The Regional Water Cycle and Heavy Spring Rainfall in Iowa: Observational and Modeling Analyses from the IFloodS Campaign

YOUNG-HEE RYU, JAMES A. SMITH, MARY LYNN BAECK, LUCIANA K. CUNHA, AND ELIE BOU-ZEID
Department of Civil and Environmental Engineering, Princeton University, Princeton, New Jersey

WITOLD KRAJEWSKI
IIHR–Hydroscience and Engineering, The University of Iowa, Iowa City, Iowa

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ABSTRACT

The regional water cycle is examined with a special focus on water vapor transport in Iowa during the Iowa Flood Studies (IFloodS) campaign period, April–June 2013. The period had exceptionally large rainfall accumulations, and rainfall was distributed over an unusually large number of storm days. Radar-derived rainfall fields covering the 200 000 km² study region; precipitable water from a network of global positioning system (GPS) measurements; and vertically integrated water vapor flux derived from GPS precipitable water, radar velocity–azimuth display (VAD) wind profiles, and radiosonde humidity profiles are utilized. They show that heavy rainfall is relatively weakly correlated with precipitable water and precipitable water change, with somewhat stronger direct relationships to water vapor flux. Thermodynamic properties tied to the vertical distribution of water vapor play an important role in determining heavy rainfall distribution, especially for periods of strong southerly water vapor flux. The diurnal variation of the water cycle during the IFloodS field campaign is pronounced, especially for rainfall and water vapor flux. To examine the potential effects of relative humidity in the lower atmosphere on heavy rainfall, numerical simulations are performed. It is found that low-level moisture can greatly affect heavy rainfall amount under favorable large-scale environmental conditions.

1. Introduction

In this paper, we present observational and modeling analyses of the atmospheric water cycle in Iowa during the Iowa Flood Studies (IFloodS) campaign period (April–June 2013), with a special focus on heavy rainfall. The IFloodS campaign period was characterized by extreme rainfall. The “regional scale” of our water cycle analyses is defined by the approximately 200 000 km² rectangular region that contains the state of Iowa. Observational analyses of precipitable water, water vapor flux, and thermodynamic properties focus on this scale.

It has been well documented that flooding events in the central United States are concentrated in late spring and early summer (Wang and Chen 2009; Villarini et al. 2011a,b; Smith et al. 2013) and that water vapor transported from the tropics is an important moisture source for precipitation and thus flooding events (Rasmusson 1967, 1968, 1971; Trenberth and Guillemot 1996; Dirmeyer and Kinter 2010; Lavers and Villarini 2013; Stevenson and Schumacher 2014; Smith and Baeck 2015). The water cycle analysis performed by Dirmeyer and Kinter (2010) reveals that flooding events are often associated with an anomalous transport of moisture from the Gulf of Mexico and Caribbean. The relationships between heavy rainfall and water vapor have received particular attention because of the projected increases in water vapor with climate change (Held and Soden 2006), especially with increasing sea surface temperatures that control low-level water vapor concentrations (Trenberth et al. 2003; Sugiyma et al. 2010). An underlying assumption of procedures used for computing probable maximum precipitation (PMP; see WMO 2009), which is a design standard for high-risk structures subject to flood hazards, is that extreme rainfall scales linearly with precipitable water [see Zhao et al. (1997) and Kunkel et al. (2013) for detailed
development. Kunkel et al. (2013) have examined climate change and PMP, concluding that the “most scientifically sound projection is that PMP values will increase in the future.” Regional climate model simulations in the Kansas–Oklahoma area of Zhao et al. (1997), which were designed to examine the “moisture maximization” assumptions linking precipitable water to PMP, point to pronounced changes in storm structure with increasing humidity and nonlinear relationships between precipitable water and maximum rainfall.

The Great Plains low-level jet has been identified as a key contributor to the meridional moisture transport in the central United States (Mo et al. 1995; Helfand and Schubert 1995; Dirmeyer and Kinter 2010). Helfand and Schubert (1995) demonstrated that almost one-third of the moisture transported into the continental United States is delivered by low-level jets. The low-level jet has a pronounced diurnal cycle, exhibiting strong southerly flows at night. An important aspect of heavy rainfall in the central United States is the pronounced diurnal cycle in the frequency and magnitude of heavy rainfall (Wallace 1975; Baeck and Smith 1995; Dai et al. 1999). It has been shown that the diurnal variation of precipitation in the central United States is characterized by a pronounced midnight to early morning maximum during the spring and summer months (Wallace 1975; Dai et al. 1999; Chen et al. 2009), and that the low-level jet has been closely associated with the enhanced nocturnal precipitation (Higgins et al. 1997; Mo et al. 1997; Trier et al. 2010). There have been previous modeling efforts to reproduce the diurnal cycle of precipitation; however, many global and regional models have experienced difficulties in capturing the nocturnal precipitation maxima, in particular over the Great Plains (e.g., Lee et al. 2008, and references therein). Berbery and Rasmussen (1999) reported that the National Centers for Environmental Prediction (NCEP) Eta model forecast precipitation shows a dry bias over the central United States and that this is in part due to a model humidity bias in the lower atmosphere. Since their work, the performance of the U.S. operational numerical weather prediction models in simulating low-level jets has improved. Lee et al. (2008) found that the simulated diurnal cycle of precipitation over the Great Plains by a general circulation model is sensitive to the choice of convection triggers, especially as they relate to saturation properties of the lower atmosphere.

