H$ are still too high, and LE is too low at the south end of the track. These biases remain even if we shift the data to allow for fetch (less than one grid point, based on results of Song and Wesely 2003) or to correct for flux height changes.

For completeness, $G_{\text{fsc}}$ is plotted as a function of latitude in Fig. 21 for run 19, the one closest to observations. Note a reasonable consistency with the surface data, although there are only three points for comparison.

Three more runs were conducted: (i) run 20, testing the effect of setting $\Theta_2 = \Theta_1$, where $\Theta_2$ is the volumetric soil moisture in the second soil layer (10–40 cm), (ii) run 21, like run 20 but changing $z_{0m}$ for grass from our default value 0.12 m to a smaller value, 0.08 m, and (iii) run 22, like run 20 but with $F_g = 0.25$. Runs 20 and 21 are compared with run 19 in Figs. 22 and 23. From the figures and Tables 1 and 5, the first two runs each improved agreement with observations, but the effect was small. Reducing $F_g$ superficially eliminated the improvements from (i) and (ii), making run 22 more like run 19 in Table 5 and plots (not shown) more like run 19 in Figs. 22 and 23.

5. Discussion

a. Physical significance of changing $C$

The coefficient $C$ in (9) impacts the coupling between the surface and atmosphere: it changes the surface en-
ergy balance and $T_s$ though altering the eddy exchange coefficient $c_h$, as illustrated in Fig. 18. This in turn follows from the Monin–Obukhov similarity flux–profile relationship [(A4) in Chen et al. 1997]. A smaller $c_h$ in (4) reduces the transport of heat $H$ from the surface, resulting in an increase in $T_s$. From (3), a higher $T_s$ translates into more heat flux into the soil; the increase in $G_{sfc}$ with $C$ is obvious from Tables 2–5. How LE should be affected is less obvious: from (1b), the slight reduction of LE with increasing $C$ in Fig. 19 suggests that decreased $R_{net}$ and increased $G_{sfc}$ (both due to higher $T_s$) more than offsets the decrease in $H$. Mitchell et al. (2004) found that changing $C$ from 0.2 to 0.05 increased midday available energy and $H$, with $T_s$ and $G_{sfc}$ decreasing and LE staying about the same, consistent with what we found.

The association of the results here along with those of Trier et al. (2004) and GS with summer conditions in

![Fig. 17. For $F_g = \Theta_s = 0.25$ and grassland points (run 11), the impact of soil type on (a) surface energy budget and (b) skin temperature at 1800 UTC.](image-url)
the southern Great Plains suggests our results might be seasonal. We speculate that lower values of \( C \) might work better for the wintertime, when vegetation is dormant. In the case of dormant grasses, \( F_{g} \sim 0 \) will treat the surface as bare ground in (3), leading to more flux into the soil than if LAI were used. As noted in the foregoing, lowering \( C \) decreases the flux into the soil, perhaps compensating for using \( F_{g} \) instead of LAI in (3).

b. Relationship between (7) and \( \Delta LE/\Delta H \)

In the foregoing, we discussed the possibility that the \( H \) coefficient in (7) could be related to \( \Delta LE/\Delta H \). For the first 19 runs with control soil types, Fig. 24 compares the scatterplot based HRLDAS slope \( \Delta LE/\Delta H \) at 1800 UTC, to the slope based on the \( H \) coefficient in (7) evaluated for the same time using (8) to calculate \( K_{r} \). Points with \( \Theta_{i} < 0.15 \) were not considered for the \( H \) coefficient or the slope. At lower \( \Theta_{i} \) values, (i) the curve used to estimate (8) becomes more nonlinear, and (ii) \( LE \to 0 \), as illustrated for run 2 at the northern end of the track in Fig. 13. These low \( LE \) values lead to a flatter slope on the high-\( H \) end of the \( H \)-versus-\( LE \) plots. In Fig. 24, \( \Delta LE/\Delta H \) remains negative, although the two slopes diverge from the 1:1 line for larger negative values. Predictions would be perfect if the \( H \) coefficient and right-hand side of (7) were constant, suggesting there is more variability in both when the slopes are steeper (and \( C \) is larger, Tables 2–5). Despite this departure, correlation coefficients for the HRLDAS best-fit slope lines remain consistently high.

