Initiation of Deep Convection along Boundary Layer Convergence Lines in a Semitropical Environment

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ABSTRACT

The initiation of deep convection through forcing along boundary layer convergence lines is examined using observations from the Convection and Precipitation/Electrification Experiment conducted in east-central Florida during the summer of 1991. The study is concerned with the evolution and interaction of two converging air masses that were initially separated by an intervening boundary layer characterized by neutral stability and horizontal convective rolls. As anticipated, major thunderstorms erupt when the east coast breeze eventually collides with thunderstorm outflows from the west, but unexpected convection takes place prior to their merger along a well-defined confluence zone associated with a persistent quasi-stationary roll vortex signature. Analyses using wavelet transforms confirm that linear boundary layer reflectivity features are strongly correlated with radial convergence associated with roll vortices. In this study, complementary interactions between roll vortex convergence lines and the sea-breeze front are not sufficient to trigger deep convection. However, organized convergence along the eastward-spreading thunderstorm outflows did interact periodically with roll vortex convergence maxima to initiate a series of new storms.

Results from two-dimensional numerical model simulations replicate many of the observed boundary layer features. Surface heating produces circulations similar to sea-breeze frontal zones that appear near the coastlines and progress steadily toward each other as the interior boundary layer deepens. Vertical velocity maxima develop over the associated convergence zones, but weaker periodic maxima also occur within the interior air mass at intervals similar to the spacing of observed horizontal roll vortices. As the boundary layer deepens, a layer immediately above it cools, confirming organized large-scale ascent within the interior air mass. When surface heating is removed, circulation associated with this large-scale ascent collapses to a near-steady state where the width of the remaining prominent updraft is similar to its depth. This results from a balance between momentum advection by large-scale circulations and excess pressure developed at low levels near the center of the interior domain.

1. Introduction

Most of the work relating convection initiation to clear-air boundary layer convergence lines is based on observations over the semiarid High Plains of the United States (Wilson and Schreiber 1986; Wilson et al. 1988; Wilson et al. 1992). The Convection and Precipitation/Electrification Experiment (CaPE) conducted along Florida's east-central coast in the summer of 1991 (see Fig. 1) afforded the opportunity to study such correlations in a moist and generally more unstable semitropical environment. The ocean—land sea-breeze front is a recurring feature of the coastal circulation, and its interaction with thunderstorm outflows and with convergence zones of other origins was a primary focus of the CaPE research. High instability and daily development of sea-breeze fronts along both coasts accounts for a high frequency of thunderstorm days in the Cape Canaveral area (15–20 in the months of July and August; Watson 1990) and insured a plentiful number of research opportunities during the six weeks of the CaPE field operations.

Several boundary layer convergence zones of diverse origin were observed during the afternoon of 2 August 1991 within the CaPE network of observing facilities. Most of the phenomena observed have been identified in earlier studies as effective mechanisms for the initiation of deep convection (e.g., Wilson and Schreiber 1986; Wakimoto and Wilson 1989; Wilson et al. 1992). This case study is concerned with the evolution and interaction of three distinct air masses and the anticipated convective development that does and does not occur at the boundaries between them. Two converging air masses were initially separated by an intervening neutrally buoyant boundary layer characterized by horizontal convective rolls. The eastern member of the converging set was the westward-moving east coast sea-breeze front that had been modified

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somewhat by early afternoon convection that developed south of the CaPE network. The western boundary marked the leading edge of outflow from thunderstorms that had developed initially along the west coast sea-breeze front and spread steadily eastward across the peninsula during the afternoon. Outflow from earlier storms to the south also began to spread northward into the CaPE network during late afternoon.

Conventional wisdom would have predicted the onset of deep convection with the merger of the two major airmass boundaries. While this eventually did occur, convective development took place initially within the thermodynamically well-mixed intervening air mass several minutes prior to the merger of the principle airmass boundaries. This development occurred along a distinct confluent zone formed along the axis of ascent along a preferred quasi-stationary horizontal convective roll signature that persists as other roll structures disappear. Overall convective development patterns, confirmation of roll vortex structures and their interaction with airmass boundaries, and the origin of the intermediate convergence zone are of principle interest in this study.

A large and varied array of instrumentation was available in CaPE (Fig. 1). Data systems of primary importance to this study are the C-band Doppler radars operated by the National Center for Atmospheric Research (NCAR, CP-3 and CP-4) and Lincoln Laboratory (FL-2), the NCAR PAM II surface mesonetwork, mobile and stationary NCAR CLASS upper-air sound-
ings, and a special sounding system operated by Lowell University. Simultaneous kinematic and thermodynamic measurements in the boundary layer were provided by two similarly equipped twin-engine turboprop aircraft—the NCAR King Air (N312D) and the University of Wyoming King Air (N2UW). Cloud photography was also routinely available from a number of remote sites. A general description of observational capabilities, analysis techniques, and research strategies available from an integrated mix of these data sources is given in Wilson et al. (1992).

2. Large-scale synopsis

Several studies have shown that interaction of the large-scale synoptic pattern with the east and west coast sea-breeze fronts serves to modulate the timing, location, and intensity of convective development over the Florida peninsula (e.g., Byers and Rodebush 1948; Estoque 1962; Frank et al. 1967; Pielke 1974; López et al. 1984; Arritt 1993). Estoque (1962) found that an offshore ambient wind component intensified the sea-breeze perturbation by concentrating the horizontal temperature gradient in the frontal zone. Frank et al. (1967) concluded that the progression of convective activity from the windward coast across the peninsula in the direction of the basic synoptic flow and the development of a late afternoon maximum near or along the leeward coast are features of both easterly and westerly flow regimes. López et al. (1984) report, however, that the most widespread and vigorous convective development occurs under general southwesterly synoptic-scale flow. On such days the axis of the Atlantic or Bermuda high is positioned south of the area, and the southern extension of a midlatitude trough usually lies to the west, providing the possibility for large-scale ascent over the peninsula. Under these conditions there is typically a south to southwesterly low-level flow with moisture extending vertically through a deeper tropospheric layer (Blanchard and López 1985). Conditions of this type prevailed over central Florida on 2 August 1991. There was a trough of low pressure extending from Georgia southwestward into the central Gulf of Mexico. Winds aloft shifted to southerly in the midtroposphere and to easterly aloft, indicating a shift in the Atlantic ridge axis from south to north of the area with height. Midday surface winds were light south-southwesterly across the peninsula (Fig. 2a), matching conditions identified by Arritt (1993) as favorable for east coast sea-breeze frontogenesis.

