Nocturnal Elevated Convection Initiation of the PECAN 4 July Hailstorm

J. W. WILSON AND S. B. TRIER
National Center for Atmospheric Research, Boulder, Colorado

D. W. REIF
School of Meteorology, University of Oklahoma, Norman, Oklahoma

R. D. ROBERTS AND T. M. WECKWERTH
National Center for Atmospheric Research, Boulder, Colorado

(Manuscript received 15 June 2017, in final form 2 October 2017)

ABSTRACT

During the Plains Elevated Convection at Night (PECAN) experiment, an isolated hailstorm developed on the western side of the PECAN study area on the night of 3–4 July 2015. One of the objectives of PECAN was to advance knowledge of the processes and conditions leading to pristine nocturnal convection initiation (CI). This nocturnal hailstorm developed more than 160 km from any other convective storms and in the absence of any surface fronts or bores. The storm initiated within 110 km of the S-Pol radar; directly over a vertically pointing Doppler lidar; within 25 km of the University of Wyoming King Air flight track; within a network of nine sounding sites taking 2-hourly soundings; and near a mobile mesonet track. Importantly, even beyond 100 km in range, S-Pol observed the preconvection initiation cloud that was collocated with the satellite infrared cloud image and provided information on the evolution of cloud growth. The multiple observations of cloud base, thermodynamic stability, and direct updraft observations were used to determine that the updraft roots were elevated. Diagnostic analysis presented in the paper suggests that CI was aided by lower-tropospheric gravity waves occurring in an environment of weak but persistent mesoscale lifting.

1. Introduction

The Plains Elevated Convection at Night (PECAN) field campaign was conducted in the central plains 1 June–15 July 2015, and its focus was on nocturnal convection initiation (CI) and mesoscale convective systems (MCSs) and their environment (Geerts et al. 2017). Nocturnal CI is difficult to predict, compared to the prediction of daytime CI. This poor skill is likely because of the relative lack of observations above the surface at night (Davis et al. 2003; Clark et al. 2007; Weisman et al. 2008). One of the five scientific focus areas of PECAN was advancing knowledge of the processes and conditions leading to pristine nocturnal CI. While PECAN investigators (Geerts et al. 2017; Stelten and Gallus 2017; Reif and Bluestein 2017) have not provided a consistent definition of “pristine CI,” this case fits their definitions of pristine CI because it was more than 100 km from any other storms and was unassociated with any near-surface convergence lines.

On the night of 3–4 July 2015, an isolated supercell hailstorm developed on the western side of the PECAN study area during intensive observation period (IOP) 18. The IOP was called to study nocturnal elevated convection initiation (NECI) in the western PECAN area. Because of the strong static stability of the nocturnal boundary layer, convection at night tends to be elevated rather than surface-based. For this paper, elevated convection is defined as convection that is not rooted in the boundary layer (e.g., Colman 1990; Bluestein 1993). The storm is classified as a supercell hailstorm based on a radar Doppler velocity mesocyclone, radar reflectivity (maximum 71 dBZ), and an extensive flare echo (discussed later). There were no hail reports, likely
because the storm occurred in a remote area during the night.

Reif and Bluestein (2017) reported that for the occurrence of pristine NECI, important features include the nocturnal low-level jet, midtropospheric moisture maximum, and a midtropospheric warm advection. Other studies have discussed nocturnal initiation being associated with bores and gravity waves (Pitchford and London 1962; Whiteman et al. 2006; Knupp 2006; Koch et al. 2008; Browning et al. 2010; Marsham et al. 2011) or warm advection and mesoscale convergence within the exit region of the low-level jet (e.g., Tuttle and Davis 2006; Stelten and Gallus 2017). Wilson and Roberts (2006) noted that midlevel convergence and conditional instability were major factors in initiating nocturnal storms during IHOP_2002 (Weckwerth et al. 2004).

This current PECAN case was unique in that it was the only case where data were collected on convection initiated more than 100 km from any other storms and without a near-surface convergence line or bore causing the CI. In addition, the forecast for CI provided an opportunity to collect a considerable amount of high-resolution data from a variety of measurement systems near the time and location of the CI. Figure 1 shows the forecasts that were issued by the PECAN forecasters at the 3 July afternoon briefing for the likelihood of CI that night. Forecast discussions from the Goodland and Dodge City, Kansas, National Weather Service offices, as well as from the PECAN forecasters, mention the following phenomena could contribute to the favorable conditions for CI: warm air advection, a short-wave trough, isentropic ascent, elevated convective available potential energy (CAPE), and the veering of the low-level jet. Based on the forecast, instruments including the University of Wyoming King Air (UWKA) research aircraft, fixed and mobile PECAN Integrated Sounding Arrays (PISAs), and an NSSL mobile mesonet (MM1) were deployed in the region of anticipated NECI (Fig. 2); all of these instruments will be discussed later.

This case is of particular interest for the following reasons: (i) it was unique in that the good forecast for NECI provided the opportunity to deploy many PECAN facilities in desirable locations, likely making this the most well-observed case of isolated, nonsurface-forced NECI to date; and (ii) a hailstorm with supercell characteristics occurred. The primary objective of this study was to identify possible CI forcing mechanisms that initiated the storm, which eventually evolved into a hailstorm and later into a north–south squall line. In this paper, we examine only the early stages of CI.