Previous studies have relied heavily on reanalysis fields to examine the atmospheric water cycle. These analyses inherently rely on model parameterization schemes, model resolution, and data sampling frequency. For example, Ryu et al. (2015) reported that North American Regional Reanalysis (NARR) considerably underestimates the diurnal variation in water vapor flux in the eastern United States, compared to observationally derived water vapor flux analyses. We use a new method proposed by Ryu et al. (2015) for computing vertically integrated water vapor flux using Doppler velocity measurements from Weather Surveillance Radar-1988 Doppler (WSR-88D) radars, precipitable water measurements from the global positioning system (GPS) network, and humidity profiles from the radiosonde network. Our new observationally derived water vapor flux dataset provides much higher temporal resolution (can be subhourly) and vertical resolution (50 m) than model-produced analysis products. The high temporal resolution of our dataset, in particular, can provide insight into the water cycle at diurnal (or shorter) time scales. Hence, we aim to provide a new type of observational-based dataset that can aid in overcoming the limitations of model-produced analyses and/or other types of observational datasets having coarser resolutions.

The objectives of this study are to characterize the regional water cycle focusing on water vapor transport during the IFloodS campaign period (April–June 2013) over the Iowa study region, including anomalies during the 2013 period, and to complement these observational analyses with an examination of the potential effects of saturation properties in the lower atmosphere on heavy rainfall using a numerical model. Questions that motivate the study include the following:

1) How do distributional properties of warm season rainfall over Iowa during heavy rainfall seasons differ from rainfall properties during other years?
2) How does the regional distribution of heavy rainfall in the central United States vary with precipitable water, water vapor transport, and saturation properties of the atmosphere?
3) How does the diurnal variation of water cycle components affect the regional distribution of heavy rainfall?
4) What are the mechanisms by which low-level moisture affects heavy rainfall in numerical model simulations?

The data and methods utilized in this study are described in section 2. In section 3, the rainfall during the IFloodS campaign, atmospheric water cycle, and their diurnal cycles are analyzed. Numerical model results are also presented and analyzed in section 3. Summary and conclusions are given in section 4.

2. Data and methods

In this study, we utilize radar rainfall fields from the NCEP stage IV products (Baldwin and Mitchell 1998;
Lin and Mitchell 2005); precipitable water from the GPS network (Ware et al. 2000); and vertically integrated water vapor flux data computed from WSR-88D velocity–azimuth display (VAD) wind profiles, GPS precipitable water, and radiosonde humidity profiles (Ryu et al. 2015). The locations of key observing systems within the study region are shown in Fig. 1.

Precipitable water [kg m^{-2} or mm (dividing by the density of water, 1000 kg m^{-3}, and multiplying by 1000 mm m^{-1})] can be represented in terms of the vertical profile of water vapor density as follows:

$$ W = \int_0^{z_{\text{TOA}}} \rho_v(z) \, dz, \tag{1} $$

where $\rho_v$ is the water vapor density at elevation $z$ [above ground level (AGL)] and $z_{\text{TOA}}$ denotes the elevation of the top of the atmosphere. In this study, hourly precipitable water time series are constructed from 33 GPS precipitable water stations that are located in and around Iowa (Fig. 1). We interpolate the precipitable water time series onto a regular grid ($\approx 11 \times 15$ km$^2$) using the inverse distance weighted interpolation method with a weighting exponent of 2 and integrate to derive an areally averaged precipitable water (used in Figs. 6 and 9, described in greater detail below). We also derive precipitable water from radiosonde sites at Topeka, Lincoln, Omaha, Davenport, and Minneapolis (used in Fig. 10, described in greater detail below).

The vertically integrated water vapor flux vector $Q = (Q_x, Q_y)$ (where $Q_x$ and $Q_y$ have units of kg m$^{-1}$ s$^{-1}$) is given by

$$ Q_x = \int_0^{z_{\text{top}}} \rho_v(z) u(z) \, dz \quad \text{and} \quad \tag{2} $$

$$ Q_y = \int_0^{z_{\text{top}}} \rho_v(z) v(z) \, dz, \tag{3} $$

where $u$ is the zonal (east–west) component of the wind (m s$^{-1}$; positive eastward) and $v$ is the meridional (north–south) component of the wind (m s$^{-1}$; positive northward). Vertically integrated water vapor flux time series are computed at the locations of WSR-88D radars (Fig. 1) using VAD profiles of the horizontal wind, GPS precipitable water time series (interpolated to radar locations from GPS stations), and normalized profiles of water vapor density derived from radiosonde observations [see Ryu et al. (2015) for details on the procedure]. For the station where the radiosonde observation is not available (i.e., Des Moines), the normalized profiles of water vapor density from other radiosonde observations (Springfield, Topeka, Lincoln, Omaha, Davenport, Aberdeen, Green Bay, and Minneapolis) are interpolated using the inverse distance weighted method. For the stations where radiosonde and radar observations are colocated (or closely located), a spatial interpolation of normalized profile of water vapor density is not applied. In the present study, $z_{\text{top}}$ is set to 4 km AGL because of the reliability limitations of VAD data and because of our focus on water vapor flux, which is dominated by water vapor below 4 km. The water vapor density at 4 km AGL, for example, decreases to about 18% of its surface value. The VAD data availabilities on average for April–June 2013 at Des Moines are about 94% for the layer from the surface to 1 km AGL, 75% for the layer of 1–2 km AGL, and 53% for the layer of 2–3 km AGL.

We use the NCEP hourly stage IV radar rainfall fields. These fields are composite radar rainfall analyses at hourly time scale and on the Hydrologic Rainfall Analysis Project (HRAP) grid [approximately 4-km horizontal resolution; see Cunha et al. (2015) for details]. The hourly stage IV rainfall fields in 2013 include some periods with temporal discontinuities in rainfall fields for 1200 UTC observations. For rainfall analyses during 2013, therefore, the hourly stage IV rainfall at 1200 UTC is replaced with the Hydro-NEXRAD rainfall product (Cunha et al. 2015) when it shows discontinuities.