Cumulus cloud streets indicative of horizontal boundary layer roll vortices were prominent in visual satellite imagery throughout the Florida peninsula by midday (Fig. 2a). Alignment was generally along the south-southwesterly synoptic-scale flow, and their minimum spacing was on the order of 4–6 km. At the same time, cloud development is discernible along sea-breeze fronts on both coasts. Earliest deep convection formed along the west coast sea breeze near Tampa–St. Petersburg near midday. Steering-level winds resulted in generally southwest to northeast cell movement. Low-level southwesterlies combined with downdraft outflows from the west coast activity to cause a fairly rapid eastward progression of downdraft-modified air. This outflow boundary triggered successive bursts of new convection and is seen approaching the Orlando vicinity by 1600 LDT (Fig. 2b). The area swept out by these storms is evident as the cloudless region over the west-central portion of the peninsula in Fig. 2b. As the east coast sea-breeze front moved slowly inland weak storms formed along it south of the Melbourne area. Farther south the interaction between the east coast sea-breeze front and a lake breeze from Lake Okeechobee triggered somewhat more vigorous activity. The anvils from these storms can be seen streaming westward across the peninsula in Figs. 2b and 2c. Low-level outflow from this activity reached the southern edge of the CaPE surface mesonet around 1800.

Comparison of Figs. 2a and 2b shows that cloudiness associated with roll structures decreased with daytime heating and that spacing between rows increased from midday to late afternoon. Both of these tendencies are consistent with the observations of Plank (1966). Cloud streets that were clearly visible on satellite images through early afternoon had nearly disappeared over the CaPE area by 1600 (Fig. 2b), apparently in response to a warmer and deepening boundary layer. Discussion in a subsequent section will show, however, that radar signatures of roll vortex structure persisted at this time in the south-southwesterly airflow west of the east coast sea-breeze front.

Three cloud lines are visible at 1800 in the western sector of the southern CaPE dual-Doppler lobe (Fig. 2c). Clouds on the western edge are developing along the western outflow boundary. The broken cloud line extending through the center of the lobe shows the position of the modified east coast sea-breeze front. The third line of clouds is forming over a boundary layer convergence line lying nearly equidistant between them. Between 1800 and 1900 a three-way collision of airmass boundaries, involving the east coast sea-breeze front, the outflow boundary approaching from the west, and the northward-moving outflow from storms to the south, occurred. Storm development patterns that took place along these boundaries will be discussed in more detail in the following section.

3. Storm development patterns

Figure 3 illustrates the evolution of events occurring over the west-central portion of the CaPE mesonet

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1. All subsequent reference to time will designate local daylight saving time (LDT), which is UTC minus 4 h.
Contoured gray shade represents radar reflectivity observed on low-elevation (0.3°) PPI scans from the CP-4 radar. Reflectivity with $Z_r \leq 20$ dBZ is primarily from clear-air return, and darker shades ($Z_r \geq 20$ dBZ) designate locations of rain shower and thunderstorm cells. At 1800 (Fig. 3a) two lines of enhanced “thin line” echo (separated by approximately 20 km and extending from north to south across the grid) mark the boundaries of air masses whose earlier history is tracked in Fig. 4. The band extending through the eastern sector of the grid marks the location of the westward-moving sea-breeze front, while the line of clear-air return to the west represents the leading edge of outflow from the recurring development of storms over the peninsula. In the northeast quadrant, alignment of the sea-breeze front parallels the east coast, but modification by downdrafts from earlier convection causes the eastern boundary to penetrate inland more rapidly south of the CP-4 radar. It was moving westward at a speed of 3.5 m s$^{-1}$ in this sector. The rain-cooled air mass approaching from the west was spreading eastward at a rate of approximately 4 m s$^{-1}$. While this air may have also originated as the west coast sea breeze, it too has been modified substantially by successive waves of new convection as seen in satellite images (Fig. 2).

The continuity of precipitating convective cells is identified from frame to frame in Fig. 3 by alphabetical lettering that also approximates their order of appearance. Since PPI (plan position indicator) scans are at a low elevation angle, reflectivity is from precipitation quite near the surface and generally represents a period late in the lifetime of evolving cells (i.e., the actual cell initiation might have occurred some 15−20 min earlier). Cells A, B, C, and D form over or near the western boundary and move north-northeastward at about 5 m s$^{-1}$. Cold-air production by downdrafts from these storms gives impetus to the east-southeastward movement of the western outflow boundary. The spacing between cell pairs A−B and C−D suggests that they may have been triggered by enhanced vertical velocity at locations where the outflow boundary intercepted the nearly stationary updraft axes of horizontal roll vortices. The implication is that periodic storm development is related to enhanced convergence along the parallel-oriented roll axes. Complementary interactions between preexisting convergence lines and boundary layer roll vortices have been found in numerical simulations (Crook et al. 1991) and reported in the observations of Wilson et al. (1992). We will examine this issue further in section 6.

A second family of cells (E, F, G, and H) forms along a confluence line (heavy dashed line in Fig. 3) embedded within the well-mixed air mass occupying the region between the converging western outflow boundary and the westward-moving sea-breeze front. Cells I and J form along outflow boundaries moving into the grid from the southwest and south, respec-
tively. Finally, as might be anticipated from earlier work (e.g., Wilson and Schreiber 1985), an outbreak of strong convection (cells L, M, and N) occurs after 1830 when the two major convergence boundaries eventually collide.

It is evident from the events illustrated in Fig. 3 that a variety of forcing mechanisms exist. Although (as will be shown in the next section) the thermodynamic properties of the air masses involved did not differ greatly, the development that occurred along the vari-
4. Kinematic and thermodynamic airmass properties

Surface conditions at 1815 (about 15 min prior to the collision of the western outflow boundary with the east coast sea-breeze front) are illustrated by mesonetwork data plotted in Fig. 5. Analog records of surface parameters from selected stations that expe-
The passage of the east coast sea-breeze front and the western outflow boundary (as illustrated in Fig. 4) are given in Fig. 6. Surface winds in the intermediate air mass are light south-southwesterly 2–5 m s$^{-1}$, similar to synoptic-scale surface airflow over the peninsula at midday. Except where stations are shaded by scattered cumulus, potential temperature $\theta$ within the intermediate air mass is generally within a degree of 305 K and mixing ratio ranges between 15 and 17 g kg$^{-1}$.

