Modulation of the Diurnal Cycle of Warm-Season Precipitation by Short-Wave Troughs

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ABSTRACT
Traveling deep tropospheric disturbances of wavelengths (~1500 km) have long been known to play an important role in the initiation and maintenance of warm-season convection. To date, relatively few studies have formally documented the climatology of short waves and their relationship to the diurnal heating cycle, the topography, and the diurnal cycle of precipitation. Those that did had to rely on low-resolution global analyses and often could not track short waves across mountain barriers. In this study, 10 yr of the (32 km) NCEP North American Regional Reanalysis (NARR) are used to objectively identify and track short waves in the North American domain. Statistics of short-wave span, duration, phase speed, latitudinal extent, and locations of preferred intensification/decay are presented. Some of the key findings from the climatology include that the lee (windward) of mountain barriers are preferred regions of intensification (decay) and short waves show little evidence of a diurnal cycle and can pass a given point at any time of the day. The second part of the study focuses on the role that short waves play in modulating the diurnal cycle of propagating convection east of the Rocky Mountains. Depending on the timing of short-wave passage, short waves may either significantly enhance the precipitation above the mean or completely disrupt the normal diurnal cycle, causing precipitation to develop at times and locations where it normally does not. While short waves play an important role in modulating the mean precipitation patterns their role is considered to be secondary in nature. The diurnal precipitation signature is prominent even when short waves are not present.

1. Introduction
The eastern slopes of the Rocky Mountains are a preferred region of convective initiation during the afternoon hours with subsequent propagation eastward during the evening leading to a nocturnal maximum in the central Great Plains (Maddox 1980; Cotton et al. 1983; Wetzel et al. 1983; McAnelly and Cotton 1989; Augustine and Caracena 1994). The Great Plains nocturnal convection may also develop in situ because of forcing of mesoscale ascent by fronts, troughs, nocturnal low-level jets (LLJ), and drylines (Maddox et al. 1986; Augustine and Howard 1988, 1991; Rodgers et al. 1983, 1985). The LLJ in particular is thought to be a major contributor to the forcing and maintenance of nocturnal convection (Maddox and Grice 1983; Pitchford and London 1962; Augustine and Caracena 1994; Hering and Borden 1962; Arritt et al. 1997; Stensrud 1996; Tuttle and Davis 2006; Trier et al. 2006) through the transport of moisture from the Gulf of Mexico (Means 1952; Helfand and Schubert 1995; Higgins et al. 1997) and convergence at the northern terminus of the LLJ (Maddox and Grice 1983; Augustine and Caracena 1994).

Carbone et al. (2002) and Carbone and Tuttle (2008) examined the statistics of warm-season precipitation events over the conterminous United States (CONUS) using a composited national radar dataset of 4- and 12-yr lengths, respectively. They further clarified the nature of the diurnal cycle between the Rocky and Appalachian Mountains and documented the common occurrence of long-lived propagating convective episodes initiating over the elevated terrain immediately east of the Continental Divide. In diurnal composites the phase-locked propagating signal was dominant whether using a 4- or 12-yr average. The diurnal excitation of propagating convection is prevalent in other continents as well including Asia (Wang et al. 2004, 2005), Africa (Laing et al. 2008),
A necessary condition for long-lived propagating convection is that there be sufficient instability and shear. The large-scale environmental conditions are, to first order, governed by the quasi-stationary Rossby wave pattern. Embedded within the long-wave pattern are deep tropospheric traveling disturbances, often called short-wave troughs (hereafter denoted simply short waves) that have considerably smaller scales and traveling with approximately the midtropospheric flow. Short waves are readily observable on 500-hPa charts and are features on which National Weather Service forecasters have historically focused because of their importance for the initiation and maintenance of convection. Traditionally, forecasters have noted that jet maxima (or jet streaks) associated with short waves induce vertical circulations that destabilize the environment locally and facilitate the development of convection (Beebe and Bates 1955; Miller 1955; Newton 1967; Uccellini and Johnson 1979).

Sanders (1988) performed a comprehensive analysis of moving short waves by manually tracking waves as undulations in the 5520-m contour on the 500-hPa height charts of the Northern Hemisphere National Meteorological Center (NMC) analyses for a period covering October 1976–May 1986 (summer months were excluded from the study). On any given day between 8 and 15 short waves could be identified. Short waves had a mean phase speed of 13 m s\(^{-1}\) and a median duration of 12 days. While a wide distribution of short-wave origins was found, there was a preference for short waves to originate in the lee of high terrain, notably the Rocky Mountains and the mountains north of the Tibetan Plateau. In a later study Lefèvre and Nielsen-Gammon (1995) developed an objective method to track 500-hPa vorticity features in the NMC analysis for the 1969–88 period. They identified 27 469 short waves with a median duration of 4 days and also found the lee of mountain barriers to be favored genesis regions. Regions of favored short-wave termination were found upwind of mountain ranges directly upstream of the genesis regions. They were able to track relatively few short waves over mountain barriers.