The Advanced Research version of the Weather Research and Forecasting (WRF) Model, version 3.5.1 (ARW; Skamarock et al. 2008), is used for hydrometeorological analyses of a heavy rainfall period on
26–27 May 2013, with a particular focus on potential impacts of land surface processes. Figure 2 shows the domain configuration. The simulation is initialized at 1800 UTC 24 May and integrated through 1200 UTC 29 May 2013. The results from 0000 UTC 25 May 2013 and beyond are presented in this study. Model physics options used are Noah land surface model, WSM6, Yonsei University (YSU) boundary layer scheme, Dudhia shortwave radiation scheme, and the Rapid Radiative Transfer Model (RRTM) scheme for longwave radiation. Cumulus parameterization is not used for either domain and no data assimilation is used in the model simulation. There are 40 vertical layers in total and 11 layers below 2 km AGL, and the lowest model level is 28 m AGL.

The WRF Model is run twice in this study. In the first run, the model is run with the outermost domain with a 9-km grid spacing and its one-way nested domain (d02 in Fig. 2) with a 3-km grid spacing. In the second run, after the first run is finished, we conduct a simulation with a 1-km grid spacing (d03 in Fig. 2) and its initial and boundary forcings are provided from the simulation of the 3-km gridded domain. The initial and boundary forcings are prepared by using the “ndown” tool in the ARW system. The reason for running the model twice separately is to constrain the initial and boundary forcings for multiple simulations over the innermost domain (d03). To examine the role of low-level moisture that can be greatly influenced by land surface processes, in the finest domain, we conduct two simulations: one is the control simulation and the other is a simulation in which the latent heat from the surface into the atmosphere is not considered (NO LE simulation). In the land surface model, the latent heat exchange (and also the other components of surface energy budget) between the surface and the atmosphere is actually computed, but the latent heat is not released into the atmosphere in the model. So, there are only very small differences in sensible heat flux between the control simulation and NO LE simulation. For the NO LE simulation, only in the innermost domain is latent heat not emitted into the atmosphere. Because the boundary forcing is provided from the 3-km gridded domain that allows latent heat flux from the surface, there is moisture transport through the domain boundaries even in the NO LE simulation. The reason to constrain latent heat flux only for the innermost domain is that we aim to examine land surface effects at the regional scale. We analyze and present results for the innermost domain only.

3. Results

a. Rainfall

Rainfall during the IFloodS campaign exhibits large spatial variability as illustrated by the map of the 3-month rainfall field (Fig. 3). The total accumulated rainfall for the 3-month period is computed as follows:

\[ T(x, y) = \sum_{i=1}^{n_d} \int_0^{24} P_i(t, x, y) \, dt, \]  

(4)

where \( n_d \) is the number of days for the period, \( t \) is the time in hourly intervals, and \( P_i \) is the hourly rainfall (mm h\(^{-1}\)) on day \( i \) at spatial location \((x, y)\). Accumulations greater than
than 600 mm are concentrated in the northeastern part of Iowa, with a maximum accumulation of 782 mm. Rainfall accumulations in areas of northwestern Iowa are 350–400 mm. During the IFloodS campaign, there were multiple operational and experimental systems in place for monitoring rainfall. We use the stage IV rainfall fields because of their coverage, time resolution, and accuracy [see Cunha et al. (2015) for additional details and analyses].

Rainfall over the Iowa study region during the April–June 2013 period is unusually large. Mean rainfall over the study region, which is computed as

\[ T = \sum_{i=1}^{n} A^{-1} \int_{A}^{24} P(t,x,y) \, dt \, dx \, dy, \quad (5) \]

where \( A \) is the areal extent of the study region, for 2013 is 485 mm (Table 1). This accumulation is comparable to the 498 mm accumulation in 2008, when record flooding occurred across the state (Smith et al. 2013; Budikova et al. 2010; Coleman and Budikova 2010). The 2013 accumulation is almost twice the 2012 accumulation; severe drought occurred over the region in 2012 (Hoerling et al. 2014). The 2013 accumulation is 55% larger than the climatological mean. The climatological mean is computed using the National Climatic Data Center (NCDC) rain gauge data for 102 stations over Iowa for a 54-yr period (1960–2013), because stage IV data are only available since 2002. The 54-yr mean accumulation from the rain gauge data is 312 mm. Note that the 3-month accumulations from stage IV data and from the rain gauge data for the period of 2002–13 are very well correlated with an \( R^2 \) of 0.98 and a mean bias difference of 9.6 mm.

Rainfall during the 3-month period is associated with a series of major rain events. We compute the fraction of days with daily rainfall exceeding a threshold of 0.5 mm during April–June for each year. Note that the daily rainfall is computed over the 24-h period ending at 1200 UTC. The 2013 rainfall is characterized by the highest frequency of rain events during the last 12 years (2002–13; Table 1). The fraction of days with rainfall is 0.68 in 2013, and the smallest value, 0.44, occurred during the severe drought year of 2012.

The 2013 April–June period had an unusually large number of days with heavy rainfall (taken here to be daily rainfall exceeding 25 mm) covering large areas of the state. During 2013, there were 56 days with daily accumulations exceeding 25 mm. The number of days for which the area with rainfall accumulations exceeding 25 mm was more than 160, 1600, and 16 000 km² was 48, 36, and 14 days, respectively. The three values in 2013 are larger than the 2002–12 mean values (Table 1). Heavy rainfall events covering an area of 160 km² occurred most frequently in 2013. For heavy rainfall days covering areas of 1600 and 16 000 km², the 2013 period is comparable to 2008 and 2010.