Section 2 shows, from a radar perspective, the initiation and evolution of the convective storms and their relationship to the synoptic and mesoscale environment. Section 3 contains a description of the instruments used in this study and their locations. Section 4 discusses CI observations and provides an analysis of possible CI-triggering mechanisms, and section 5 provides a summary and conceptual model of the initiation phase.
2. Convection initiation and its environment

The evolution of the convective storms on the night of 3–4 July is shown in Fig. 3, as observed by the NCAR S-Pol radar (Lutz et al. 1995) at an antenna elevation angle of 1.7°. The greenish echo is below 5 dBZ and is primarily return from insects and cloud, which will be discussed later. The NEXRADs show one other storm not shown in Fig. 3 that was 160 km northwest of the starting location of the hailstorm in Fig. 3a. CI is at 0405 UTC for the storm that evolves into the hailstorm (red track); this storm will be called CI-H. At 0356 UTC, the maximum reflectivity for the storm to become CI-H was 18 dBZ at a height of 6 km, and by 0405 UTC, the maximum reflectivity was 41 dBZ at 4 km. Figure 3a shows that the hailstorm moved toward the south, which was to the right of the storm-steering winds shown by nearby soundings. The extensive flare echo observed in Figs. 3d and 3e is a common radar artifact echo observed at S band in the presence of large hail (Zrnic 1987; Wilson and Reum 1988). During the flare echo stage, there was a mesocyclone at a height between 1 and 5 km that had a Doppler velocity rotational couplet (Brown et al. 1978) with a differential velocity of 25 m s⁻¹ (not shown).

Seven convective storms formed, and each was concentrated within a 75-km east–west envelope. Only three of these storms reached 40 dBZ. The first of these was the hailstorm that initiated at 0405 UTC, and the storm with the blue track in Fig. 3a (storm CI-2) initiated next at 0526 UTC, 45 km east of the MP4 sounding. The third storm track (yellow) initiated at 0542 UTC in almost the same location as CI-H.

There were no surface fronts or significant surface convergence features at 0400 UTC (Fig. 4), although a cold front had moved through Kansas into southern Oklahoma during the previous 24 h. The synoptic pattern at 700 hPa at 0000 UTC 4 July consisted of a ridge over the western United States, with northwesterly flow and cooler temperatures east of the ridge over much of the central plains, including the PECAN region. A north–south band of high relative humidity at 700 hPa extended from the Rocky Mountains east to the region of CI (Fig. 5a).

Animation of infrared satellite imagery indicates extensive cloudiness over the Southwest United States rotating anticyclonically about a center over the Southwest (not shown). Cloudiness from this region merged with clouds from afternoon thunderstorms over Wyoming, which then moved to western Kansas by nightfall (~0230 UTC).

Comparison of the operational soundings from Denver, Colorado (DNR), Dodge City, Kansas (DDC), and Topeka, Kansas (TOP), at 0000 UTC 4 July with the PECAN MP4 sounding launched at 0300 UTC 4 July indicates that the vertical profiles of equivalent potential temperature θe between 820 and 620 hPa were very similar between MP4 and DNR, much more so than the DDC and TOP soundings to the east, even though MP4 was closer to DDC (see Fig. 5b). Because θe is approximately conserved within air parcels, this suggests that the relatively high θe above 820 hPa at MP4 had similar origins to that of the air near the Rocky Mountains, where it was observed earlier. This point is further advanced in Fig. 6, which is a west–east vertical cross section of horizontal winds and specific humidity. The location of the cross section is provided by the line labeled A–B in Fig. 5a. The location of CI-H (near MP4) is at the eastern edge of the greater depth of high specific humidity.

The 0300 and 0500 UTC MP4 and FP5 soundings (Fig. 7) were located closest to CI-H (Fig. 2). We speculate the MP4 sounding was more representative of the inflow to CI-H. At both times, the most unstable air parcels from MP4 were located between 850 and 800 hPa, where the CAPE for the most unstable 50-hPa-deep averaged parcel increased from 1700 to 1900 J kg⁻¹, as convective inhibition (CIN) decreased from 12 to 3 J kg⁻¹.
The corresponding FP5 soundings (Fig. 7b) had less CAPE and more CIN (not shown) and became more stable with time, whereas MP4 became more unstable with time.

Both the CI-H and CI-2 storms (see Fig. 3a) have attributes of supercell thunderstorms, including a deep mesocyclone (CI-H) and storm motion to the right of the mean flow within the cloud-bearing layer, as determined from the 0500 UTC MP4 hodograph (Fig. 9, red arrow). This latter aspect of supercells arises from the location of the storm updraft being biased by shear-induced vertical pressure gradients (e.g., Rotunno and Klemp 1982), which influences the intensity and longevity of these storms. However, the rightward propagation in the current case is less than that empirically predicted by Bunkers et al. (2000), based on a large sample of right-moving supercells (Fig. 9, blue arrow). This is consistent with the current storms being somewhat weaker and shorter lived than classic supercells.

Mesoscale vertical motion was calculated kinematically for the MP4–FP3–FP2 sounding triangle (Fig. 10a) using the methodology employed in Trier et al. (2017), which is based on Bellamy (1949). Both storm initiations (CI-H and CI-2) occurred near the northern leg of this triangle, and the storms subsequently moved southward through the triangle (Fig. 10a). Though maximum upward motions were weak ($\omega \approx -3 \, \mu s^{-1}$), they were

\[ \frac{Dp}{Dt} \approx \omega \] is the vertical velocity in pressure coordinates; see Holton (1992, 165-175). One $\mu s^{-1}$ at 1000 hPa implies a vertical velocity of approximately $-0.01 \, m \, s^{-1}$. 

FIG. 3. S-Pol radar reflectivity image at an elevation angle of 1.7° showing the evolution of the hailstorm. Note the flare echo in (d) and (e). (a) Storm tracks of all the convective echoes from 0300 to 0700 UTC. The dashed tracks indicate the maximum reflectivity was $<40 \, dBZ$ and solid tracks indicate $>40 \, dBZ$. The time at the north end of the track indicates the time the cell was first observed. Times at the south end of the track are provided only for those three storms that persisted until 0658 UTC. The waves referred to in the text are indicated. An enhanced dBZ scale is presented on the far right. The most outer S-Pol range ring is 150 km.
persistent, and 2-h averaged ascent from 0300 to 0500 UTC extends from 800 hPa up through 550 hPa (Fig. 10b). Mesoscale horizontal convergence and upward motion is common in environments of lower-tropospheric warm advection, which, itself, is inferred from the veering wind profiles, particularly between 850 and 700 hPa in nearby soundings (Fig. 7).