Bell and Bosart (1989) did a comprehensive survey of 500-hPa closed cyclone and anticyclone centers over the Northern Hemisphere. Over North America they found preferred genesis regions of summer cyclones along the southern coast of Alaska, the Pacific coastal regions of California and Oregon, the Hudson Bay region, and in the lee of the Canadian Rockies. In the Southern Hemisphere Piva et al. (2008) used 24 yr of the National Centers for Environmental Prediction–National Center of Atmospheric Research (NCEP–NCAR) reanalysis and identified on average 868 short waves per year with a median lifetime of 3.5 days and mean phase speed of 13.6 m s\(^{-1}\).

The relationship between short waves and convective precipitation has been the focus of several studies. Maddox et al. (1979) found that midtropospheric meso-\(\alpha\)-scale troughs were a common factor in triggering and focusing flash flood events. Bieda et al. (2009) investigated the relationship of transient inverted troughs to the diurnal cycle of precipitation over the North American monsoon region of the southwestern United States and northwestern Mexico. While accounting for only 38% of all sample days, inverted troughs days produced the majority of total lightning strikes over Arizona and northern Sonora. In a study of midtropospheric perturbations (MP) of wavelengths 700–1500 km, Wang et al. (2009a) concluded that MPs could be linked to up to 60% of July–August rainfall across the northern plains of the United States. It should be noted that the MPs of the Wang et al. (2009a,b) studies are considered distinct from the short waves being investigated here. The MP vorticity signatures are mostly confined below 400 hPa and are thought to have their origins primarily over the Rocky Mountains. Short waves have deep tropospheric extent and frequently have origins over the Pacific Ocean well west of the North American continent.

Considering both the dominance of the diurnal propagating precipitation signature in the lee of elevated terrain and the relative abundance of short waves (and their apparent preference to develop or intensify in the lee of mountain barriers), it is important to establish the cause-and-effect relationship between the two phenomena. In the first part of this study the climatology of short waves in the North American domain will be documented using a 10-yr period of the NCEP North American Regional Reanalysis (NARR) during the warm season (previous short-wave climatologies have emphasized the cold season). The NARR dataset is considered to be a major improvement upon the NCEP–NCAR reanalysis both in resolution and accuracy (Mesinger et al. 2006). An objective method for detecting and tracking short waves is developed and with the improved regional analysis a better assessment of short-wave climatology (compared to the earlier studies that relied on coarser datasets) is expected. Statistics of short-wave span (distance traveled), duration, phase speed, latitudinal extent, and preferred growth/decay locations will be presented. The second part of the study will incorporate radar data and focus on the relationship of short waves to convection, particularly the role that short waves play in modulating the diurnal cycle of precipitation.
2. Data and methodology

The primary data used in this study are the NARR and the WSI Corporation National Operational Weather Radar (NOWrad) national composite radar reflectivity. The NARR is a dynamically consistent atmospheric and land surface hydrology analysis of the North American domain spanning roughly from the central Pacific to the western Atlantic with spatial and temporal resolutions of 32 km and 3 h, respectively (Mesinger et al. 2006). It incorporates many data sources including sondes, aircraft, surface, satellite, precipitation, and the Comprehensive Ocean–Atmosphere Data Set (COADS) ship and buoy data and is considered to be a major improvement to the NCEP–NCAR global reanalysis. For ease of use the NARR data are interpolated to a regular latitude–longitude grid with a grid spacing of 0.25°.

The precise algorithm for creating the radar composite is information proprietary to WSI but is usually described as the maximum value of radar reflectivity as measured by any Weather Surveillance Radar-1988 Doppler (WSR-88D) at any height in a vertical column. The properties of the product include an approximately 2-km latitude–longitude grid with 15-min temporal resolution at 5-dBZ intervals. The reflectivity values $Z$ (mm$^6$ m$^{-3}$) are converted to a rainfall rate $R$ (mm h$^{-1}$) using a standard $Z$–$R$ relationship ($Z = 300R^{1.6}$).