Air in the southeasterly flow behind (east of) the sea-breeze front is cooler ($\theta \sim 301–304$ K) and more moist with mixing ratio varying from 17 to 19 g kg$^{-1}$. The transition from the inland air mass to sea-breeze air is most evident in moisture and wind direction traces, with the largest temporal gradients occurring in 3–5 min. With a westward movement of 3–4 m s$^{-1}$, the sea-breeze frontal zone is between 500 m and 1 km wide at the surface. Moisture increases about 1 g kg$^{-1}$, and $\theta$ drops only about 1 K with the arrival of the western outflow boundary. The sharpest gradient there is in wind direction that shifts abruptly from south-southwest to west-northwest. Transitions in virtual potential temperature $\theta_v$ across both boundaries mirror variations in $\theta$.

Four soundings (Fig. 7) are used to illustrate the vertical structure and evolution of the lower troposphere within the inland intermediate air mass (Fig. 7a).
Fig. 4. Isochrone analysis of the position of the east coast sea-breeze front between 1600 and 1800 LDT and the western outflow boundary between 1700 and 1800 LDT. The southern CP-3–CP-4 dual-Doppler lobe is also indicated. Dots show locations of PAM II mesonet network sites, and the hatched line extending southwestward from the CP-3 radar shows the location of the prominent quasi-stationary horizontal roll vortex signature and associated intermediate confluence line. Grid corresponds to dashed rectangular boundary designated in Fig. 1.

and the sea breeze (Fig. 7b). Inland soundings were made near the UND (University of North Dakota) radar at 1553 (LWL) and near PAM mesonet station 23 (MCL) at 1831. Soundings within the sea breeze were released at Maddon (PAM station 12) at 1615 (TCO) and near PAM site 27 (DPK) at 1830 (see Fig. 1 for radar and PAM station locations).

A well-mixed dry-adiabatic layer is seen on the inland soundings (Fig. 7a), which warms from approximately 304 to 305 K and increases in depth from around 1 to 1.6 km in about 2.5 h. On both soundings, mixing ratio decreases sharply in the layer just off the surface to a nearly constant value of 14 g kg$^{-1}$ through the mixed layer—1 to 2 g kg$^{-1}$ lower than surface values. Winds veer from south-southwest to southwest with height and increase from 2 to 3 m s$^{-1}$ near the surface to about 8 m s$^{-1}$ at 2 km.

In the sea breeze (Fig. 7b), $\theta$ has a value of approximately 303 K through the mixed layer, which is 300 m deep at TCO and $\gg 1$ km at DPK. We note that both soundings were released shortly after the sea-breeze front had passed to the west of each site, but a deeper mixed layer at DPK may result partly from modifications caused by outflow from the earlier coastal thun-

Fig. 5. Gray-shade radar reflectivity as in Fig. 3c from CP-4 PPI scan at 1815:00 LDT. Potential temperature and mixing ratio are given at PAM II sites indicated by dots with wind vectors. Identification numbers for stations whose time series appear in Fig. 6 are included. Line segment with circles at 1-min intervals shows the flight track of the University of Wyoming King Air flown at 300 m AGL between 1812:00 and 1819:00 LDT. Aircraft wind vectors are plotted at 10-s intervals along the track and point in the direction of airflow. Vector scale for both aircraft and mesonet winds is 1 m s$^{-1}$ km$^{-1}$. Balloon symbols show locations of the MCL and DPK soundings plotted in Fig. 7.

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2 No soundings were made behind the western outflow boundary.

3 Errors in original aircraft track and winds have been corrected using techniques described by Rodi et al. (1991).
Flight-level winds shift from southeast to south through southwest and then to west-northwesterly as the aircraft passes through the different air masses. The sharpest windshifts occur at the airmass boundaries (identified by bands of enhanced clear-air reflectivity) and at a zone of confluence encountered near the middle of the intervening air mass. This is shown more graphically in an analog plot of wind direction illustrated with other kinematic and thermodynamic variables in Fig. 8a. In contrast to the rather broad frontal zone found at the surface, a near instantaneous shift in wind direction from 150° to 180° occurs when penetrating the sea-breeze front. At the confluence zone midway through the intermediate air mass, a sharp shift from 180° to 230° is encountered, and at the western outflow boundary, a shift from approximately 240° to 300° is observed. Convergence concentrated in the wind-shift zones produces three distinct peaks in vertical velocity with the one associated with the intermediate confluence zone being the narrowest and strongest.

Thermodynamic variables in Fig. 8a clearly distinguish the three different air masses, and the sharpest gradients are found to coincide with the wind shift zones. Air in the intermediate air mass is horizontally uniform and flight-level values agree well with those shown on soundings in Fig. 7a. Thermodynamic parameters display greater variability in the sea breeze, suggesting the presence of large-scale (1.5–2.5 km) eddies just east of the frontal boundary. Transitions in parameters observed at the western boundary are indicative of typical thunderstorm outflows, with sharply cooler and moister conditions observed immediately behind the outflow boundary.
Wind shears and thermodynamic gradients at air-mass boundaries were not as sharp at the 600-m flight level (Fig. 8b); however, vertical velocity peaks were somewhat stronger, particularly over the intermediate confluence zone. As at 300 m, thermodynamic conditions across the intervening air mass were remarkably uniform, and comparison with values at lower levels verifies the neutrally buoyant lapse rates shown on soundings in this sector (Fig. 7a). A wind shift to westerly at the western boundary is not observed at 600 m, and cooling is less than that at lower levels, indicating that the outflow air is fairly shallow. As at the lower flight level, large temperature and moisture fluctuations behind (east of) the sea-breeze front suggest overturning by large-scale eddies. Dual-Doppler radar analyses and numerical modeling results to be presented in later sections also show evidence of overturning circulations at scales indicated here.