This persistent weak mesoscale ascent from 0300 to 0500 UTC is broadly consistent with both cooling and increases in relative humidity from 800 to 500 hPa at MP4 (Fig. 7a), which was located ~90 km west of the triangle center, where the diagnosed average vertical motions are valid. The moistening of the lower part of this layer (~800–680 hPa) during the 2 h, with concurrent decreases in CIN (Fig. 8), facilitated triggering of CI. Meanwhile, cooling in the upper portion of this layer from 600 to 500 hPa (Fig. 7a) contributed to the CAPE increases observed for most parcels originating from levels below 680 hPa (Fig. 8).

In summary, soundings near the western edge of the PECAN domain exhibit a deep elevated moist layer that facilitates nocturnal CI. This moisture was the eastern extension of an elevated moist layer originating along the Colorado and Wyoming Front Range of the Rockies. Weak, but persistent, mesoscale ascent near the CI location made this layer even more favorable for permitting initiation or sustenance of deep convection during the late evening. However, an adjacent colder and drier air mass within a contrasting northerly airstream east of the synoptic midtropospheric ridge (Fig. 5a) may have limited the eastward progression of new CI. In section 4, we examine possible finescale triggering mechanisms for the observed nocturnal CI.

3. Instrumentation

The full complement of instruments available during PECAN is discussed in Geerts et al. (2017). For this study, soundings were used from six fixed PISA soundings (Holdridge and Turner 2015; Vermecsh 2015; Clark 2016; UCAR/NCAR Earth Observing Laboratory 2015, 2016b,c) and four mobile PISA soundings (Knupp 2015; Wagner et al. 2016b; UCAR/NCAR Earth Observing Laboratory 2016).
Laboratory 2016d). MS3 was operated by Colorado State University. The following other instruments were utilized for this study: S-Pol (UCAR/NCAR–Earth Observing Laboratory 2016a), MP3 Doppler lidar (Wagner et al. 2016a), TWOLF (Truck-Mounted Wind Observing Lidar Facility) Doppler lidar (Bluestein et al. 2014; Reif et al. 2016), MM1 (Waugh and Ziegler 2016), UWKA aircraft, the University of Oklahoma Shared Mobile Atmospheric Research and Teaching (SMART) C-band radars SR-1 and SR-2 (Biggerstaff et al. 2005; Biggerstaff 2016), and the Dodge City (KDDC) and Goodland, Kansas (KGLD), NEXRADs. Based on the forecast, facilities were deployed near CI-H (Fig. 2); note in particular that TWOLF and the MP3 Doppler lidars were very close to the CI-H location. This storm initiated directly over TWOLF, but unfortunately, TWOLF was inoperative from 0325 to 0407 UTC, precisely when the storm was developing directly overhead. By 0407 UTC, the developing storm had moved southeast of TWOLF. A second cluster of facilities was deployed about 150 km farther south (partially shown in Fig. 2). The only data from this southern array shown in this paper are gravity waves observed by the SR-2 mobile radar.

The storm initiated within range of the NCAR S-Pol radar that was doing both PPI and RHI scans. Soundings were available from the eight PISA sites and MS3 shown in Fig. 2 and were launched at 2-h intervals (0300, 0500, and 0700 UTC). The UWKA flew from 0245 to 0720 UTC in the western PECAN area, where storm initiation took place. UWKA instruments used were the up-looking backscatter Wyoming Cloud lidar, a downward-looking compact Raman lidar (Wang et al. 2011), and in situ probes to measure atmospheric state parameters. The UWKA mostly flew at an elevation of roughly 2.15 km MSL. The primary leg of interest was flown from west to east between 0408 and 0432 UTC. Convection initiation occurred 25 km north of the track at 0405 UTC. Other flight legs helped identify the location of a wind-shift line to be discussed later. Particularly important were Doppler lidar observations from TWOLF and MP3 to help determine vertical wind profiles and cloud base and to detect atmospheric gravity waves. In addition, there was a mobile mesonet that drove back and forth on an east–west road from about 0130 to 0650 UTC, recording winds, humidity, and temperature. The road was directly below the east–west UWKA track shown in Fig. 2.

4. Observations of CI and analysis of possible CI-triggering mechanisms

a. Cloud origin and characteristics

1) RADAR CLOUD ECHO

Figure 11 shows PPIs and RHIs from S-Pol at 0336 UTC, about 30 min prior to CI-H. In Figs. 11a and 11b, the northwest–southeast-oriented echo beyond the 75-km range ring will be referred to as “cloud echo.” The RHIs from S-Pol show convective, cell-like features that have reflectivities between $-1$ and $-9$ dBZ and differential radar reflectivity $Z_{DR}$ around 0 dB (not shown). Radar-observed cloud bases were about 3.5 km MSL, and the tops were about 5.4 km MSL. There was no precipitation falling from this cloud echo.

Detection of cloud without precipitation (i.e., detecting cloud water droplets) is generally thought unlikely with S-band radar at long ranges. Knight and Miller (1993) have discussed that S-band radar can detect nonprecipitating cumulus clouds via Bragg scattering, which is caused by strong moisture gradients along the edges of clouds. It is unknown if the cellular-type echo in Fig. 11c is from Bragg or Rayleigh backscattering. Gossard (1990; see his Figs. 3.7, 3.8) showed that a radar with the sensitivity of S-Pol observing echoes between $-1$ and $-9$ dBZ and differential radar reflectivity $Z_{DR}$ around 0 dB could detect large cloud water droplets ($\sim 50\mu m$) at far ranges. Because $Z_{DR}$ is near 0 (not shown), it could be from either Rayleigh or Bragg backscattering; regardless, it is from cloud and not precipitation, thus the name “cloud echo.”