In this study 10 yr (1998–2007) of warm-season (June–August) NARR and radar data are examined. Starting with the 500-hPa wind field (Fig. 1a) a bandpass filter using the Blackman window (DeFatta et al. 1988) is applied to the $u$ and $v$ components of the wind, thereby obtaining a perturbation wind field. The weights of the Blackman window are defined by $w(n) = 0.42 - 0.5 \cos [2\pi n/(N - 1)] + 0.08 \cos [4\pi n/(N - 1)]$ where $N$ is the sample width and $n$ is an integer index over the $u$ and $v$ data points ($0 \leq n \leq N - 1$). The weights are chosen such that wind perturbations of wavelengths 600–2500 km (determined by the 3-dB cutoff wavelengths of the bandpass filter) are retained. Next, the component of the perturbation wind field that is perpendicular to the large-scale flow is found (Fig. 1b, color-filled contours). The large-scale flow is determined by applying a filter with uniform weights chosen to retain scales greater than 3000 km. Short waves are identifiable as couplets of velocity in the perturbation field (Fig. 1b). Consider an east–west section of the perturbation data centered on the velocity couplet at 51°N, 108°W (location of section is shown by the white dashed line in Fig. 1b). The data along the section closely resemble a sinusoid (open circles of inset of Fig. 1b) and by correlating a sine function of varying wavelengths with the data, the wavelength of the couplet can be found. Using the obtained wavelength a least squares sine function (shown as black sine curve in inset) is fitted to the perturbation data from which the amplitude of the short-wave perturbation is obtained. This procedure is repeated at each grid point in the domain. Only those perturbations having cyclonic circulations are considered. Although the NARR domain extends far enough south to detect easterly waves, westerly waves are the focus of this study.

From Fig. 1c a sense of the scales that are being detected can be seen. For example, while the long-wavelength trough centered near 48°N, 118°W is filtered out by the analysis, smaller-scale features embedded within the trough (45°N, 122°W and 50°N, 110°W) are readily detected having amplitudes of 6.0 and 12.0 m s$^{-1}$, respectively.

Following the procedure of Carbone et al. (2002) the short-wave amplitude and radar-derived rainfall fields are averaged in one dimension (usually latitude) and presented in time–longitude space. Long-lived coherent
short-wave and rainfall events appear as streaks propagating from the west to east. In the Carbone et al. (2002) study statistics of event span, duration, and zonal phase speed were determined from the time–longitude plots. The interest here is to document the life cycle of short waves and their relationship to topographical features and the diurnal cycle of precipitation. Thus we need to carry the analysis of Carbone et al. (2002) a step further to obtain the two-dimensional short-wave tracks as a function of time.

Consider a short-wave analysis at two adjacent times, 0900 (TIME1) and 1200 UTC (TIME2) 6 August 1998 (Fig. 2). At each time short waves are identified through an iterative process where the highest contour of short-wave amplitude that encompasses an area of 200 km$^2$ or more is found. The contours start at 9 m s$^{-1}$ and decrement by 1 m s$^{-1}$ to a minimum of 4 m s$^{-1}$. Short waves with amplitudes less than 4 m s$^{-1}$ are not considered. Next the centroid of each short wave is calculated and a 10$^9$ × 10$^9$ array of the data, centered on each of the centroids, is taken from TIME1. Using Tracking Radar Echoes by Correlation (TREC) the array of data at TIME1 is correlated with similarly sized arrays of data at TIME2 (Rinehart and Garvey 1978; Tuttle and Foote 1990). The location of maximum correlation defines the endpoints of the motion vector for each short wave. The centroid at TIME2 that is nearest the endpoint of a vector is considered to be a continuation of the centroid identified at TIME1. This process continues for all time steps, thus finding the tracks of short waves. When a centroid is not found within a prescribed maximum distance (set to 5°) of a vector endpoint the short-wave track is presumed to terminate. The 4 m s$^{-1}$ threshold, although chosen somewhat arbitrarily, is used to filter out small, short-lived features and is based on statistics showing that the vast majority of short waves traveling 500 km or more were those having amplitudes exceeding 4–5 m s$^{-1}$. Because short-wave features generally have large spatial separation a more complex method of identifying and tracking short waves is not needed.

3. Climatology of short waves

Figure 3 shows an example of time–longitude plots of the short-wave amplitude at the 500- and 300-hPa levels (averaged between 34° and 55°N spanning 90° of longitude for June 1998 and August 1999. It is evident that coherent features having spans of several thousands of kilometers and lasting 4–8 days are common and can be found during both the early and late warm season. Several of the features entered the western edge of the NARR domain (examples can be seen at 160°W on 1, 14, 15, and 22 June) indicating even longer lifespans than can be tracked with the NARR data.