The horizontal wind field at 1817 derived from dual-Doppler radial velocities observed by the CP-3 and CP-4 radars is given in Fig. 9 at an altitude of...
400 m. Airflow at this level corroborates observations at the surface and 300-m flight levels. Wind-shift zones are collocated with the major clear-air reflectivity bands at the leading edges of the converging peripheral air masses. Some spotty clear-air echo is also seen along the intermediate confluence line, and when horizontal airmass convergence is calculated from the winds in Fig. 9, the strongest convergence is found along this line. The confluence zone develops along the axis of the prominent and persistent horizontal roll signature identified in Figs. 4 and 5, and reflectivity exceeding 20–25 dBZ in Fig. 9 is from precipitating convection developing above it. Given the neutral stability and low-level convergence emphasized in this section, this early convective development is not surprising. The structure and evolution of horizontal roll vortices and detailed interactions between them and the major convergence boundaries are discussed in the next two sections.

5. Clear-air boundary layer reflectivity features

Gray-shade echoes in Fig. 10 illustrate clear-air radar reflectivity patterns observed at low elevation angle by the FL-2 radar at 1645 (Fig. 10a) and the CP-4 radar at 1730 (Fig. 10b). The area scanned lies within the dashed box outlined in Fig. 1. Although satellite imagery at 1600 (Fig. 2b) indicates minimal cloud cover over the southern CaPE dual-Doppler lobe, there is marked evidence of linear reflectivity structure in Fig. 10a, which implies that horizontal boundary layer roll vortices may persist. As with the cloud streets seen on midday satellite images (Fig. 2a), alignment of lineal

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4 Synthesis of dual-Doppler radar data is obtained using the CEDRIC software described by Mohr et al. (1986).
reflectivity maxima is essentially parallel to the mean surface airflow indicated by surface mesonetwork wind vectors. Noticeable changes occur in the linear echo structures in the 45-min period between Figs. 10a and 10b. Lines of reflectivity maxima become less distinct, and the average spacing between them increases from around 3 to 5 km at 1645 to more like 6 to 8 km at 1730. A boundary layer depth of 1.5–1.6 km produces an aspect ratio of 3–5, which agrees quite well with theoretical and empirical expectations (Kuettnet 1971; LeMone 1973). An increase in transverse wavelength with time is consistent with an observed deepening of the boundary layer during the afternoon (see Fig. 7a).

Quantitative verification of the quasi-steady nature of the radar signatures in Fig. 10, their lateral spacing, and linear structure is obtained by applying a slight variation of two-dimensional wavelet analysis to the reflectivity data. Wavelet analysis is a signal analysis method that has recently been shown to have useful applications in meteorology (Jones et al. 1993; Meyers et al. 1993; Gamage and Hagelberg 1993; Gamage and Blumen 1993), such as filtering, harmonic signal decomposition, and feature detection. Our interest here is in feature detection, and we correlate a cosine function of a prescribed wavelength $a$ with the reflectivity data. The correlation is performed over sectors of the data that are $a$ km $\times$ $a$ km in size, and a correlation value is obtained at each range-azimuth grid point. A complete description of the technique is given in the appendix.

Figure 11a shows contours of the correlation coefficient obtained when applying the analysis to reflectivity data measured by the FL-2 radar on a 1730 PPI scan at 0.5° elevation for $a = 6$ km. Highest correlations are found along the east coast sea-breeze front and at the western outflow boundary where, according to Wilson et al. (1994), one could expect a concentration of particulates to produce enhanced reflectivity. But there are also four or five distinct bands with high correlation values that have alignment and spacing comparable to reflectivity features illustrated in Fig. 10b. A significant amount of structure along the bands is implied by the periodic maxima in correlations found along their longitudinal axes. The quasi-steady nature in the lateral spacing and position of these bands is substantiated by comparison of the wavelet transform from prior and subsequent PPI scans, and an increase in wavelength indicated between Figs. 10a and 10b was also corroborated by the analysis. Arrows mark the prominent quasi-stationary roll signature highlighted in Fig. 4.

\footnote{Note that the grid is shifted northward in Fig. 11.}
along which the concentrated zone of convergence illustrated in Fig. 9 eventually develops.

Prior studies (Christian and Wakimoto 1989; Wilson et al. 1992) have shown that linear bandingness as seen in Figs. 10 and 11a likely results from concentration of radar reflective particulate by organized linear convergence between the axes of horizontal roll vortices. Unfortunately, scanning in the dual-Doppler mode required to produce horizontal vector fields equivalent to those presented, for example, by Kropfl and Kohn (1978) and Doviak and Berger (1980) was not available, so it was not possible to evaluate the horizontal convergence straightforwardly. However, if bands of enhanced reflectivity are correlated with convergence, one should expect to find convergence in the radial velocity from a single Doppler radar when scanning is transverse to the roll axes. Miller and Andersen (1993) have examined the convergence of radial velocity in the present case and found that coherent signals were not detectable above the background noise. However, when wavelet analysis of the type described above was applied to the convergence of the radial velocity signal for the same scan, the results illustrated in Fig. 11b were obtained. Although correlation coefficients are considerably lower (the outer contour is zero), due to the noise introduced by taking the radial derivative of the velocity field, a distinct linear structure appears that matches the reflectivity features found in Fig. 11a. This is evidence that the reflectivity bands are a consequence of convergence associated with organized convective circulations within horizontal roll vortices.

Further evidence of linear forcing along the convergent axes between horizontal rolls is presented in Fig. 12. This photograph was taken from a field site located at the CP-3 radar \((x, y = -19.4, 12.6)\) about 6 min after the radar scans depicted in Figs. 10b and 11. The looking angle is between 180° and 190°, or nearly due south from the CP-3 radar. A continuous cloud line, apparently associated with the prominent reflectivity band marked by arrows in Fig. 11a, is clearly visible on the right, and at least two shorter and more irregular cloud lines are shown in the near and distant field of view on the left. These are presumably associated with fragmented roll signatures lying between the major roll axis and the sea-breeze front as illustrated in Fig. 11a.