Essentially all the other echoes visible in Fig. 11 have large $Z_{DR}$ values typically associated with insects (Wilson et al. 1994; Drake and Reynolds 2012). Thus, it is most likely that S-Pol observed an area of small,
nonprecipitating cumulus clouds with bases at about 3.5 km and tops at about 5.4 km MSL near the CI-H location (red plus symbol in Fig. 11a). S-Pol shows that after 0336 UTC, the uniform reflectivity structure of the cloud echo, seen in the PPI in Fig. 11a, disappears and becomes much more discrete and cellular in nature (Fig. 3). The half-power beamwidth of S-Pol was on the order of 1–2 km in the region of the cloud echo, so it would not be able to identify individual convective cells with widths on a smaller scale. It is then likely that the uniform pattern of reflectivity of the cloud echo in Fig. 11a was the result of a field of small cumulus clouds having individual diameters of \( \sim 2 \text{ km} \). The PPIs and RHIs during the following 30 min show that this field of small cumulus clouds broke down into progressively fewer but larger cumulus clouds. Where CI-H occurred, the sizes of the convective cells are relatively large, with diameters \( > 5 \text{ km} \) (Fig. 3b at 0355 UTC). The growth of the clouds in the cloud echo is discussed further in section 4c(2).

2) CLOUD ORIGIN

Figure 12 shows the IR4 infrared satellite cloud images\(^2\) at about 50-min intervals in the PECAN area. Notable is the presence of northwest–southeast-oriented cloud bands moving from the northwest that gradually increase the cloudiness over the western PECAN area. The S-Pol cloud echo is visible by 0300 UTC. This cloud echo is outlined in Fig. 12d and closely follows the satellite infrared cloud image. Satellite imagery indicates the cloud echo is shallow and not under cirrus cloud. At this time, the infrared temperatures of the cloud tops were mostly above freezing in the area of the S-Pol observed cloud echo; however, the IR cloud-top temperature was between \(-1^\circ\) and \(-5^\circ\)C near the CI-H location, indicating higher tops in that region. The soundings (Fig. 7) show the freezing level was about

\(^2\) Wavelength 10.2–11.2 \( \mu \text{m} \); pixel size 4 km
4.5 km MSL, which, combined with the satellite cloud-top temperatures and cloud-base measurements from S-Pol, indicate that the cloud echo was shallow and, except in the CI-H region, above freezing.

The satellite images in Fig. 12 indicate numerous large areas of cold cloud tops; however, mosaics of the NEXRADs show that only a few of the cold cloud-top regions were associated with rain of any intensity. Examination of these clouds over time shows they originated in Wyoming from afternoon mountain and plains thunderstorms. The coldest cloud tops, just to the southwest of the PECAN domain, are associated with active thunderstorms along the cold front mentioned earlier. The layers of relatively high dewpoints in the MP4 0500 UTC sounding between 6.0 and 7.0 km MSL (Fig. 7) suggest layers of upper-level cloud, which is similar to the FP5 0500 UTC sounding above 9.0 km MSL. It is likely that many of the clouds observed by satellite were anvil remnants of cumulonimbus from Wyoming.

The cloud echo initiated between 0245 and 0330 UTC along the northeast edge of one of the northwest–southeast-oriented cloud fingers that was advancing from the northwest (Fig. 12). Initiation of convective echo in advance of anvils has been discussed by Knight et al. (2004) and Fovell et al. (2006); however, it is unlikely the processes they discussed are applicable in this case (anvil rain and gravity waves in advance of a squall line). The coarse time and space resolution of the satellite IR data made it difficult to follow the evolution and possible cause of the generation of this low-level cloud.

3) CLOUD BASE, TOPS, AND UPDRAFT ROOTS

Of interest is whether the CI-H storm updraft was elevated or surface-based. The closest soundings to the CI-H location are MP4 (82 km) and FP5 (90 km). These are likely the most representative soundings of the conditions where CI-H occurred. At 0330 UTC, the IR satellite data show that neither one of these soundings is located within the cloud echo (Fig. 12d). The FP5 0300 UTC sounding is released in a region of no cloud, and the MP4 sounding in a region of high cloud. Based on the MP4 0300 UTC sounding, it is likely there is no low cloud at MP4.

The FP5 and MP4 0300 and 0500 UTC soundings show shallow surface-based inversions or stable layers (Fig. 7). The mobile mesonet vehicle that drove back and forth below the east–west leg of the UKWA and just south of the CI-H location recorded a temperature of 23°C and a dewpoint of 16.5°C (relative humidity of 67%) closest to the location of CI-H (Fig. 13). These temperatures agree closely with the observations from the 0300 UTC MP4 sounding. Because of the near-surface stability, surface air would need forced lifting for at least 500 m to reach the lifted condensation level (LCL) and develop a convective cloud (Fig. 7a). Neither the mobile mesonet nor the S-Pol showed any near-surface convergence lines that could have supplied a lifting mechanism for the surface air.

The 0500 UTC MP4 sounding, which is the most unstable sounding, was used to estimate possible cloud bases. A parcel lifted from 1.4 km (400 m AGL) would reach the lifted condensation level at 2.2 km MSL, and with about 15 J kg⁻¹ of CIN to overcome would reach the level of free convection (LFC) at 2.8 km MSL. As shown in Fig. 8, MP4 50-hPa-deep parcels lifted from
about 1.7 km MSL (840 hPa) would have only 3 J kg\(^{-1}\) of CIN to overcome to reach their LFC. Between 2.1 and 3.4 km MSL, there were individual points that were saturated or nearly saturated (Fig. 7a). Thus, the MP4 sounding suggests that parcels lifted from above the stable layer would have cloud bases located anywhere between 2.2 and 3.4 km MSL.