Comparing the 500- and 300-hPa analysis shows that the same features are identifiable at both levels and in most cases the 300-hPa amplitude is stronger. This illustrates that the majority of the short waves analyzed here extend into the upper troposphere and are not the MPs studied in Wang et al. (2009a,b). The vertical extent is further elucidated in Fig. 4 showing a series of vertical sections of short-wave amplitude and total meridional wind for three short waves (locations indicated in Fig. 1c): one located north of the polar jet (point a), one embedded within the jet and of weaker amplitude (point b), and one located in the prime convective region of the central plains of the United States (point c). All three sections show deep vertical structures extending from around 800 hPa to near the tropopause. While there is little evidence of the short waves extending to the surface in Fig. 4, examination of other sections have shown that weak signals can often extend to just above the surface. Although the maximum amplitude of short waves lie between the 400- and 200-hPa levels, we monitor and
track disturbances at 500 hPa because this level has been used in earlier tracking studies and because the amplitude at this level may be more directly relevant to the forcing of lower-tropospheric ascent that pertains to convective initiation and modulation. Note that the section in Fig. 4b also cuts through the southern portion of the short wave in Fig. 4a (seen at 108°W in both panels).

Using the tracking methodology described in section 2, 1378 short waves that had spans (distance traveled) longer than 500 km were identified.¹ This equates to a

¹ To remove short-lived transient features from the analysis, only short waves having spans greater than 500 km are considered.
rate of about 11 week$^{-1}$, demonstrating the relative abundance of short waves during the warm season. Stratifying by month, 466, 460, and 452 short waves were observed in June, July, and August, respectively, indicating a flat distribution of activity over the warm season.

Figure 5b shows the 10-yr-averaged density plot of short-wave tracks in plan view, highlighting those areas that have more frequent short-wave activity. The plot is created by accumulating the number of tracks that pass through grid boxes of size 2°×2°. The dominant path of activity occurs in a west–east band centered between 50° and 55°N stretching from the eastern Pacific Ocean to the Hudson Bay region of Canada. Within this band are two regions of enhanced activity: along the southern coast of Alaska and the Hudson Bay region. Both of these active areas are located just to the north of the mean 500-hPa jet positions (Fig. 5a). A secondary active region extends southward from the primary path along the coasts of Oregon and California. There is little activity over the midlatitudes of the Pacific Ocean west of 130°W.

Figures 5c and 5d show the average short-wave amplitude and intensification rate [m s$^{-1}$ (6 h)$^{-1}$]. The intensification rate at time $t_o$ is computed from the difference between the maximum short-wave amplitude at $t_o + 3$ h and $t_o - 3$ h. There are two primary areas where short waves have their maximum amplitudes (Fig. 5c). One is in a band stretching southeastward from the Aleutian Islands of Alaska to near the Oregon coast. The second is a north–south band from the eastern edge of the Hudson Bay to the New England area of the United States. There is a general area of weaker amplitudes over the Rocky Mountains of Colorado and Wyoming that continues northwestward to the Canadian Rockies.

**FIG. 4.** East–west vertical sections of short-wave amplitude (color filled) and meridional winds (contours) at the locations indicated in Fig. 1c. The first solid and dashed contours indicate 2.0 and −2.0 m s$^{-1}$, respectively. Contour interval is 4.0 m s$^{-1}$. Topography is shown in black at the bottom of each panel.

**FIG. 5.** Ten-year average (June–August 1998–2007) of (a) 500-hPa flow [colored contours represent speed (m s$^{-1}$)] and short-wave (b) track density (No.), (c) amplitude (m s$^{-1}$), (d) intensification rate [m s$^{-1}$ (6 h)$^{-1}$] and (e) phase speed (m s$^{-1}$).
The highest intensification rates (Fig. 5d) occur over the eastern Pacific near the coasts of Oregon and California (indicative of intensification of short waves moving southward along the coast) and in the lee of the Rockies (eastern Colorado/Wyoming, central Montana, and Alberta). The primary area of short-wave decay is noted upwind of the Rockies, especially near the Utah–Nevada border. A second area of decay can be seen in the Midwest. Note that the apparent north–south band of decay along the eastern edge of Fig. 5d is an artifact and is due to short waves moving out of the domain.