Cross-line and mean longitudinal properties of horizontal roll vortices have received exhaustive treatment in the literature (see, e.g., Brown 1980; Ettling and Brown 1993), but longitudinal irregularities of the kind presented, for example, in the analyses of Christian and Wakimoto (1989) and Kristovich (1993) and exhibited by roll structures in Figs. 10 and 11 have received little observational or theoretical emphasis. Theory and observations show that Rayleigh-type convection gives way to two-dimensional longitudinal instabilities in the presence of shear, that is, convection forces waves with axes parallel to the plane of the velocity shear. In modeling the unstable boundary layer, Deardorf (1972) found that as thermal instability increased the tendency for convection to be elongated with the shear diminished. From instability parameter analyses, Grossman (1982) concluded that the transition from two-dimen-
sional roll structures to more three-dimensional random convection occurred over a small range of stabilities and shear stresses. Strong shear and weak thermal forcing favored roll formation, while weak shear and strong heating lead to more random convective structures. The latter conditions would appear to apply in the present case as two-dimensional cloud streets prominent in midday satellite images (Fig. 2a) evolve to fragmented roll vortex signatures during the afternoon. To explain this trend we must appeal to decreasing vertical shear, since surface heating will also decrease after midday. Decreasing wind shear is a likely consequence of thermally driven convection acting through a major portion of the daylight hours.

6. Interaction between horizontal rolls and airmass boundaries

a. Structure near and along the sea-breeze front

Figure 13 shows the track of King Air N2UW flying at 900 m above ground level (AGL); first northward along the axis of the sea-breeze front, then westward and southward into the intermediate air mass, and eventually eastward again across the sea-breeze front. The NCAR King Air flew the same track simultaneously at an altitude of 1.6 km. Gray shade represents clear-air echo, and as in earlier figures, the band of highest reflectivity shows the location of the sea-breeze front. There is some evidence of banded structure in the clear-air echo over the northwest sector of the grid, and as shown in Figs. 10b and 11, separation between the better-defined bands in this sector is around 6–8 km. Linear structure is less pronounced near the sea-breeze front; however, patchy echo does appear to be aligned preferentially along the ambient south-southwesterly airflow.

Selected parameters measured at a frequency of 1 Hz along the track in Fig. 13 are plotted in analog form in Fig. 14. Periodic updraft maxima are marked by vertical lines and are numbered sequentially forward and then backward from 1 to 6. The position of each maximum is also plotted as a large dot on the track in Fig. 13. The intent of the numbering system is to imply a relationship between numbered pairs, which when connected linearly would reflect orientation of rolls as seen in the northwest quadrant here and in Figs. 10 and 11. Although there is no one-to-one correspondence between updraft maxima and lineal echo structures as was found, for example, by Reinking et al. (1981), several maxima in the well-mixed intermediate air mass do tend to fall over or near bands of clear-air echo. Others (e.g., the northern maximum labeled 5) clearly do not. Over the sea-breeze front, strong updraft maxima labeled 2 and 3 correlate well with strong reflectivity centers that presumably reflect zones of maximum low-level convergence along the front.

Except over the sea-breeze front, where there is strong organized forcing from below, there is little evidence of high correlation between updrafts and the temperature and moisture parameters, as was reported, for example, by Atlas et al. (1986) for longitudinal rolls observed in cold air spreading over a warmer ocean surface. A lack of correlation in the thermodynamically well-mixed air to the west of the front may
by the NCAR King Air were also weaker than those found over the sea-breeze front at lower altitude (Fig. 14). This damping with height is consistent with analysis presented in the top panel, where it can be seen that strong compensating divergence directly overflies the low-level convergence maximum. Sea-breeze air approaches the front from the southeast, ascends in the low-level convergent zone, but is then sheared sharply toward the northeast by increasing southwesterlies near the top of the boundary layer. Soundings in Fig. 7b show that sea-breeze air lifted from the surface must also overcome a moderate amount of negative buoyancy to reach the LFC. For air converging from the western side of the front, the detrimental effects of vertical wind shear, as embodied in the theory of Thorpe et al. (1982) and Rotunno et al. (1988), would appear to apply since winds increase with height on the western side while opposing winds decrease and, in fact, reverse sign on the sea-breeze side of the front.

b. Structure near and along western outflow boundary

Between 1700 and 1800, the position and movement of the western outflow boundary was monitored by low-elevation (0.5° and 1.0°) PPI scans from the FL-2 radar (see Fig. 4). During the same period, the history of convective-cell development was examined using 7.5-min PPI scans from the CP-4 radar at elevations of 0.3°, 3.0°, 5.0°, and 8.0°. Tracks of seven cells that developed near the western outflow boundary during this period are plotted in Fig. 16a. Only history of those echoes appearing above the minimum threshold of −15

result from evolving convective plumes that, following an afternoon of heating by surface fluxes from below, have begun to dominate tendencies for horizontal roll structures. In any event, the interaction between horizontal roll vortices and lines of strong convergence (represented here by the sea-breeze front), as proposed by Wakimoto and Wilson (1989) and Wilson et al. (1992), does not appear to be an adequate mechanism for initiating deep convection in the present case.

Further explanation for the lack of significant convective development along the sea-breeze front is presented in Fig. 15. The position and orientation of this vertical cross section derived from dual-Doppler analysis of CP-3 and CP-4 radial velocity measurements is given in Fig. 13, and analyses represent the spatial average across a swath 5 km wide and oriented normal to the front at the indicated location. In both panels gray shade identifies the location of the maximum low-level radar reflectivity. The upper panel shows that maximum low-level convergence (solid contours) is concentrated near or slightly east of the maximum reflectivity. Vertical velocity (solid contours, bottom panel) reaches a maximum of around 2 m s⁻¹ in the lowest one-third of the boundary layer (as defined by conditions west of the front) and decreases to near zero at the top. Vertical velocity maxima observed at 1.6 km
dBZ that survived for longer than 7.5 min were considered in the analysis. Cells first appeared in the altitude range of 3–5 km, and as shown schematically, they were usually located over or near a quasi-stationary linear reflectivity signature. The position of first echo appearance is also plotted relative to the eastward-moving surface outflow boundary in Fig. 16b, and we find that five of the seven cells first appeared within 0–3 km of the outflow boundary. Cells C and D formed near a roll vortex signature one wavelength east of the outflow boundary, but they did not intensify appreciably until that roll axis was later intercepted by the outflow. Cell pairs A–B and C–D in Fig. 3 are heavily precipitating mature versions of A2–B2 and C–D in Fig. 16.