<table>
<thead>
<tr>
<th>Year</th>
<th>Total rainfall [mm (91 days)⁻¹]</th>
<th>Fraction of rain days</th>
<th>No. of heavy rainfall days for given rain area</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2013</td>
<td>0.68</td>
<td>&gt;160 km²</td>
</tr>
<tr>
<td></td>
<td>2012</td>
<td>0.44</td>
<td>48</td>
</tr>
<tr>
<td></td>
<td>2011</td>
<td>0.65</td>
<td>30</td>
</tr>
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<td></td>
<td>2010</td>
<td>0.63</td>
<td>45</td>
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<td>2009</td>
<td>0.64</td>
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<td>2008</td>
<td>0.67</td>
<td>45</td>
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<td>2007</td>
<td>0.51</td>
<td>33</td>
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<td>2006</td>
<td>0.64</td>
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<td>0.58</td>
<td>34</td>
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<td>2004</td>
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<td></td>
<td>2003</td>
<td>0.55</td>
<td>40</td>
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<tr>
<td></td>
<td>2002</td>
<td>0.57</td>
<td>40</td>
</tr>
<tr>
<td></td>
<td>Avg</td>
<td>0.59</td>
<td>42</td>
</tr>
</tbody>
</table>

Fig. 4. Fraction of total rainfall produced by rain rate exceeding the rain rate indicated on the abscissa. The rain rate is separated into 1 mm h⁻¹ bins.
Large rainfall rates account for a somewhat larger fraction of the total rainfall in 2013 than on average. The distribution of rainfall rate for 2013 is compared in Fig. 4 to the 2002–12 period through the fraction of total rainfall produced by rainfall rates exceeding $r$ (mm h$^{-1}$):

$$T(r) = T^{-1} \sum_{i=1}^{n} A^{-1} \int_{0}^{24} P_i(t,x,y)1[P_i(t,x,y) > r] dt dx dy. \quad (6)$$

The comparison of the distribution of rain rates shows that the fraction of total rainfall produced by rainfall rates exceeding 17 mm h$^{-1}$ is larger in 2013 than in 2002–12. This result is consistent with the larger number of heavy rainfall days in 2013 than in 2002–12.

The diurnal cycle of rainfall is prominent during the 2013 period and contrasts somewhat from climatological norms. In Fig. 5, we compare the diurnal cycles of rainfall during 2013 and 2002–12 though diurnal box plots of the probability of rainfall and rain rate. The probability of rainfall is computed as

$$\hat{Z}_i(t) = A^{-1} \int_{A} 1[P_i(t,x,y) > 0] dx dy, \quad (7)$$

and rain rate is computed as

$$\hat{Z}_i^r(t) = \int_{A} P_i(t,x,y)1[P_i(t,x,y) > 0] dx dy. \quad (8)$$

The diurnal cycle in 2013 had the lowest probabilities of rainfall during the early afternoon and two local maxima in probability in the late morning and in the late afternoon, in terms of the median values (Fig. 5a). The climatology of the diurnal cycle of rainfall during 2002–12 also shows low probabilities during early afternoon (Fig. 5b). Unlike 2013, the probability of rainfall during late morning is low in 2002–12. The rain rate in 2013 shows a more pronounced diurnal cycle in which the rain rates are generally high during the nighttime hours with a maximum of 1.1 mm h$^{-1}$ at 0100 LST for the median value (Fig. 5c). Compared to the nighttime, the rain rates during the daytime hours are generally low. The nighttime rain rates exhibit larger variabilities than the daytime rain rates. The rain rates in 2002–12 are also lower during the daytime hours than nighttime hours and show two maxima near dawn and 1800 LST.
diurnal cycle of rain rate in 2013 exhibits a more pronounced nocturnal enhancement than that in 2002–12. The enhanced nighttime probability of rainfall is likely associated with mesoscale convective complexes (MCCs) as they have significant diurnal cycles with nocturnal maxima (McAnelly and Cotton 1989; Villarini et al. 2011a,b).

b. Atmospheric water balance analyses

In this section, we examine components of the atmospheric water balance during the IFloodS campaign period and compare the 2013 observations to observations from other years. A particular focus of analyses presented in this section is the diurnal cycle of water cycle components.

The area-averaged precipitable water derived from GPS observations (Fig. 6) shows a weak but clear diurnal variation with an amplitude of ~1 mm. These results are similar to those found in other regions (Dai et al. 2002; Wu et al. 2003; Hanesiak et al. 2010; Ryu et al. 2015). The diurnal cycle of median precipitable water reaches its maximum during late afternoon (1600 LST). This feature is likely associated with the sharp peak in evaporation rate in the early afternoon (not shown) and low rainfall rates occurring in the early afternoon (Fig. 5c). The diurnal cycle of median precipitable water reaches a minimum during the morning, following the nocturnal period of very low evaporation and the nocturnal maximum in rainfall rate.

The vertically integrated water vapor flux in the meridional direction in the center of the Iowa study region (at Des Moines radar site, Fig. 1) exhibits a strong diurnal cycle (Fig. 7b), while that in the zonal direction exhibit little diurnal cycle (Fig. 7a).

We use the spatially interpolated GPS precipitable water and the spatially interpolated water vapor density profiles from radiosonde observations to compute water vapor flux (see section 2). To ensure that the spatially interpolated values reproduce the water vapor flux, we compare the water vapor flux obtained from the interpolated variables with radiosonde observations at Omaha and Davenport (Fig. 8). For the comparison, only 0000 and 1200 UTC data are used. It is seen that the vertically integrated water vapor flux from the interpolation using GPS and VAD wind data is slightly higher than that from the radiosonde observations in general. This can be to some extent attributed to the VAD wind contamination due to bird migrations (see below and appendix A). Apart from the positive biases, the comparison of vertically integrated water vapor flux still shows very good agreements at both the sites.

The meridional water vapor flux is enhanced in the nighttime compared to daytime (Fig. 7b). Given that precipitable water does not show such a strong diurnal variation, we infer that the pronounced diurnal variation of water vapor flux is principally due to diurnal variation of winds. The enhanced southerly flow (and correspondingly the enhanced southerly water vapor
transport) has been shown in previous studies to be associated with nocturnal low-level jets over the central United States (Bonner 1968; Higgins et al. 1997; Mo et al. 1997; Trier et al. 2010). Our analyses show that enhanced southerly transport of water vapor is associated with the enhanced nocturnal rain rates, as shown in Fig. 5.