Measured cloud bases within the cloud echo were roughly 3.4 km MSL from S-Pol (Fig. 11c), 3.1 km MSL from the UWKA cloud lidar (not shown), and between 3.1 and 3.6 km MSL from TWOLF [(section 4c(2)]. The cloud-base measurements are then in the upper height range, suggested by lifted parcels obtained from the MP4 soundings (2.2–3.4 km MSL). Because the lowest parcel-estimated cloud base of 2.2 km MSL is associated with a parcel lifted from 1.4 km MSL (400 m AGL), and the observed cloud bases are higher (between 3.1 and 3.4 km MSL), it is most likely that updraft roots were above 400 m AGL. These updraft roots for CI-H will be discussed further after examination of the updrafts observed by TWOLF.

It was shown that the S-Pol cloud echo was shallow. As convective clouds grew above 4.0–4.5 km MSL, they likely started entraining dry air, which was evident in the soundings starting above a height of 4.0 km MSL (Fig. 7). Using an entrainment rate of 0.1 (e.g., Romps 2010), Trier et al. (2015) determined that simulated CAPE for the most unstable lifted parcel (MUCAPE) was reduced approximately in half in a southern plains dryline environment with large, undiluted MUCAPE and very dry mid- and upper-tropospheric conditions, similar to the current case (Fig. 7). This left a sufficient amount of remaining CAPE in their simulations, wherein supercell storms developed near the dryline. Thus, it is unlikely that dry-air entrainment alone can explain the relatively small number of deep convective storms in the current case. However, when this effect is combined with the large 800–600-hPa vertical shear in the layer near and above cloud base (Fig. 7), which itself is detrimental to the early stages of CI, and the fact that the mesoscale forcing in this case (Fig. 10b) was not strong, it is not surprising that relatively few deep convective storms occurred.

b. Winds and thermodynamic stability

Insight into the winds and stability in the vicinity of CI-H is provided in Fig. 14. Shown are the wind, temperature, and dewpoint data from the 0300 UTC PECAN sounding releases plotted on constant pressure surfaces (850, 790, 700, and 600 hPa). Doppler lidar winds from MP3 and TWOLF at 790 and 700 hPa are also included.

The easterly and southeasterly winds at 850 hPa at the more central sounding sites shift to northerly winds at 790 hPa, resulting in a wind-shift line (see dashed line in Fig. 14b). At 700 and 600 hPa (Figs. 14c,d), all winds have shifted to westerly or northerly. In Fig. 15, the depth of the low-level easterly and southeasterly winds is plotted based on the radiosonde and Doppler lidar data. The depth of the easterlies increasing toward the west is apparent in Fig. 15 as well as in the S-Pol RHI of Fig. 11d. Also, it was shown in Fig. 6 that the increasing height of the top of the surface-based layer with southeasterly winds was a large-scale phenomenon associated with the sloping terrain. S-Pol agreed with the height on the eastern side showing the
height of the wind shift east of the radar between 1.6 and 1.9 km MSL (not shown).

Summarizing to this point, there is a ~2.0-km-deep layer of nonprecipitating cloud that has a base at about 3.5 km MSL embedded in light west-northwest flow. The most representative sounding (MP4) of conditions likely to occur in the CI area shows that the 2.0–3.5-km MSL layer was nearly saturated with conditionally unstable lapse rates. The CIN in this layer is small (Fig. 8), and the layer has persistent weak mesoscale upward motion (Fig. 10b), conditions that are favorable for CI. At least 1 km below cloud base, the westerlies shift to southeasterlies. The height of the top of the southeasterly wind layer beneath the cloud slopes upward, reaching a height of 2.6 km MSL on the far western side of the PECAN area, compared with a height of 1.6 km MSL on the far eastern side (Fig. 6). So the entire cloud is above the southeasterly winds. The convective cloud updraft roots were likely elevated above 400 m AGL.

c. Storm triggering

While the environment does have sufficient instability for initiation of convective storms, a dominant triggering mechanism is not obvious. Two possible triggers are investigated: an elevated wind-shift line (Fig. 14b) and gravity waves.

1) WIND-SHIFT LINE

At its closest point, the UWKA passed 25 km south and 25 min after CI-H. The location of this leg is shown in Figs. 11a and 11b; the flight altitude was 2.15 km MSL (~790 hPa). The time of this leg is from 0408 to 0432 UTC. Figure 16 is a time series of wind direction, wind speed, mixing ratio, and vertical motion on this west–east
The wind shift from about 180° to 300° was observed near −100.4° longitude, or 70 km west of S-Pol. CI-H occurred ~120 km west of S-Pol. This wind shift is at a lower height than the cloud echo. The wind-shift line on the 790-hPa (2.15 km MSL) surface that the UWKA flew through is where the 790-hPa surface intersects the sloping depth of the easterly and southeasterly winds.

Time series of observations along the west–east flight leg of the UWKA show the CI region is associated with local maxima in horizontal wind speed, upward vertical velocity, and water vapor mixing ratio that were located ~50 km west of the wind shift (Fig. 16). In contrast, there was no distinct maximum in vertical velocity and mixing ratio that would be favorable to CI near the wind-shift line.