East–west spacing between cell pairs and the location of successive development indicates that their parent clouds and updrafts were spawned ahead (east) of the outflow boundary over axes of linear convergence between horizontal roll vortices that were essentially parallel to the advancing outflow boundary (see Fig. 10b). The cells moved north-northeastward as the outflow boundary swept beneath them, and they typically reached their maximum intensity over or behind the outflow boundary, where their downdrafts gave added impetus to the eastward-spreading outflow. A similar pattern of cellular development and intensification was reported by Kingsmill (1993).

7. Origins of the intermediate convergence zone

We have made several references to an organized zone of confluence situated approximately midway between the advancing flanking airmass boundaries (cf., Figs. 4, 5, and 11), and in Figs. 3 and 9 it was shown that convection develops along this feature prior to the merger of the major boundaries. Timely prediction of this unexpected development requires knowledge of the origins and evolution of the intermediate convergence line.

Single Doppler radar and surface wind measurements reveal that the convergence zone develops in situ and intensifies with time. Streamlines in Fig. 17 were derived from the motion of clear-air reflectivity targets using the TREC cross-correlation technique developed by Rinehart and Garvey (1978) and Tuttle and Foote (1990). As described in these works, the correlation analysis was done over two-dimensional reflectivity arrays 5 km on a side. Because of the relatively large array size and the filtering and interpolation process, only scales on the order of 5–10 km and larger are retained. Data are shown for the 0.5° elevation and are representative of the air flow 150–600 m above the ground. The streamline pattern at 1600 (Fig. 17a) reflects the generally uniform south-southwesterly airflow found across the peninsula near midday (cf., Fig. 2a). However, by 1710 (Fig. 17b) a distinct zone of confluence is evident extending from the southwest corner through the central portion of the grid. This trend toward confluence is confirmed by temporal changes in wind direction at mesonetwork sites lying on opposite sides of the confluence line. Wind direction at stations on the eastern side displayed a tendency to back with time (see, e.g., the wind direction trace at PAM 18 in Fig. 6 prior to the passage of the sea-breeze front), while wind direction at stations on the western side remained steady or veered with time.

In their early study of convective forcing over the Florida peninsula, Byers and Rodebush (1948) point out that development of an interior convergence zone is a natural consequence of landmass heating and the inland progression of coastal sea-breeze fronts. We now examine the development and evolution of these phenomena through use of a numerical model. Al-
though the formation of Florida sea breezes has been the subject of several previous studies with detailed three-dimensional models (see, e.g., Pielke 1974), the development of convective structures and horizontal divergence between the sea breezes has not received as much attention.

The model used is the anelastic, nonhydrostatic model described in Clark (1977) and Clark and Farley (1984). It is run in a two-dimensional mode with the 400-km domain oriented perpendicular to the Florida peninsula (aligned along azimuth 70°–250°). Horizontal resolution is 1 km, and vertical resolution is 200 m.
Sensible heating of the Florida peninsula is simulated by a surface heat source over the inner 200 km of the domain. Heating is held horizontally constant over the peninsula but allowed to vary as a function of the diurnal daytime incident solar radiation. The sensible heating is taken as 20% of the incident solar radiation so as to match flux measurements taken at a number of special inland surface sites on this day. Surface drag is imposed over the peninsula with a constant value of $5 \times 10^{-3}$ s$^{-1}$.

The initial conditions are horizontally uniform and are taken from a CLASS sounding launched at 0700 from a site about 10 km south of PAM station 37 (see Fig. 1). This is also the start time for the simulation. The velocity component parallel to the domain is $+3$ m s$^{-1}$ (i.e., toward 70°) in the lowest 4 km, goes to 0 at 7.0 km, and then reaches $-20$ m s$^{-1}$ at 10 km. Vertical velocity and buoyancy fields after 5 h (midday) are shown in Figs. 18a and 18b, respectively. Periodic structures in the vertical velocity field between the two flanking sea-breeze fronts extend through the depth of the boundary layer and have a wavelength at the time shown of approximately 8 km. Following trends in the observed radar reflectivity, this wavelength increases during the day from approximately 5 km at 0900 when the rolls first become apparent to about 15 km at 1800.

The updraft–downdraft structures shown in Fig. 18a are the representation of boundary layer convective eddies in a two-dimensional model. Similar structures have been found in other two-dimensional simulations (e.g., Clark et al. 1986; Sha et al. 1991). The assumption of two-dimensionality does require that we interpret the model results with some caution. However, there is no good reason to believe that many of the results will not carry over to three spatial dimensions, especially for flows that exhibit a high degree of two-dimensionality such as those that produce longitudinal cloud lines as shown in Fig. 2a. It should be noted that if we interpret the updraft–downdraft structures in Fig. 18a as representing longitudinal rolls, then, by definition, these rolls are aligned perpendicular to the model domain. Thus, a limitation of these results is that the roll orientation is determined entirely by the chosen domain orientation and not by the wind shear, as it would in a three-dimensional system. In a three-dimensional flow, the fastest growing modes are aligned parallel to the low-level shear vector (Asai 1970). In the two-dimensional simulations we have selected only those modes that are aligned perpendicular to the model domain. To examine the sensitivity to domain orientation, we performed a number of simulations with varying orientations. Convective structures developed in all of these simulations, with the spacing and strength of these structures varying by less than 20% between simulations.

Dashed contours in Fig. 18b show that positive buoyancy generated by surface heating in the boundary layer has produced a maximum buoyancy at this time of just over 3°C. The two sea breezes are delineated by horizontal buoyancy gradients and by the damping of convective activity behind the sea-breeze fronts. Solid contours in Fig. 18b also indicate that cooling has occurred above the boundary layer. A layer of cooling approximately 1 km deep extends across the whole region between the two sea breezes and reaches a maximum of approximately 1.5°C. This cooling is a result of large-scale ascent over the peninsula, which in turn is forced by heating of the land surface. The fact that the cooling occurs across the entire domain lying between the two sea-breeze fronts indicates that the ascent is also occurring over that scale.