It should be noted that the VAD winds can be contaminated by bird migrations. Birds migrate northward over the central United States in spring, preferably in the nighttime to benefit from the southerly winds. We compare the meridional VAD winds and radiosonde winds for the five stations (Topeka, Lincoln, Omaha, Davenport, and Minneapolis) for the period of 2007–14 in appendix A. The VAD winds show positive biases in April–June relative to the radiosonde winds, and we perform simple quality control for VAD winds based on the analysis (see appendix A for the details). The vertically integrated meridional water vapor flux using the corrected VAD winds also shows a clear diurnal variation with a nocturnal enhancement (Fig. A2), which does not change our conclusion.

Precipitable water has long been viewed as a key variable for heavy rainfall (Bermowitz 1975; Maddox et al. 1978; Bretherton et al. 2004). During the IFloodS study period, precipitable water and precipitable water change are only weakly related to rainfall (Figs. 9a,b, respectively). For these analyses, we use areally averaged daily rainfall that is computed for the 24-h period from 1200 to 1200 UTC. We use precipitable water at 1200 UTC for the beginning of the 24-h period, and the precipitable water change is computed for the same 24-h period as is used for the daily rainfall. The correlation between daily rainfall and precipitable water is 0.20 and the correlation between precipitable water change and daily rainfall is 0.10. Large values of precipitable water (e.g., 30 mm, somewhat larger than the 75th percentile value of the 3-month precipitable water) are clearly neither necessary nor sufficient for heavy rainfall.

FIG. 8. Scatterplots of vertically integrated (top) zonal and (bottom) meridional water vapor flux from radiosonde measurements and from GPS and VAD measurements at (a),(b) Omaha and (c),(d) Davenport.
We examine the relationship between rainfall and differential (vertically integrated, hereafter it is omitted) water vapor flux in the meridional direction. The differential meridional water vapor flux is computed as the average of the meridional water vapor fluxes at Topeka and Lincoln (two points to the south of Iowa) minus the meridional water vapor flux at Minneapolis (a point to the north of Iowa). So, a larger value of the differential water vapor flux means stronger southerly water vapor flux in the south than in the north (a weaker northerly water vapor flux is also possible, but it is rather unlikely). Because we take the difference, the influence of bird migrations would be smaller than when taking water vapor flux at a point. The differential water vapor flux has a slightly stronger relationship (correlation of 0.28) with areally averaged precipitation (Fig. 9c) than precipitable water. It is not the case, however, that the largest rainfall accumulations are associated with the largest values of the differential water vapor flux.

Relative humidity of the lower atmosphere plays an important role in determining rainfall amounts for periods of large southerly water vapor flux (Fig. 9d). Large values of rainfall accumulation are associated with enhanced values of relative humidity near the surface. The relative humidity near the surface is referred to as the value at the lowest level of the radiosonde measurements. The correlation between rainfall and relative humidity near the surface is the largest (0.53) among the variables we examine. The correlation between rainfall and mean relative humidity in the lower atmosphere (below 1.5 km AGL) is also relatively high, 0.49 (not shown). If we restrict consideration to periods of large differential water vapor flux (75th percentile corresponding to 258.2 kg m$^{-1}$ s$^{-1}$; the red triangles in Fig. 9d), the correlation increases to 0.67. These results point to the importance of thermodynamic processes controlling relative humidity as central players in controlling heavy rainfall production during periods of strong water vapor transport [for related studies, see Doswell et al. (1996), Junker et al. (1999), Market and Allen (2003), and McCaul and Cohen (2002)]. A number of previous studies with a
focus on seasonal and/or interannual time scales have pointed to the crucial role of water vapor transport (or water vapor convergence) in heavy precipitation (Trenberth and Guillemot 1996; Dirmeyer and Kinter 2010; Smith and Baeck 2015). Our analyses point to the important role of thermodynamic properties in determining heavy rainfall.

We compare the interannual variation in anomalies of precipitable water, relative humidity near the surface, level of free convection (LFC), and geopotential height at 850 hPa from the radiosonde observations at Topeka, Lincoln, Omaha, Davenport, and Minneapolis from 2008 through 2013 (Fig. 10). The anomalies are computed as the deviations from the 6-yr monthly mean values over the five stations. The precipitable water in 2013, in terms of the median, is close to normal even though the 3-month rainfall accumulation in 2013 is second largest followed by 2008 (Table 1). Surprisingly, the median value of 2008 precipitable water is lowest during the 6-yr period. The precipitable water in the other wet year (2010) and normal wet year (2011) are high, but the precipitable water is not a strong
The relative humidity near the surface is highest in 2013, and it is higher in the wet years (2008, 2010, 2011, and 2013) than in the normal year (2009) or in the dry year (2012; Fig. 10b). The LFC in 2013 is lowest [positive anomaly (hPa) means a lower altitude], and LFC in 2012 is highest (Fig. 10c). So, it can be interpreted as favorable (unfavorable) thermodynamic conditions in 2013 (2012) together with the lower (higher) lifting condensation level (LCL) as can be inferred from the high (low) relative humidity near the surface in 2013 (2012). The other wet years also show lower LFC except for 2011. The anomalies of geopotential height at 850 hPa in 2012 reflects the unfavorable conditions for heavy rainfall. The anomalous conditions in 2012, including low absolute and relative humidities and high pressure anomalies, are consistent with findings of Hoerling et al. (2014).

We examine the interannual variation in differential meridional water vapor flux from 2008 through 2013 (Fig. 11). Because most GPS precipitable water data in the study region are available since 2008, we compute the water vapor flux from 2008. It is observed that the differential water vapor flux also exhibits the nighttime enhancement in general, in terms of the median and 75th percentile values. The wet years of 2008, 2010, 2011, and 2013 show larger daily mean values

![Diurnal box plots of differential (vertically integrated) meridional water vapor flux in (a) 2008, (b) 2009, (c) 2010, (d) 2011, (e) 2012, and (f) 2013. The differential water vapor flux is computed as the average of meridional water vapor flux at Topeka and Lincoln minus the meridional water vapor flux at Minneapolis.](image-url)
(greater than about 47 kg m\(^{-1}\) s\(^{-1}\)) than the normal year of 2009 (40.5 kg m\(^{-1}\) s\(^{-1}\)) or the dry year of 2012 (28.2 kg m\(^{-1}\) s\(^{-1}\)). The upper envelope of the differential water vapor flux (here, 75th percentile) is more closely correlated with the 3-month rainfall accumulations rather than the median values. The nocturnal enhancement seen from the 75th percentile values is clearer in the wet years than the dry or normal years. For example, the majority of nighttime 75th percentile values exceed 100 kg m\(^{-1}\) s\(^{-1}\) in the wet years. However, the nighttime 75th percentile values in the dry or normal year do not show such large values and are even comparable to the daytime values. Therefore, from the interannual analyses, it can be argued that a strong southerly water vapor flux (stronger in the south than in the north) should be accompanied with humid conditions for heavy rainfall. This is consistent with the finding from Fig. 9d.