c. Flux into the soil

From the foregoing, we can make several observations about \( G_{sfc} \). First, other things being equal, \( G_{sfc} \) increases in magnitude as \( F_{g} \) decreases, a result of reduced “insulation” by the vegetation as represented by \( e^{-2F_{g}} \) in (3) (Tables 2–5; Fig. 7). Second, \( G_{sfc} \) is largest for sand, a result of its high thermal conductivity compared to other equally moist soils (Table 3; Fig. 21). And finally, \( G_{sfc} \) increases with \( C \), which contributes to the better match of \( H \), \( LE \), and \( H/LE \) to observations for the \( C/0.5 \) runs (Tables 1–5).

Since we measured \( F_{g} \), \( T_{s} - T_{i} \), \( \Theta \) at 5 cm, NDVI, and \( G_{sfc} \), we can estimate \( K_{r} \) from (3) and (8) as a consistency check. We did this for the two sites (2 and 3) with good data. In the observed case, \( G_{sfc} \) is calculated from the measured flux at 5 cm below the surface by using a correction based on the temperature change and heat capacity of the soil layer between the flux plate and the surface. The heat capacity is estimated from soil water content, soil bulk density, and the density of mineral particles, all of which were measured at each site. Based on surface NDVI measurements and (10), we estimated an \( F_{g} \) of 0.39 at site 2, and 0.36 at site 3—close to the control values.

Using data from 1700 to 2100 UTC in (3), \( K_{r} \) was 1.6 \( \text{W m}^{-1} \text{K}^{-1} \) for site 2 and 1.7 \( \text{W m}^{-1} \text{K}^{-1} \) for site 3. These values correspond to saturated soils at both sites based on (8)—clearly not correct. If we use 0.7 for NDVI_{max} and 0.08 for NDVI_{min} in (10) following Alfieri et al. (2007), \( F_{g} \) reduces to 0.24 and 0.21, respectively, and \( K_{r} \) becomes 1.20 for site 2 and 1.28 for site 3. From (8), these values correspond to \( \Theta_{i} \) of 0.3 at site 2 and 0.2 at site 3, closer to what is observed (Fig. 10). Applying Alfieri’s values in (10) for the aircraft data lowers the “observed” \( F_{g} \) from \(~0.25\) to \(~0.13\), which produces only a minor effect on \( H \) and \( LE \) from the foregoing.
FIG. 19. Effect of the Zilitinkevich coefficient $C$ on $H$, LE, and $T_s$ for control (CN) values of $F_g$ and $\Theta$ (Figs. 7–8): $C = 1$ (triangle; run 16), $C = 0.5$ (square; run 17), and $C = 0.1$ (upside-down triangle; control). Observations: aircraft (diamonds) and surface (squares). HRLDAS: $H$ and LE averaged 1700–2100 UTC and $T_s$ 1800–1900 UTC. Observations: $H$ and LE averaged between 1730 and 2030 UTC and $T_s$ from leg 8 (~1830 UTC).
FIG. 20. As in Fig. 19, but looking at the effects of soil moisture at $C = 0.5$ with control $F_g$ (Fig. 7) for runs 14 (triangle down, $\Theta_1 = \text{DN}$), 17 (connected squares, $\Theta_1 = \text{CN}$), and 19 (triangle up, $\Theta_1 = \text{WWN}$). Input $\Theta_1$ plotted in Fig. 8.
d. Sensitivity of fluxes to $z_{0m}$

In this study, the “grassland” pixels are assigned a $z_{0m}$ value of 0.12 m. Given the sparseness of the grasses along the western track, we thought it would be useful to test the impact of assuming a smaller value.

Thus, we decided to perform run 21, which was identical to run 20 except that $z_{0m} = 0.08$ m for grasslands. In Fig. 22, the grassland points associated with the same soil type (one of two silt loam points and four of eight clay loam points) stand out for runs 19 and 20 as slightly higher for $H$ and lower for LE than the cropland points. Likewise, $T_s$ tends to be higher for the grassland points than for the cropland points in Fig. 23. These excursions, small to begin with, are somewhat reduced, but not eliminated, for run 21.