Evidence for this cooling can be found in many sounding pairs taken during CaPE. For example, Fig. 7a shows that the layer between 1.3 and 3.5 km at MCL at 1831 is approximately 1°C cooler than the same layer at LWL 3 h earlier. Comparing soundings taken at the same location (LWL), the air above the boundary layer cools by about 2°C between 1000 and 1300 and by a further 1°C between 1300 and 1600. Sounding pairs taken on other days during CaPE consistently show this cooling above the boundary layer, with the largest change occurring between early morning and early afternoon soundings.

Results illustrated in Fig. 18 reveal three components to the circulation generated over a heated peninsula: periodic convective roll structures, the sea-breeze circulations, and large-scale convergence between the two
sea breezes. As the day progresses, the sea-breeze fronts move steadily inland, overtaking the convective roll circulations. Late in the afternoon, around 1730, the modeled sea-breeze fronts are approximately 30 km apart. At that time, there is one dominant updraft between the two sea-breeze fronts (Fig. 19). This intermediate updraft first appears approximately 30 min prior to the time shown in Fig. 19 and then strengthens throughout that period. This bears a close resemblance to the observed late-afternoon structure, as seen, for example, in Fig. 9. It should be noted, however, that the structure of the intermediate updraft was sensitive to some of the model parameters. A number of simulations were performed with varied conditions. Although an intermediate convergence zone occurred in all cases, grid resolution, drag coefficient, and heating rate all had some effect on the position, depth, and strength of the circulation.

We now consider possible explanations for the development of the intermediate updraft. A partial explanation is that the two wind surges continue to overtake the convective updrafts until only one remains between them. This phenomenon was indeed observed in conjunction with the western outflow boundary (section 6). However, the observations and model results also indicate that the intermediate convergence zone develops and strengthens in situ. A more complete explanation involves the manner in which surface heating affects the circulations. It has already been shown that surface heating forces large-scale convergence between the two sea breezes. It will now be shown that if the forcing is switched off, then the broad convergence zone collapses down to a width similar to the depth of the boundary layer.

We first note that there is a steady-state solution for a convergence line in an unstratified fluid between two horizontal surfaces. This solution is shown in Fig. 20a and is obtained from the equation for conservation of horizontal vorticity in a two-dimensional system (see, e.g., Crook 1991). The analytical form for this solution is given below the figure. The important point to note is that the characteristic width of the updraft $W$ is equal to its depth $H$. A convergence zone broader than this steady state can be derived by increasing $W$. The resultant flow with the width increased by an order of magnitude is shown in Fig. 20b. This flow is then used as initial conditions for a numerical integration.

Figure 20c shows the convergence zone that develops after 30 min of integration. Although some differences are apparent, it is clear that the broad convergence zone has collapsed down to a structure similar to the steady-state solution. To explain this collapse, we note that in the steady-state solution along the lower surface, a balance exists between the advection of momentum $u \partial \psi / \partial x$ and the pressure gradient $- \partial p / \partial x$, with a pressure maximum existing at the position of maximum convergence. This balance does not exist in the broad convergence zone; the advection of momentum exceeds the retarding pressure gradient. This creates unsteadiness in the flow so that higher momentum air is advected toward the updraft maximum on both sides. This advection continues until a pressure bulge develops to balance the momentum advection.

It is also interesting to examine the flow evolution in terms of vorticity production and advection. In a two-dimensional system, horizontal vorticity can be produced only by the baroclinic effect and redistributed by diffusion (the stretching and tilting terms vanish). This means that over a heated peninsula, large-scale vorticity is first generated at the coastlines due to the land–sea temperature contrast and then advected inland with the sea breezes. There is no mechanism for large-scale vorticity production within the interior air mass. However, turbulent motions will tend to diffuse the vorticity at the sea-breeze head into the interior air mass producing a broad zone of increasing vorticity. In comparison, the steady-state solution (Fig. 20a) has uniformly distributed vorticity on one side of the convergence line and then a discontinuous jump to equal and opposite vorticity on the other side. In the broadened convergence zone (Fig. 20b) this vorticity discontinuity has been smoothed out so that the vorticity transitions smoothly from one far-field value to the other over a distance of approximately 10 layer depths. This distribution is comparable to the observed stage.
when the two sea breezes have approached close enough that their diffused vorticity gradients are beginning to interact. At that stage, we hypothesize that in the absence of thermal forcing, the convergence will collapse down to a width similar to that in the steady-state solution.

A schematic representation of the late afternoon development is shown in Fig. 21. In the early afternoon, when the solar heating is near its maximum, there exists three components to the flow over the peninsula: 1) the circulations that initially develop at the two coastlines and move inland (in the present case, a modified east coast sea breeze and a west coast gust front, 2) the small-scale interior convective circulations, and 3) the large-scale ascent between the west coast and east coast wind surges (Fig. 21a). Late in the afternoon, as the solar heating decreases, the convective circulations begin to decay. However, the large-scale convergence decays more slowly, because of its greater inertia. Furthermore, since the circulation is no longer forced, it tends to collapse until its horizontal scale is similar to the depth of the boundary layer (as shown in Fig. 21b). This schematic picture describes many of the low-level features observed on this day and most of those obtained in the numerical simulations.

8. Summary and conclusions

We have examined a variety of boundary layer forcing mechanisms observed over the east-central Florida peninsula on a particularly active convective day during CaPE. Some of the convergence zones responsible for the initiation of deep convection have been identified in studies conducted over the High Plains, but the
events described here pertain to a much shallower boundary layer with considerably higher moisture content. The prevailing south-southwesterly synoptic airflow regime has been identified as favorable for widespread and vigorous convective development that did indeed occur over the interior and leeward coastal regions of the peninsula on this day.

Horizontal roll vortex signatures manifest as cloud streets and linear clear-air radar reflectivity bands were prominent midday and early afternoon features of the boundary layer. Through innovative radar signal processing, the enhanced reflectivity bands are shown to be convergence zones between axes of roll vortex pairs. One function of the horizontal roll circulations is to mix the boundary layer and deepen it with time. This eventually leads to a neutrally buoyant dry-adiabatic lapse rate with constant mixing ratio through the depth of the well-mixed layer. With sustained daytime heating and a deepening boundary layer, the distance between the roll axes increases with time and comparatively weak vertical wind shear combined with strong thermal forcing leads to less distinct and more fragmented roll signatures in the radar reflectivity patterns.