Based on the previous studies of low-level jets and precipitation in the central United States, one can raise a question whether frequent occurrence of low-level jets is associated with large rainfall accumulations. We examine the interannual variation in the frequency of low-level jets at Des Moines (Fig. 12). The conditions/criteria of detecting low-level jet are the same as those used in Wang and Chen (2009), one of which is that the vertical profile of the meridional wind speed depicts a jet core below 700 hPa where the maximum meridional wind speed larger than 10 m s\(^{-1}\) is located. Because we use hourly data, we apply one additional criterion that the low-level jet should last for at least 4 h. The frequency of low-level jet is higher in the nighttime than in the daytime for all years of 2008–13. It is striking, however, that the frequency of nocturnal low-level jet occurrence is not very different from year to year. Nocturnal low-level jets occurred most frequently in

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**Fig. 12.** Diurnal variation in the number of low-level jet (LLJ) occurrence (days) at Des Moines for the 3-month period (April–June) in (a) 2008, (b) 2009, (c) 2010, (d) 2011, (e) 2012, and (f) 2013.
2013 over the period of 2008–13. So, this might provide the favorable conditions for moisture supply from the south. However, the frequency of nocturnal low-level jet occurrence in 2012 is comparable to that in the wet years and is even higher than that in 2008. Note that the strength of low-level jet (defined as the maximum wind speed) also exhibits little interannual variation for the period of 2008–13 (not shown). Our findings are consistent with the findings of Song et al. (2005), who demonstrate that frequent and strong southerly jets do not necessarily lead to more precipitation in a small region in south-central Kansas.

c. Simulation results

From the observational analyses, we found that the relative humidity near the surface under conditions of strong southerly water vapor transport plays an important role in heavy rainfall. The moisture and relative humidity near the surface can be greatly influenced by land surface processes. To answer the question of whether the effects of surface latent heat fluxes have an appreciable impact on heavy rainfall, we use the WRF Model and perform a simulation in which the surface latent heat is not released into the atmosphere (NO LE simulation) as described in section 2. The WRF Model performs reasonably well in simulating evaporation rate, and the comparison with observation-based evaporation rates is given in appendix B.

Figure 13a shows the time series of area-averaged rain rate from observations (stage IV) and the two model simulations for a heavy rainfall period of 25–29 May 2013. The control simulation shows, in general, lower rain rates than the observation for the period from 0000 UTC 25 May to 1800 UTC 26 May 2013, and there is little difference between the control and NO LE simulations during this period. This period is characterized by nighttime rainfall that is initiated around midnight (0000 UTC) when latent heat flux is very low. The control simulation, however, produces larger amounts of rainfall when rainfall is initiated in the afternoon (e.g., approximately 1800 UTC 26 May, 2100 UTC 27 May, and 2100 UTC 28 May) than the NO LE simulation. For example, the daily rainfall for the period from 1800 UTC 26 May to 1800 UTC 27 May reveals much larger rainfall in the control simulation than in the NO LE simulation (Figs. 13c–e). The observations also show heavy precipitation in this period, exceeding 100 mm for 24 h for some regions (Fig. 13b).

The control simulation shows reasonable performance in capturing some features of precipitation such as the heavy rainfall in the central part of Iowa and the peak rain rate. The model cannot, however, reproduce the heavy rainfall in the northwestern part of the domain and the exact timing of peak rain rate (the modeled peak timing is 2 h earlier than the observed one). One possible reason is that the model simulates the location of the system, including the front in a different place. We compared the vertical profiles of temperature and dewpoint temperature at Omaha at 0000 UTC 27 May from the observations and simulation (not shown here), and the comparison showed that the model does not capture the vertical profiles well at Omaha when the observed front was located near Omaha. The vertical profile for the simulation at a point near the simulated front, which is located north of Omaha, was similar to the profile observed at Omaha. It should be highlighted that our aim of conducting the numerical simulations is to examine the effects of land surface evaporation on precipitation under favorable heavy precipitation conditions. Therefore, a more important aspect from this perspective is to obtain reasonable performance in simulating the environmental conditions that are associated with heavy precipitation rather than in perfectly reconstructing the precipitation distribution.

The difference in rain rate between the control simulation and NO LE simulation is attributable to the low-level moisture originating from the surface evaporation because the boundary forcings at the domain boundaries as well as the initial conditions are identical in the two simulations. We compare the thermodynamic characteristics in the region, as an example, at point P (point P is denoted in Fig. 14) that is located on line AB (just north of A; line AB is denoted in Figs. 13c–e). Figure 14 shows LCL for the parcel with the maximum equivalent potential temperature $\theta_e$ at 0000 UTC (=1400 LST) 26 May in the two simulations. It is seen that the LCL in the region roughly below the northwest–southeast diagonal line of Iowa is considerably lower in the control simulation than that in the NO LE simulation. For example, the LCL at point P in the control simulation is 1.2 km AGL and that in the NO LE simulation is 1.3 km AGL. Because the LCL is related to the relative humidity near the surface, this result is consistent with our observational analyses.