To understand why the impact is so small, we consider the Monin–Obukhov relationships:

$$U = \frac{u_g}{k} \left[ \ln \frac{z}{z_{0m}} - \Psi_m \left( \frac{z}{L} \right) \right]$$

(11)

and

$$T_s - T_0 = \frac{H}{\rho c_p u_g k} \left[ \ln \frac{z}{z_{0h}} - \Psi_h \left( \frac{z}{L} \right) \right],$$

(12)

where $L$ is the Monin–Obukhov length, $z$ is height above the ground, and $\Psi_h$ and $\Psi_m$ are stability corrections to the logarithmic profiles.

For $U = \text{constant}$ in (11), a decrease in $z_{0m}$ decreases $u_g$. A smaller $u_g$ in (9) produces a larger value of $z_{0h}$. Thus, in (12) a decrease in $z_{0m}$ decreases both $u_g$ and $\ln(z/z_{0h})$, so the two effects compensate. From (9), (11), and observed data, $\ln(z/z_{0h})$ decreased more than $u_g$, leading to a slight decrease in $T_s - T_0$ in (12) and a corresponding decrease in $T_s$ in Fig. 23 for run 21. The effect on $H$ and LE is more subtle, but results in values of LE slightly closer to observations, with little change in $H$.

e. The impact of changing initial conditions at 0000 UTC 29 May

In this study, we view HRLDAS as a tool to provide an initial physically realistic soil moisture profile along the western track. The soil moisture was subsequently modified at the top level to provide different values to explore sensitivity and to get closer to observed values. In run 20, setting the soil moisture at level 2 to the level 1 value improved results, but only slightly. Since roots extend down to level 3, “correcting” soil moisture at that level might improve HRLDAS results a little more. However, observations of the soil-moisture profile indicate that the high values of soil moisture had not yet had time to penetrate to those depths (see Fig. 11 of LeMone et al. 2007a).

It is clear, however, that the imposed soil moisture for the “sand” point was consistently too high. While the HRLDAS control $\Theta_i$ at the sand point at $\sim$36.83°N was drier than the adjacent points (Fig. 8, presumably due to more rapid drainage), the imposed values in the sensitivity runs were intermediate between values on either side. These artificially high values probably exaggerated HRLDAS sand-value departures of $H$, LE, and $T_s$ from the values on either side. Thus, with the exception of the control run, we used “no sand” values in Tables 2, 4, and 5 for parameters strongly influenced by the sand point.

6. Conclusions

Sensitivity studies using HRLDAS to represent the fluxes and surface temperature along the IHOP_2002 western track in the Oklahoma Panhandle revealed that adjusting HRLDAS itself is necessary to achieve good agreement with observations. While modifying the input soil moisture improved horizontal variability, we needed to increase the empirical coefficient $C$ in the Zilitinkevich equation [(9)] for obtaining $z_{0h}$ from its default value of 0.1 to 0.5 in order to achieve a good match to observations. This is an appealing result since the observed value of $\rho c_p u_g U$ relating $T_s - T_0$ to $H$ in (4) at the surface sites also suggested $C = 0.5$. Given this value of $C$, our best estimate of near-surface soil moisture ($\Theta_i$ decreasing from 0.4 at the south end of the track to 0.22 at the north end) led to the best agreement with observations (runs 19–21, Figs. 20–23). However, control $F_g \sim 0.4$ provided slightly better results than $F_g \sim 0.25$ estimated from aircraft data (run 22).
Fig. 22. The $H$ and LE along the western track for run 19, run 20 [like run 19, but with $\Theta_2 = \Theta_1$, where $\Theta_2$ is the volumetric soil moisture at HRLDAS level 2 (0.1–0.4 m)], and run 21 [like run 20, except that $z_0$ (grassland) = 0.08 m]. Observations averaged between 1730 and 2030 UTC; HRLDAS results for 1700–2100 UTC. Control $F_g$ (Fig. 7); $\Theta_1$ = WWN (Fig. 8). Land use indicated for soil types with more than one vegetation type.
and none of the runs succeeded in replicating $H$ and LE at the southernmost points, something we speculate is related to widespread ponding there (Fig. 12) or possibly the neglect of soil albedo. A half-hour shift in input solar radiation data led to too low $H$ at 0000 UTC, but there was negligible effect for the period of aircraft–HRLDAS comparisons (1700–2100 UTC).