2) GRAVITY WAVES

The S-Pol scans using 0.5° and 1.0° antenna elevation angles show waves with horizontal wavelengths of about 5 km, whose phase lines are oriented along 290° to 110° and move from the southwest at about 6 m s⁻¹. The waves can be faintly observed near the radar in Fig. 3a at an elevation angle of 1.7°. These waves are more visible at 0.5° and 1.0° elevation angles. SMART R-1, SMART R-2, and the KDDC WSR-88D also observed the waves and showed similar orientation, wavelength, and movement. An example is shown in Fig. 17 from SMART R-2 that was located 130 km south-southwest and was upstream from where CI-H occurred (see Fig. 2). The waves appear as alternating bands of...
enhanced reflectivity; the scatterers are most likely insects. The SMART R-2 image at 0425 UTC in Fig. 17 was chosen only as a visual example of the waves being observed at S-Pol, SMART R-1, and KDDC prior to, during, and after initiation CI-H. The waves in Fig. 17 were similar to those observed by S-Pol, with wavelengths of about 4 km, and were moving from the southwest at about 5 m s$^{-1}$. Because all the radar ranges are over 100 km from the CI-H location, the beamwidth was about 2 km wide, and the lowest elevation angle (0.5°) was at a height of at least 1.5 km, making it unlikely to observe waves directly over the initiation location of CI-H.

The vertical velocity panel in Fig. 16 showed the UWKA also observed the gravity waves while flying at an altitude of about 2.15 km MSL (~1.2 km AGL). Taking into account the roughly 30° angle between the airplane and the wave crests, the wavelength was about 5 km, similar to that observed by the radars. The vertical velocities were <1 m s$^{-1}$, except when flying through the edge of the developing CI-H radar echo. West of the developing hailstorm, the amplitude of the vertical velocity from crest to trough was only $|\Delta w| \sim 0.2–0.3$ m s$^{-1}$. However, the amplitude of these wavelike vertical velocity oscillations increased significantly to $|\Delta w| \sim 0.5–1.5$ m s$^{-1}$ east of the developing storm. In contrast, the other parameters generally do not show consistent wavelike features.

The vertical pointing lidars also show what appear to be wavelike features (updraft–downdraft pairs) in the vertical velocity time series (Fig. 18). Vertical velocities are observed by TWOLF and MP3 for a several-hour period near the time of CI-H. The bottom panel in Fig. 18 shows expanded images of vertical velocity measurements from TWOLF between 0227 and 0324 UTC; recall that TWOLF lost data at 0324 UTC. The TWOLF expanded panel has been edited at the height of cloud base (roughly 3.5 km) to simplify viewing because the data above cloud base are rapidly attenuated and unreliable.

The alternating pairs of updrafts and downdrafts that pass over the MP3 lidar about every 10–15 min roughly agree with the 5-km wavelength and 6-m s$^{-1}$ movement of the waves observed by the radars. The lack of appreciable vertical tilt in most of these wavelike features (Fig. 18a), together with their spatial and temporal coherence, suggests they are most likely vertically trapped or “ducted” gravity waves, as shown in images and discussed by Durran (1986; see his Fig. 20.3), Gossard and Hooke (1975), and Nappo (2002).

Possible mechanisms contributing to the vertical trapping of waves in Fig. 18a are deduced from an analysis of a simplified version of the Scorer parameter (e.g., Nappo 2002):

$$\hat{\ell}^2 = \frac{N^2}{(U - c)^2} \frac{d^2 U/dz^2}{U - c},$$

in which layers of $\hat{\ell}^2 < 0$ prohibit vertical propagation of gravity waves. FP3 is located 92 km northeast of MP3 and 50 km north of S-Pol (Fig. 2), where wavelike signatures are also evident. Note that FP3 is also situated in the environment of consistent higher-amplitude vertical velocities continuing east of the shaded CI-H region along the UWKA flight track (Fig. 16). Because it is the closest available sounding to MP3 and S-Pol,
FP3 is used to calculate $\ell^2$ from (1), in which $N^2 = (g/\theta_v)(\partial \theta_v/\partial z)$ is the static stability, $c = 6$ m s$^{-1}$ is the approximate horizontal phase speed of the waves, and $U$ is the wind in the direction of horizontal wave propagation, which is oriented south-southwest–north-northeast from 210° to 30° azimuth.

Vertical variations in the magnitude of the vertical shear along the direction of the horizontal wave propagation [term 2 on the right side of Eq. (1)] are the dominant contributor to two prominent trapping layers located beneath 2 km MSL (Fig. 19). The strongest trapping layer, which prohibits downward propagation of waves from above, is located directly above a ~150-m-deep, surface-based nocturnal inversion and results from a midlayer $U$ maximum in the presence of weak static stability (Fig. 19). In contrast, the second significant trapping layer from 1.55 to 1.8 km MSL, which prohibits upward propagation of waves, arises from the vertical shear of $U$ becoming more negative with height within this layer (Fig. 19). This second trapping layer marks the location of the abrupt transition from the lower-tropospheric synoptic southeasterlies to strong northerlies above (Fig. 6; −100° to −98°) located east of the midtropospheric ridge (Fig. 5a).
Together, the two trapping layers constitute a lower-tropospheric wave duct (Fig. 19) that allows the waves to propagate horizontally. The depth and location of the wave duct in Fig. 19 is broadly consistent with the wavelike vertical velocity signatures in the MP3 lidar data (Fig. 18a), though some of the updrafts in the lidar data extend up to 1 km above the second trapping layer. This could indicate some vertical “leakage” (i.e., only partial trapping) of the waves and/or spatial and temporal variations in the structure of the wave duct.

It is assumed that the horizontal scanning radars and vertical pointing lidars observed the same gravity waves, though at different times and locations. Analysis was also conducted at surface stations MP3, MP4, and MM1 of pressure, wind direction, and wind speed to detect gravity waves. While they showed weak perturbations, they were not consistent among variables and stations. MP3 was an exception, where an analysis based on Gossard and Hooke (1975) did show a wave propagation direction from 233° consistent with the direction observed from S-Pol and SR2.