Figure 15 compares the skew(Temperature)-log(pressure) (i.e., skew T-log p) diagrams for point P at 2000 UTC 26 May in the two simulations. Because of the evaporation, the mixing ratio is highest near the surface (i.e., 13.4 g kg$^{-1}$ at the lowest model level) in the control simulation, while it is almost constant (e.g., 12.2 g kg$^{-1}$ at the lowest model level) in the lower atmosphere in the NO LE simulation. We compute LFC, convective inhibition (CIN), and convective available potential energy (CAPE) for the parcel with the maximum $\theta_e$. The CIN, LFC, and CAPE at point P in the control (NO LE) simulation are 0.0 (16.0) J kg$^{-1}$, 1.2 (1.7) km AGL, and 2192 (1445) J kg$^{-1}$, respectively. Under favorable conditions for deep convection (unstable conditions with the low-level advection of warm and humid air and with midlevel advection of cool air as seen in Fig. 15), the land surface evaporation can result in lower LCL, LFC, and CIN and higher
CAPE, and can in turn result in larger rainfall amounts. These results are expected and are consistent with previous studies indicating that the larger evaporation from the land surface leads to higher boundary layer $u$, larger CAPE, and lower LCL and LFC (Betts et al. 1996; Eltahir 1998; Findell and Eltahir 2003). Unlike previous studies that mostly focused on triggering and/or frequency of convection, our results focus attention on low-level moisture as a key determinant of heavy rainfall.

Another setting in which low-level moisture plays a central role in heavy rainfall production occurs in the afternoon on 26 May. Figure 16 shows the temporal evolution of rainfall during the afternoon, with very intense rain rates in the control simulation. The maximum hourly rain rate in the control simulation is 92.5 mm h$^{-1}$ at 2300 UTC 26 May and that in the NO LE simulation is 71.5 mm h$^{-1}$ at 0100 UTC 27 May. The rainfall locations in the early afternoon (at
2000 UTC/1400 LST) in the two simulations are similar to each other (Figs. 16a,e), but the storms in the control simulation develop into much stronger storms and produce much larger rainfall accumulations. It should be noted that the small difference in peak rain rate at 0800 UTC 27 May between the two simulations (Fig. 13a) is likely attributed to the storms that are initiated outside the domain and advected from the west boundary to Iowa (not shown). That is, these storm elements likely do not experience the surface-originating water vapor inside the innermost domain when they initially develop.

The environmental conditions for these storms were favorable for MCC development, as Colman (1990) identified the following thermodynamic characteristics (some listed): strong warm air advection at 850 hPa and a sharply defined front associated with a strong horizontal thermal contrast. The large-scale environment for MCC development is also well documented in Laing and Fritsch (2000). Figure 17 shows the horizontal wind and $\theta_e$ fields at the 850-hPa level in the control and NO LE simulations. The wind fields at the 850-hPa level are similar between the two simulations; the patterns of $\theta_e$ fields are also similar in general, but the magnitudes are slightly larger in the control simulation than in the NO LE simulation. The sharply defined front (the $\theta_e$ gradient with high $\theta_e$ to the south and low $\theta_e$ to north) is clearly seen in the northeastern part of Iowa, and a less well defined cold pool near the convection is seen in the
northwestern part of Iowa. It is also seen that the high $\theta_e$ air is associated with the strong southerly winds. The vertical cross sections of wind and $\theta_e$ along line $AB$ (marked in Figs. 13c–e and also Fig. 17) show the high $\theta_e$ advection from the south below about the 2-km level (above mean sea level) in both the simulations (Figs. 18a,d).

The $\theta_e$ vertical cross sections in both the simulations also reveal the frontal zone characterized by the surface-based layer of low $\theta_e$ in the north of the cross section.

As Moore et al. (2003) pointed out by analyzing storm-relative composites, large-scale isentropic ascent is instrumental in lifting of parcels to saturation. The
noticeable difference between the two simulations is that the boundary layer $u_e$ (below approximately 1.7 km) is significantly higher in the south of the cross section when the surface latent heat is released into the atmosphere (control simulation). This low-level high $u_e$ greatly influences the development of deep convection and in turn heavy rainfall. The preexisting storm approaching from the west encounters this high $u_e$ layer at 0000 UTC 27 May (Fig. 18b) and develops into an intense storm with heavy rainfall (Fig. 16d). When the surface latent heat is ignored, the preexisting storm from the west encounters the relatively low $u_e$ layer at 0100 UTC 27 May (Fig. 18f). This results in relatively weaker convection and smaller rainfall (Fig. 16h). Our study stresses that the ingestion of surface-originating water vapor within the boundary layer that meets isentropic lift strongly influences the intensity of convection and heavy precipitation. The important role of low-level moisture in a heavy-rain-producing convective system is also highlighted in the numerical study of Schumacher (2015). It is found that a small change in moisture profile in the near-surface layer (with a precipitable water change only by 1%) leads to a large change in rainfall accumulation, by as much as 15% in terms of the domain-averaged rainfall. This is because when near-surface air is more humid (and thus has larger CAPE, smaller CIN, and a lower LFC), more back-building convection occurs and the convection is more intense. There is a possibility that the modified cold pools in the NO LE simulation, which can be stronger because of the drier near-surface air, can influence the convective activities as demonstrated by Schumacher (2015). Further analysis regarding modified cold pools would be interesting and would be required in a future study.

4. Summary and conclusions

We examine the atmospheric water cycle of heavy rainfall during the IFloodS campaign period, April–June 2013, through observational analyses and analyses based on downscaling simulations using the WRF Model. Water cycle analyses are carried out for the approximately 200,000 km² rectangular domain containing the state of Iowa. Observational analyses are based on spatially and temporally distributed water cycle measurements over the study region; these provide fields of precipitation, precipitable water, and water vapor flux. The principal conclusions are summarized below.