When $C$ was increased from 0.1 to 0.5, several things improved. First, the coefficient $\rho c_p c_H U$ from HRLDAS matched the observed values. Second, the flux into the soil increased to values well bracketed by the observations. This reduced the available energy to values close to observed values, with $H$ and LE both closer to observations. Third, the horizontal variability in $H$ and LE as represented by the slope $\Delta LE/\Delta H$, which changed little for different runs with $C = 0.1$, matched observations for $C = 0.5$. And finally, $T_s - T_0$ matched observations reasonably well. The HRLDAS $T_s$ values for runs 19–21 (Fig. 23) were slightly higher than the observations partially because observed $T_s$ was $\sim 1$ K too low and input $T_0$ was $1–2$ K too high.

The slope $\Delta LE/\Delta H$ proved useful not only as an additional variable to guide our sensitivity tests, but it provided a justification for trying to replicate only the larger-scale horizontal trends in the observed fluxes rather than the smaller-scale maxima and minima. Both observations and the HRLDAS equations strongly support an anticorrelation between $H$ and LE horizontally. In addition, the slope defines the ratio of the horizontal amplitudes of $H$ and LE, helping us to understand how
arbitrary horizontal changes in soil moisture to match small-scale LE fluctuations led to unrealistically large variation in $H$ at those scales (run 12; Fig. 15).

In the sensitivity tests, we varied $\theta$ far more than $F_g$ because of the inferred large horizontal $\theta$ range and the suspicion that the control $F_g$ value represented an upper limit in this sparsely vegetated and drought-affected region. Thus, it is not surprising that sensitivity tests showed that the results were most sensitive to $\theta$ for a given value of $C$. Because $C$ was the only thing that would bring modeled and observed variables into agreement, and because of the clear dominance of $\theta$ for the range of parameters examined, we felt we did not need to sort out the nonlinear interactions between the input parameters as done by Niyogi et al. (1999).

We could replicate neither the low values of $H$ or $T_s$ nor the high values of LE at the southern end of the track for realistic initial conditions even with the “best” sensitivity runs and allowing for height differences and fetch for $H$ and LE. We speculate that $H$, LE, and $T_s$ could be better replicated by HRLDAS at the southern end of the track if effects of the observed widespread ponding (Fig. 12) were represented. For example, comparing the aircraft videos to radiometric surface temperature for puddles large enough to be measured by the Heiman radiometer, which looked at a 15° swath (~35 m for flight legs at 65 m AGL), the temperatures were lowered by as much as 5 K relative to nearby ground.

Likewise, accounting for soil radiative properties and their changes with surface soil moisture could yield improvement for this sparsely vegetated area (one could clearly see the color of the soil from the air). For example, soil albedo drops from about 0.3 to 0.4 for dry soils, to about 0.1 for saturated soils (e.g., Idso et al. 1975; Lobell and Asner 2002). Accounting for this effect would increase the available energy at the southern end of the track, potentially steepening the modeled $\Delta$LE/$\Delta H$ slope. However, the effect on $H$, LE, and $T_s$ is unclear.

How general are the results? The fact that we are studying 1 day is mitigated by considering 19 points (or 57 km) along a line with significant horizontal changes in soil moisture, soil type, and vegetation, for several hours. Since many of the same discrepancies exist for the eastern track—too high $H$, too low LE, and too shallow $\Delta$LE/$\Delta H$ slope—we suspect $C$ is too low for that track as well. Moreover, the work of Trier et al. (2004) and GS suggest that $C > 0.1$ might work better than $C = 0.1$. However, all the cited work is for the spring–summer regime in the southern Great Plains; $C = 0.1$ could work better other times of the year or in other locations. Furthermore, other methods such as described in Carlson and Boland (1978) and Zeng and Dickinson (1998) should be considered.

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