The gravity wave vertical motions shown by the vertical-pointing Doppler lidars (Fig. 18), were relatively weak, with \( |w| < 1 \text{ m s}^{-1} \). This is consistent with the vertical velocities measured by the UWKA. However, because of the low CIN, even weak updrafts could trigger CI. Applying the simplifications of parcel theory, the minimum updraft required to lift air to its LFC is \( w_{\text{min}} = \sqrt{2 \text{CIN}} \) (e.g., Markowski and Richardson 2010, p. 194). CIN of 3–4 J kg\(^{-1}\) in 50-hPa averaged layers centered between 835 and 725 hPa (Fig. 8) would require \( w_{\text{min}} \) of \(-2.4\)–2.8 m s\(^{-1}\), which are larger than the maximum updrafts measured by lidar prior to the loss of data. The MP3 lidar continues to measure similar weak updrafts during the data loss at TWOLF. Nevertheless, the similarity of the theoretical required and actual measured updraft magnitudes, combined with the broad region over which persistent weak mesoscale ascent is diagnosed, suggests the possibility that the gravity waves could be sufficient to trigger isolated CI somewhere within this region.

The cloud-base height at TWOLF decreased from about 3.6 to 3.1 km MSL until signal was lost at 0324 UTC. Starting just before 0315 UTC and ending just before data were lost, a longer period of weak updraft develops below cloud base, just above 2.5 km MSL (Fig. 18). Note that this updraft is not directly connected to the gravity wave updraft. It is speculated that mesoscale gradual ascent (Fig. 10b) and horizontal advection (Fig. 6), with time, contributed to gradually increasing the relative humidity and decreasing the height of cloud base. It is further speculated that the longer duration of updraft between 0315 and 0322 UTC above 2.5 km MSL coincides with the development of the updraft roots responsible for CI-H. It is likely during the period of no TWOLF data that the vertical velocity increased significantly as CI-H developed overhead.

In section 4a(3), based on sounding analysis and observed cloud bases, it was concluded that the roots of the updrafts forming the cloud echo were above 400 m AGL. Updrafts in Fig. 18 from the Doppler lidar vertical time series show that updrafts with the gravity waves start slightly above the first recorded gate. The first recorded gate for MP3 is 15 m AGL, and for TWOLF, it is 400 m AGL. The base of the likely cloud updraft at TWOLF between 0315 and 0323 UTC is near 2600 m. Because of (i) potential problems with vertical sampling of slanted updrafts, (ii) distinguishing cloud updraft roots from gravity wave updraft roots, and (iii) unknown performance of the lidars in sampling vertical velocities near the ground, the base of the cloud updrafts is difficult to estimate from the vertical-pointing lidar data.

The observed waves are most likely trapped gravity waves having their origin from an MCS in the Texas Panhandle (cf. Fig. 4), located about 300 km to the southwest of S-Pol. We specifically examined the possibility that gravity waves from the advancing cloud from Wyoming interacted near TWOLF with the gravity waves..
waves from the southwest to initiate CI-H. While waves from the Wyoming storms cannot be totally ruled out, they were not apparent.

5. Summary and conceptual model

a. Initiation of the hailstorm

There were no surface fronts near the area of the initiation of the hailstorm (CI-H), but importantly, there was a north–south band of enhanced moisture at 700 hPa that stretched from the Colorado/Wyoming Rockies to western Kansas and Nebraska. An analysis of the MP4 sounding from 0300 and 0500 UTC showed a layer from 800 to 620 hPa of increasing relative humidity and weak, but persistent, mesoscale ascent. By 0300 UTC, a field of small cumulus clouds was observed by radar (cloud echo) to be developing on the northeast side of a high-level cloud (anvil) observed by satellite. The leading edge of the anvil cloud had its origin from afternoon thunderstorms in the mountains of Wyoming. After 0336 UTC, the S-Pol radar showed the uniform field of small cumulus evolving into fewer but larger-scale convective cells, which were likely the result of updrafts having increasing vertical velocities with increasing horizontal sizes. The Doppler lidar positioned under the growing hailstorm showed that the cloud base descended from 3.6 to 3.1 km MSL by 0324 UTC, after which time data were lost.

The prolonged weak (diagnosed) mesoscale lifting contributed to gradually increasing the humidity and
decreasing the CIN, which facilitated development of small, shallow cumulus by 0300 UTC. It is speculated that this environmental moistening, combined with weak (directly measured) gravity wave updrafts, enabled more persistent, larger-scale updrafts with roots near 2.5 km MSL (~1.5 km AGL) to develop by 0315 UTC. Soon, these updraft roots extended downward, ingesting larger CAPE air from below; this enabled sufficient buoyancy to maintain an updraft capable of overcoming factors detrimental to CI, including entrainment of dry midtropospheric air and large vertical shear extending from cloud base to a few km above.

There was not an observed dominant triggering mechanism for this hailstorm. Instead, its initiation (CI-H) seemed dependent on a variety of subtle factors in an environment that was becoming increasingly more susceptible to CI. Once initiated, it was not surprising the storm developed into a supercell, given the strong vertical wind shear from the surface to 6 km and moderate-to-large CAPE. There were no other storms within 100 km prior to CI-H, making this a pristine [as defined by Stelten and Gallus (2017) and Reif and Bluestein (2017)] nocturnal elevated convection initiation event.

b. Conceptual model of CI

To highlight the features of this CI event, a conceptual vertical cross section is shown through a 12-km approximate northeast–southwest line centered on the CI-H location, valid for a time just prior to CI-H (Fig. 20). The larger cloud centered at zero represents the cloud that developed into the hailstorm. The dark red horizontal line in Fig. 20 at about 2.2 km indicates the shift in wind direction between easterlies below and westerlies above. This is the wind-shift line that the UWKA flew through on a west–east leg when flying at 2.15 km MSL. The conditions along the wind-shift line were unfavorable for CI, so it was not a factor in triggering CI.