In earlier studies, interaction between convergence lines formed by roll vortex circulations and convergence zones concentrated at airmass boundaries has been identified as an effective mechanism for initiating deep convection. In the present study, the intersection of horizontal rolls with the westward-moving coastal sea-breeze front does produce local vertical velocity maxima in the middle of the boundary layer, but an unfavorable shear profile combines with divergence and stable air aloft to inhibit significant convective development over the organized low-level convergence along the front. Organized convergence also existed along the leading edge of outflow produced by thunderstorms spreading eastward across the interior of the peninsula. When it entered the CaPE array, this outflow boundary was essentially parallel to the axes of the roll vortices. As the outflow moved eastward, it periodically intercepted clouds formed over convergence lines between rolls, and the enhanced convergence occurring with each juxtaposition produced successive bursts of new convection. New cells triggered in this way moved northward, intensified, and eventually deposited downdraft that added impetus to the eastward-spreading outflow. As anticipated, major convective storms erupted with the collision of the two primary airmass boundaries; however, unexpected convective development took place several minutes earlier along an intermediate convergence zone that formed approximately midway between the converging airmass boundaries.

Results from two-dimensional model simulations replicate many of the boundary layer features observed during the day. The model, run in the dry mode, was oriented normal to the major axis of the Florida peninsula. Daytime diurnal surface heating was the only forcing applied. Circulations similar to sea-breeze frontal zones appear near the coastlines and progress steadily toward each other as the interior boundary layer deepens. Vertical velocity maxima develop over the associated convergence zones but weaker periodic maxima also occur within the interior air mass at intervals that approximate the spacing of observed horizontal roll vortices. Large-scale convergence develops over the interior, which produces cooling in a layer of 1–1.5-km thickness immediately above the boundary layer. This cooling amounts to 1 or 2 K and is also observed on soundings made within the interior air mass. With time, spacing between interior updrafts increases and eventually only a single dominant updraft remains.

When surface heating decreases, then large-scale ascent over the interior domain collapses to a near steady state where the width of the remaining prominent updraft is similar to its depth. This results from a balance between momentum advection by large-scale circulations and excess pressure developed at low levels near the center of the interior domain. The associated confluent zone was actually observed nearer the eastern coast, but this might be expected considering a prevailing south-southwesterly synoptic airflow and the fact that thunderstorm downdraft outflow caused a more rapid eastward progression of the western boundary.

We conclude with some speculative comments on the generality of the flow development on this day. The
external conditions (large-scale flow and surface heating) on this day were fairly typical for this time of year. Hence, if the flow development depicted in Fig. 21 is a natural, deterministic, response to these external conditions, then we should expect to observe this type of development on many days during the summertime. From previous studies and observations during CaPEx, it is clear that the development of sea breezes, convective rolls, and the large-scale convergence over the peninsula is a common occurrence. The collision of the wind surges from the two coasts is also probably a frequent event, although prior to this study it had not been well documented. However, whether or not the development of an interior convergence zone is a common occurrence needs to be verified through additional case studies.

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APPENDIX

Wavelet Transform Analysis

Wavelet analysis, considering first one-dimensional quantities, involves the convolution of some measured quantity, expressed here as a real function \( g(x) \), with a set of prescribed functions \( f_{ab}(x) \), known as wavelets, where

\[
f_{ab}(x) = \frac{1}{a^{1/2}} f \left( \frac{x - b}{a} \right); \quad (A1)
\]

\( a \) and \( b \) are variables that represent a scaling and translation, respectively, of the basic wavelet \( f \). The convolution results in the wavelet transform \( T_{f(b,a)} \),

\[
T_{f(b,a)} = \frac{1}{a^{1/2}} \int f \left( \frac{x - b}{a} \right) g(x) \, dx. \quad (A2)
\]

The convolution is done over localized regions giving local amplitude and phase values of each harmonic, a distinct advantage over Fourier analysis where the average values for the entire dataset are yielded. The basic wavelet can have a variety of forms (cubic splines, single cosine wave, a hyperbolic tangent, etc.), the choice of which depends upon the nature of the features being extracted from the data.

The interest in this study is in feature detection, and we seek to maximize the function \( T_{f(b,a)} \) with respect to \( b \) and \( a \) (Jones et al. 1993), that is, we want to determine the locations and scales at which the wavelet best matches the measured quantity. In the present case we extend wavelet analysis to two dimensions and modify the transform equation slightly to obtain the correlation function. In the application to two-dimensional reflectivity data in range—azimuth space we have four degrees of freedom; translation in the \( r \) and \( \theta \) directions, rotation, and scaling. We choose to eliminate the rotation component by prescribing an axial symmetric basic wavelet of the form \( \cos (2\pi r/a) \), where \( r \) is the distance from a range—azimuth data point to the center of the local analysis area and \( a \) is the variable wavelength. While this form of wavelet decreases the response of \( T_{f(b,a)} \) to linear features, it results in a significant decrease in computation time, and as shown in Fig. 11, linear features are still readily detected.

Starting at some range—azimuth grid point (the center point of the local analysis area), the processing proceeds by convolving the cosine wavelet function with the data points that lie within half a wavelength \((a/2)\) of the center point, that is, the convolution is performed over one full cosine wave. We then normalize by the product of the standard deviations of the reflectivity and the wavelet function to obtain a correlation value, that is, our wavelet transform has the form

\[
\sum \sum Z_{rj} \frac{\cos (2\pi r_j/a)}{\sigma_r \sigma_f}, \quad (A3)
\]

where the summation indices \( i \) and \( j \) are over range and azimuth, respectively, \( Z_{rj} \) is the radar reflectivity at the \( i \)th and \( j \)th grid point, \( r_j \) is the distance from the center point to the \( i \)th and \( j \)th data point, and \( \sigma_r \) and \( \sigma_f \) are the standard deviations of the reflectively and wavelet function, respectively. Again the summation is done only for \( r_j \leq a/2 \). Continuing to subsequent data points, the process is repeated until the entire dataset has been convolved obtaining a correlation value at each grid point. One then proceeds through the dataset again with a different wavelength, obtaining a correlation field for each wavelength. We applied the analysis over a number of wavelengths in the range of 3–12 km but present only the results for \( a = 6 \) and 8 km—wavelengths that maximize the response to the lineal reflectivity features seen in Figs. 10a and 10b, respectively.

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