Figure 5e shows the average propagation speed of short waves where the warm and cool colors denote eastward and westward propagation, respectively. Propagation speeds are a maximum over the CONUS in a band centered at 47°N, nearly coincident with the mean position of the jet stream during the warm season (Fig. 5a). A simple analysis was done comparing the phase speeds to the 500-hPa flow averaged over the latitudinal extent of the short waves. On average short waves were found to travel 1.2 m s\(^{-1}\) less than the airflow. Thus the steering level is near, but slightly lower than 500 hPa. Note that the dividing line between easterly and westerly propagation is near 30°N.

The zonal span, duration, latitudinal extent, and amplitude short-wave statistics are presented in Fig. 6. The latitudinal extent is calculated from the north–south extent of the 4 m s\(^{-1}\) amplitude contour. The amplitude distribution represents the mean amplitude along each short–wave track.

The span and durations distributions (Figs. 6a and 6b) are near lognormal with approximately 40% and 20% of the short waves having spans (durations) greater than 1400 km (50 h) and 2500 km (100 h), respectively. Median span and duration values are 1040 km and 44 h, respectively. There are several events having spans and durations of 6000–8000 km and 200–250 h, being limited by the NARR domain size. The median latitudinal extent of short waves is 6.8°, but short waves of greater than 8° extent occur about 25% of the time. Recall that the latitudinal extent is an average over the short-wave lifetime and it is expected that at any given time the short-wave extent could be different. The short-wave amplitude distribution (Fig. 6d) has a median value of 9.4 m s\(^{-1}\) and about 20% have amplitudes greater than 15 m s\(^{-1}\) with
values as high as 30 m s$^{-1}$. This indicates that short waves can have significant deviations from the large-scale flow.

Warm-season precipitation exhibits strong diurnal signatures in certain geographic areas (Fig. 7) as seen in a plot of the standard deviation of the diurnal maximum [see Carbone and Tuttle (2008) for a description of how this is calculated]. Over the Rocky Mountains (Fig. 7c) and along the Southeast coast (Fig. 7d) the occurrence of
precipitation is strongly peaked around 2100 UTC owing to heating over elevated terrain and land–sea breeze interactions, respectively. In the Mississippi River area of the United States the combination of precipitation developing locally in response to the solar heating maximum and propagating into the region from the west results in a more uniform temporal distribution of precipitation (Fig. 7b). The corresponding analysis for the short-wave amplitude (Fig. 8) shows a much different picture with the diurnal distribution of short waves nearly uniform at all locations. Even in the location where the distribution is most strongly peaked (southern Colorado; Fig. 8c) the deviation from the mean is only 20%–25%. Thus, a short wave can pass any given point at any time of the day and has no significant phase locking to the diurnal cycle.

Examinations of time–longitude plots of short-wave amplitude often reveal a weakening of the amplitude as short waves traverse high topography and a restrengthening on the lee side. The top panel of Fig. 9 shows several examples of this where the averaging window has been limited to latitudes spanning Colorado and northern Kentucky.
New Mexico, an area where the mean topography has a north–south orientation. The nominal location of the Continental Divide is denoted by the dashed line in Fig. 9 at 106°W. In the three examples denoted by “arrows” there is a considerable weakening (by as much as 70%) of the short-wave amplitude as they moved over the highest topography and a restrengthening downstream. It is obvious in each case that the disjointed maxima on either side of the Continental Divide are part of the same short wave as determined by the tracking method (plus signs indicate longitude–time position of tracks passing over the Continental Divide and plus signs show the longitude–time location of objectively identified tracks.

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4. Relationship of short waves to the diurnal cycle of precipitation

It is well known that precipitation tends to form on the downshear side of a short wave where mesoscale ascent and thermodynamic destabilization are the greatest (Smith and Younkin 1972). Figure 10 shows an example of this relationship for two analyzed short waves—one in the Pacific Northwest and the other over the central plains. In both cases precipitation is observed approximately 200–400 km ahead of the short-wave axis. Also of interest is that although the short waves are of similar amplitudes, the precipitation over the central plains is much more intense because of the moister, more unstable environmental air located there.

From the studies of Carbone et al. (2002) and Carbone and Tuttle (2008), the climatology of the life cycle of warm-season precipitation over the CONUS has been well established. Figure 11 shows a 10-yr average of the diurnal precipitation cycle and the 650-hPa NARR
vertical airflow as a function of longitude (Benjamin et al. 2004). For clarity of presentation the cycle has been repeated for a second day. The main features to note are the development of convection over the elevated terrain east of the Continental Divide (105°FW) near the time of maximum solar heating (2100 UTC), followed by propagation into the central plains during the evening hours (between 92° and 102°FW). West of the Continental Divide (106°FW) and east of the Mississippi River (90°FW) a strong diurnal signature is also present, but there is little evidence of any organized, coherent propagation. Also of note is the late afternoon/early evening (2100–0300 UTC) minimum in precipitation between 95° and 100°FW in the central plains. This is thought to be because of weak midtropospheric subsidence due to the mountain–plains solenoidal circulation (Tripoli and Cotton 1989a,b; Koch et al. 2001; Carbone and Tuttle 2008; Trier et al. 2010). The sloped dashed lines in Fig. 11a represent the average phase speeds of short waves (white) and precipitation (black) in the 33°–46°N latitude band. It is important to note that the phase speed of the envelope of propagating convection (~17 m s⁻¹) is much faster than that of short waves (9 m s⁻¹).