The light brown shading in Fig. 20 above 4 km represents the region of dry air. The black, wavelike features represent the trapped gravity waves, but it is unknown if these waves existed above cloud base. Note that above roughly 2 km MSL, the winds are westerly, and below 2 km, they are easterly. Particularly important for CI in
this case is the light green layer. This is the layer of decreasing CIN and increasing relative humidity that is in the layer of westerly winds. We have not added cloud updrafts to Fig. 20 because of the uncertainty in separating gravity wave updrafts from cloud updrafts. We did speculate that the cloud updraft roots initially started near 2.5 km MSL and lowered as CI-H intensified.

The forecaster discussion in the introduction mentioned five factors that could contribute to nocturnal CI on the night of 3–4 July: warm-air advection, short-wave troughs, elevated CAPE, isentropic ascent, and the low-level jet. Of these factors mentioned by the forecasters, warm-air advection and elevated CAPE clearly played a role. In particular, the soundings showed veering winds with height and warm-air advection on the 700-hPa map (Fig. 14c). There were no obvious short waves on the 850-, 700-, or 500-hPa synoptic maps. Elevated CAPE was shown to exist in Fig. 8. Isentropic ascent was only briefly assessed. The low-level jet was only starting to develop by the time of CI, and there was no obvious convergence associated with it.

c. Needed observations

Given the unprecedented large number of high-resolution observations, what additional observations would be needed to better determine the evolution and triggering of the hailstorm? The new GOES-16 satellite, with its greatly improved pixel resolution and frequency of observation, would have been very useful for examining possible causes of the of the low-level cloud along the northeast side of the anvil cloud. Certainly, vertically pointing Doppler lidar data from TWOLF up to the time of CI-H would have provided critical additional information on cloud-base height, gravity waves, and updraft evolution. Several additional Doppler lidars would also be desirable. Two major CI unknowns are (i) what caused the lowering of the cloud base prior to CI-H, and (ii) what were the specifics of the 3D air motions in its vicinity? A water vapor DIAL located near and upstream of TWOLF would have provided the needed water vapor information. Two or three radars positioned to collect dual- or triple-Doppler radar winds prior to CI would provide 3D wind fields and the possibility of observing interacting gravity waves. Sensitive Doppler radars capable of observing cloud echo and clear-air return would be required. The cloud echo was \(-10-0\, \text{dBZ}\), and the insect clear-air return was 0–10 dBZ. With the exception of S-Pol, few radars, particularly those with shorter wavelengths, would be able to observe the nonprecipitating cloud at distances more than 10–20 km. Observation of the insect echo would generally be limited to 50 km. Bragg echo would not be possible with X-band radars and only marginally possible with C-band radars. The best chance of obtaining the

![FIG. 19. The Scorer parameter $\ell^2$ [see Eq. (1)] and vertical profiles of virtual potential temperature $\theta_v$ and the wind in the direction of horizontal wave propagation $U$ calculated using the 0500 UTC 4 Jul FP3 sounding (see Fig. 2 for location).](image-url)
necessary winds and gravity waves would be Doppler lidars and multi-Doppler radar, including S-Pol and the most sensitive of the mobile radars. However, this would require the assistance of Mother Nature (March 1998) because S-Pol is not mobile.

Acknowledgments. The authors thank all participants of PECAN for their hard work and long nights during the field campaign. The expert forecasting led by William Gallus (Iowa State University) on 3 July calling for pristine CI made this study possible. Appreciation is extended to Doug Nychka and Kate Young (NCAR) for developing a new more flexible sounding plotting package; Russ Schumacher (CSU) for sharing a high-resolution simulation of this case from the CSU version of the WRF Model; David Ahijevych (NCAR) for his help in the construction of Fig. 4; Manda Chasteen (University of Oklahoma) for her assistance in collecting TWOLF data and Chris O’Handley (Simpson Weather Associates) for his help in processing TWOLF data; Sean Waugh and Conrad Ziegler (NSSL) for their a posteriori recalibration of the wind direction sensor that enabled corrected winds from the NSSL mobile mesonet (MM1); and Bob Sharman and Teddie Keller (NCAR) for informative discussions on trapped gravity waves. We thank Morris Weisman (NCAR) for a very useful internal review and Laura Hutchinson (NCAR), who provided an internal review from the perspective of a student. The paper further benefited from the constructive comments of Conrad Ziegler and two anonymous reviewers. The first author would particularly like to thank Dan Megenhardt (NCAR) for his expert help in displaying the many PECAN datasets into one integrated display. Central to the analysis was the S-Pol radar operated by the NCAR/EOL Remote System Facility. Special appreciation is extended to the Colorado State University, North Carolina State University at Raleigh, University of Oklahoma, University of Wisconsin–Madison, University of Alabama in Huntsville, University of Manitoba, Millersville University of Pennsylvania, and NCAR/EOL for operating and maintaining the fixed and mobile sounding systems. Thanks to Dave Turner (NOAA/ESRL) for overseeing the planning of the fixed profiling sites for PECAN. These collective sounding systems resulted in an outstanding radiosonde dataset, without which the current research would not have been possible. The field data used in this study were obtained from the NCAR/EOL website (http://data.eol.ucar.edu) under the sponsorship of the National Science Foundation. The research was

Fig. 20. Conceptual vertical cross section of the primary atmospheric features, roughly 30 min prior to CI-H. The southwest–northeast cross section is centered on the CI-H cell. Color shading is used to show the layer of dry air (light brown) and the layer of high relative humidity and decreasing CIN (light green). The undulating black curves represent the trapped gravity waves moving to the northeast. The temperature inversion layer (topped by the dark blue line), sloping depth of the easterly winds (dark red), depth of the west and northwest winds (red), and freezing level (light blue) are labeled.
REFERENCES


partially supported by NCAR’s Short Term Explicit Prediction (STEP) program, which is sponsored by National Science Foundation funds from the U.S. Weather Research Program. The National Center for Atmospheric Research is sponsored by the National Science Foundation. The third author was supported by Grant AGS-1560945.