1) Total rainfall during the April–June 2013 IFloodS period was unusually large and was associated with a high frequency of storm days. The total rainfall during the 91-day period of 485 mm was 55% larger than the climatological value of 312 mm and matched only by the record-breaking rainfall and flood period of 2008. The fraction of rainy days (i.e., days with mean daily rainfall over the study region exceeding 0.5 mm) during the 2013 period, 68%, is the largest during the 2002–13 period. In addition to the large frequency of storm periods, there was a slightly larger fraction of total 91-day rainfall during 2013 produced by “high” rainfall rates (exceeding 25 mm h⁻¹).

2) The diurnal variation of the water cycle during the IFloodS study period was pronounced for rainfall and water vapor flux, with modest, but significant, diurnal variation of the area-averaged precipitable water. The diurnal cycle of rainfall during 2013 exhibited a more pronounced nocturnal maximum than appears in the diurnal cycle for the period from 2002 to 2012. The diurnal cycle of water vapor flux exhibits a pronounced nocturnal maximum with a
much stronger diurnal variation of the meridional than the zonal component.

3) We show that heavy rainfall at the 200 000 km$^2$ scale of the study region is relatively weakly correlated with precipitable water, with a somewhat stronger correlation to differential water vapor flux. Extreme values of precipitable water are neither a necessary nor sufficient condition for heavy rainfall, with the same holding for water vapor flux. Thermodynamic properties tied to the vertical distribution of water vapor play an important role in determining heavy rainfall distribution, especially for periods of strong
southerly water vapor flux. Heavy rainfall is associated with the combination of strong southerly water vapor flux and high values of relative humidity near the surface.

4) The precipitable water anomaly in 2013 from the five radiosonde observations near Iowa is close to the average for the period of 2008–13. However, the relative humidity near the surface is considerably higher than normal and the LFC is considerably lower than normal in 2013, highlighting the importance of low-level moisture.

5) The analysis of the interannual variation in meridional component of the differential water vapor flux shows large nocturnal water vapor transport from the south in wet years. Although nocturnal enhancement of the differential water vapor flux is also observed in an average sense in the normal/dry years, the magnitudes are smaller than for wet years. It is found that the frequency of nocturnal low-level jet is similar between wet years and normal/dry years.

6) The model simulation suggests the importance of low-level moisture in heavy rainfall under favorable large-scale environmental conditions. The drier air in the lower atmosphere resulting from no surface latent heat release leads to higher LCL, higher LFC, higher CIN, and lower CAPE, thus decreasing rainfall amount. The low-level moisture is also found to play a key role in heavy rainfall when the water vapor can be supplied by isentropic lifting as it is advected by the southerly flow.

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APPENDIX A

Meridional Wind Comparison between VAD and Radiosonde Measurements

We examine the difference in meridional wind between VAD and radiosonde measurements for the five stations of Topeka, Lincoln, Omaha, Davenport, and Minneapolis. The differential winds (VAD winds minus radiosonde winds) at 1200 UTC (=0600 LST) are compared for the period of 2007–14. The data from the five stations are all taken into account in the comparison, and the values of five grid points in the vertical are aggregated into a representative grid (at the middle level) for the box plots in Fig. A1. With an assumption that radiosonde winds are true winds that are not affected by bird migrations, it is found that VAD measurements overestimate the meridional winds on average in all months of April, May, and June. The magnitude of

FIG. A1. Box plots of vertical profiles of differential meridional winds between VAD and radiosonde winds at 1200 UTC (=0600 LST) in (a) April, (b) May, and (c) June for the period of 2007–14. The data at Topeka, Lincoln, Omaha, Davenport, and Minneapolis are all taken into account. The left and right margins of each box indicate the 25th and 75th percentiles, respectively. The red vertical bar indicates the median, and the blue circle indicates the mean. The whiskers indicate the 5th and 95th percentiles.
differential wind in terms of the mean or median values shows its peak value of smaller than $4 \text{ m s}^{-1}$ at about 0.375 km AGL in May. Given this analysis, we perform simple quality control for VAD winds. The adjustment of VAD winds using the difference between VAD and radiosonde winds at 1200 UTC is applied only after sunset as in Benjamin et al. (2004). The mean values of difference between VAD and radiosonde winds at 1200 UTC for the five stations are taken, and their vertical profiles in April, May, and June are fitted by a sum of sine functions. Then these vertical profiles of the difference in meridional wind are subtracted from the original VAD wind profiles during the time period of 2000–0600 LST when the sun is below the horizon. Figure A2a shows the diurnal box plots of vertically integrated meridional water vapor flux at Des Moines with the bird adjustment. The abrupt change at 2100 LST is smoothed out as compared to the original box plots (Fig. 7b). The diurnal variation in the meridional water vapor flux is still clearly observed in its median values, and thus our analysis and conclusions still hold.

APPENDIX B

Evaporation Rate Comparison

We constructed evaporation rate estimates (hereafter, referred to as observation-based estimates) at 14 locations in Iowa (marked by triangles in Fig. B1) by running the Noah land surface model (Chen and Dudhia 2001) using observed downwelling shortwave and longwave radiation and standard surface meteorological measurements at the agricultural meteorology sites. In other words, the evaporation rates are model results but constrained by observed meteorological forcing. We computed the

![Fig. B2. Time series of evaporation rate of observation-based estimates (black line with circles) and control simulation (red line).](image-url)
estimates at the hourly time scale and interpolated point values to a regular grid using the same procedure employed for spatial interpolation of GPS precipitable water time series.

Figure B2 shows the comparison of Iowa-averaged evaporation between observation-based estimates and simulations. The model overestimates the evaporation on 26 May. However, considering the absolute magnitude of peak evaporation rate (≈0.2–0.3 mm h⁻¹) relative to peak precipitation rate (≈2 mm h⁻¹), it can be said that the evaporation is reasonably well reproduced in the model simulations. The daily mean evaporation rates (averaged over Iowa during 25–28 May) of observation-based estimates and simulated ones are 0.0809 and 0.109 mm h⁻¹, respectively. The mean bias error is 0.0283 mm h⁻¹, and the root-mean-square error is 0.0492 mm h⁻¹.

REFERENCES