During the afternoon and early evening (2000–2200 UTC) the vertical air motion (Fig. 11b) shows upward motion over the Rockies (~106°FW) and a broad area of weak descent over the central plains (90°–100°FW). During the night and early morning the signs of the vertical motion are reversed in these respective regions. The nocturnal upward motion in the central plains may be the combined result of the mountains–plains solenoid circulation and the convergence at the exit region of the LLJ and convection itself.

While the precipitation cycle over the Rocky Mountains and Great Plains is strongly tied to the diurnal heating cycle, short waves are not and can pass any given location at any time of the day. Thus there will be times when a short wave is in phase with the normal diurnal precipitation cycle (short wave about 200–400 km west of the normal precipitation location) and at other times the short wave may be out of phase (short wave 200–400 km east of the normal cycle). To investigate the impact of short-wave timing on precipitation 48-h diurnal composites of short-wave amplitude, rainfall and 650-hPa vertical winds are generated stratified by longitudinal bands in which short waves are found at 0000 UTC. Three such longitude bands are presented here: 112°–107°, 106°–101°, and 98°–93°FW. Considering the first band, for example, those days that had a short wave passing between 112° and 107°FW at 0000 UTC are noted and a composite is generated for those days. The composites are generated for the 9 h preceding and 39 h after short-wave passage. The latitudinal averaging is preformed between 33° and 46°FN.

Figure 12 shows diurnal composites for the three starting longitude bands. In each composite the short-wave amplitude signal (black contours) is coherent enough to be followed into the second 24-h period and changes in the precipitation patterns (color-filled contours) due to the timing of short-wave passage are clearly evident. Also shown are the NARR 650-hPa vertical winds overlaid with the short-wave amplitude. The apparent weakening of the composited short wave with time is due mainly to dispersion from the phase-locked starting time at 0000 UTC. Figure 12a represents conditions when the short-wave forcing is considered to be in phase with the normal precipitation cycle. Comparing Fig. 12a to the 12-yr average (Fig. 11) it is clear that the presence of a short wave enhances the propagating precipitation signal significantly (0000–1200 UTC, 105°–95°FW of the first 24-h period) with a near doubling of the precipitation from the 12-yr mean. Ascent of about 1 cm s⁻¹ is evident 3°–5° ahead of the short wave. Note also that the
composited phase speed of the precipitation (13 m s\(^{-1}\)) is less than the 12-yr mean (17 m s\(^{-1}\); Fig. 11a) and only modestly larger than the short-wave envelope phase speed (10 m s\(^{-1}\)). During the second 24-h period the composited short wave continues its eastward progression reaching 100°W by 0300 UTC. The normally dry central plains (95°–100°W) during the afternoon/early evening hours become filled with moderate precipitation (up to 200% of mean) because of the proximity of short waves to the west. The forcing from the short wave is strong enough to overcome the convective inhibition (presumably owing to the mountain–plains solenoidal circulation; Tripoli and Cotton 1989a,b) and convection is able to develop at times and locations where it normally does not.

**Fig. 12.** Diurnal composites of (left) radar rainfall and (right) 650-hPa NARR vertical winds overlaid with contours of short-wave amplitude for days that had short waves passing between (a) 112° and 107°W, (b) 106° and 101°W, and (c) 98° and 93°W at 0000 UTC. Composites are shown for the 9 h preceding and 39 h after short-wave passage at 0000 UTC (boldface time label). Reference lines are shown at 105° and 97°W. Contours of amplitude start at 2.0 m s\(^{-1}\) and increment by 0.6 m s\(^{-1}\).
Figure 12b shows conditions when the short-wave axis lies nearly on top of the normal precipitation cycle. In these cases convection still initiates near 105°W, but the main intense propagating precipitation signal is shifted about 5° to the east initiating around 100°W in western Kansas/Nebraska and propagates at a speed (11.0 m s⁻¹) near that of the short waves. As noted in Fig. 12a afternoon convection is apparent in the normally dry central plains starting at 1800 UTC of the first cycle. The area between 95° and 100°W contains ascending motion as opposed to the normal descent (Fig. 12d). Also note that while ascending motion is evident over the Rockies at 0000 UTC (when this region is behind the short waves) it is less than normal (Fig. 11b).

Finally, Fig. 12c shows situations when the short waves are out of phase with the normal diurnal cycle. In these cases the propagating convection signal (being far enough to the west of the short waves) initiates in the normal 105°W location. But propagating faster than the short waves the convection eventually encounters the subsidence region behind the short waves and dissipates. The normal long-lived propagating signal is clearly disrupted. To the east of the short wave relatively intense convection develops around 1800–2100 UTC (85°–95°W) where normally only moderate convection develops (Fig. 11) and is able to survive during the night (in areas where nocturnal convection normally does not occur).

In all three scenarios discussed above the intense precipitation envelopes propagated at speeds significantly less than the 12-yr mean and nearer that of the short waves. Thus, it appears that the forcing of convection in these situations is more tightly coupled to the mesoscale destabilization ahead of the short waves.

Figure 12 showed the diurnal variation of precipitation as a function of longitude for short waves passing three selected locations at 0000 UTC. Figure 13 summarizes the results by showing how the diurnal precipitation varies as a function of the relative short-wave position. This was done for four longitude zones: western Colorado (105°–108°W; WCO), eastern Colorado (102°–105°W; ECO), western Kansas (98°–101°W; WKS), and eastern Kansas (95°–98°W; EKS). To clarify how to interpret these plots consider Fig. 13b for eastern Colorado. It shows the usual diurnal cycle of precipitation with a maximum near 0000 UTC. However, the cycle is modulated by short waves approaching from the west with the precipitation reaching a maximum of about 0.48 mm h⁻¹ when short waves are about 4° to the west. When short waves are to the east of eastern Colorado the precipitation intensity is decreased to values of 0.25 mm h⁻¹. Regardless of the short-wave locations the timing of the maximum precipitation remains relatively unchanged; that is, the dominant forcing remains the heating of elevated terrain. Note that as short waves approach there is a slight slope of the main precipitation band, indicating an earlier onset of rainfall by 1–2 h. Close proximity of short waves (to the west) also results in weak precipitation lasting well into the evening/early morning hours (0600–1500 UTC, 6°–12° to the west).

The situation in WCO is essentially the same of ECO with the precipitation maximizing when short waves are about 4° to the west and a decrease in rainfall when short waves are to the east. The occurrence of maximum rainfall is about 1 h earlier in WCO than in ECO.

Farther east in Kansas, however, both the intensity and timing of the diurnal cycle are strongly modulated by short waves. When short waves are well to the west (15°–20°) the normal diurnal pattern is evident with rainfall peaking at 0700 and 1000 UTC in WKS and EKS, respectively. However, as short waves approach precipitation intensifies dramatically, reaching a peak when short waves are 7°–11° to the west. At these optimal positions the mesoscale destabilization ahead of the short wave and convergence at the exit region of the LLJ combine to produce the most intense rainfall. The timing of peak rainfall is about 2 h earlier than normal. When short waves are even closer (3°–5° to the west) the timing of rainfall changes to a late afternoon/early evening maximum, as opposed to the normal late evening/early morning maximum. After short-wave passage to the east precipitation essentially shuts off. Over the central plains short waves strongly perturb the normal diurnal pattern and are able to overcome the weaker forcing of the mountains/plains solenoidal circulation.

It is evident from Figs. 12 and 13 that short waves can strongly modulate the diurnal precipitation cycle. It would be expected that the degree of modulation would be dependent upon the short-wave strength. A cursory examination of this dependence is done by stratifying the results of Fig. 12b into two categories: those days having short waves with average amplitudes less than and greater than 10 m s⁻¹. This roughly divides the days into two equal populations. The rainfall is integrated over a 24-h period between 90° and 105°W to give an average rainfall over the central plains which is compared to the 12-yr mean. For the low- (<10 m s⁻¹) and high-amplitude (>10 m s⁻¹) days the average rainfall values are 0.22 and 0.31 mm h⁻¹, respectively. These values represent an increase in rainfall by 22% and 72% over the 12-yr mean of 0.18 mm h⁻¹. From these crude calculations it is concluded that short waves of amplitudes less than 10 m s⁻¹ have a relatively small impact on rainfall in the central plains.

Consider now those situations when no short waves are present anywhere in the Rocky Mountain or central plains regions (Fig. 14). The diurnal propagating signal is
still clearly evident and similar to the mean state. Thus the diurnal cycle of precipitation still exists even in the absence of short waves. This is consistent with the findings of Trier et al. (2010) where a 10-day integration of a numerical mesoscale model was done to investigate the role of terrain-influenced flows in organizing summer convection. The model was initialized with a month-long average of environmental conditions, thereby filtering out tropospheric disturbances. It was concluded that while tropospheric disturbances may enhance precipitation, they are not essential for the basic diurnal precipitation event.

5. Summary

Using the NARR 500-hPa wind analysis a 10-yr climatology of short waves during the warm season was presented. Here a short wave is defined as the wind perturbation having scale sizes of 600–2500 km normal to the large-scale wind field. An objective technique for identifying and tracking short waves was developed. Some key findings from the climatology include the following:

- Short waves are plentiful during the warm season, occurring at a rate of about 11 week$^{-1}$ in the NARR domain. Within the CONUS the rate is about 4 week$^{-1}$.
- Of the 1378 short waves identified (having spans greater than 500 km) 40% and 20% had spans (durations) greater than 1400 km (50 h) and 2500 km (100 h), respectively. A number of events had spans and durations of 6000–8000 km and 200–250 h, being limited only by the analysis domain.
Short waves were found to have phase speeds about 1.2 m s\(^{-1}\) less than the 500 hPa, indicating a steering level slightly lower than 500 hPa.

Short waves show little evidence of a diurnal cycle and can pass a given location at any time of day.

Short waves often strengthen (weaken) on the lee (windward) side of elevated topography.

The second part of the study focused on the relationship between short waves and the diurnal cycle of precipitation. The initiation of convection over the elevated terrain near and east of the Continental Divide followed by propagation into the central plains during the night hours has been well documented and in multiyear composites is a prominent signal in the diurnal cycle (Carbone and Tuttle 2008). Traveling deep tropospheric disturbances (short waves) are also known to play an important role in the initiation and maintenance of convection. The role that short waves play in modulating the diurnal precipitation cycle was examined by creating composites of rainfall and short-wave amplitude stratified according to the longitudinal position of short waves at 0000 UTC. Depending on the timing of short-wave passage, short waves may either significantly enhance the precipitation above the mean or completely disrupt the normal diurnal cycle, causing precipitation to develop at times and locations where it normally does not. This is especially true in the central plains where the mesoscale lift associated with approaching short waves can counteract the afternoon mesoscale descent due to the mountain/plains circulation, thereby forcing afternoon convection. Over the elevated terrain of the Rockies, precipitation modulation by short waves is less pronounced, mainly influencing the intensity of precipitation but having little impact on the timing of peak precipitation. It was also noted that convection forming in advance of short waves tends to propagate more slowly than the 12-yr mean and nearer to the phase speed of the short waves themselves. The similarity of phase speeds suggests that the short wave and rainfall are tightly coupled as one would expect in an extratropical cyclone and the destabilization associated with the wave is the dominant process in maintaining convection.

While short waves play a significant role in modulating the diurnal cycle of precipitation, their role is considered secondary to solar heating and circulations driven by heating. This is particularly true for the weaker short waves having amplitudes less than 10 m s\(^{-1}\). The diurnal cycle of precipitation is a strong signature whether a short wave is present or not. The fact that multiyear composites reveal a strong diurnal signature reinforces the idea that short waves are of secondary importance. If short waves were the dominant mechanism in forcing convection, a strong diurnal precipitation signal would not be evident in composites since short waves are evenly distributed in the time domain.

From a forecasting perspective these results are relevant for improving warm-season rainfall prediction. The intensity of rainfall over a wide region depends systematically on the position of short waves relative to the usual diurnal forcing of convection. Further, the response to the presence of a short-wave trough will depend critically on the vertical displacement of lower-tropospheric air induced by the wave and the thermodynamic properties of the air displaced. The fact that the signal of short waves is diminished over the mountains immediately upstream of a strongly conditionally unstable region likely has serious implications for forecast skill because it becomes more difficult to identify and correctly initialize in models those short waves with weakened amplitude and structures complicated by orographic influences of short waves. Wang et al. (2009b) evaluated the forecasts of MPs from the operational North American Mesoscale (NAM) model. It was found that a considerable part of QPF error was due to errors in the timing and location of MPs in the NAM. These results suggest that improved representation of the details of upstream short waves and MPs is vital to correctly anticipate the departure of convection activity from the typical diurnal cycle on any given day.

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