Observations of the Sea-Breeze Front during CaPE. 
Part II: Dual-Doppler and Aircraft Analysis

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ABSTRACT

The three-dimensional kinematic structures of offshore and onshore flow sea-breeze fronts observed during the CaPE experiment are shown using high resolution dual-Doppler and aircraft data. The fronts interact with horizontal convective rolls (HCRs) that develop within the convective boundary layer. Nearly perpendicular intersections between the HCRs and sea-breeze front were observed during the offshore flow case. Close to the front, the HCR axes were tilted upward and lifted by the frontal updrafts. Consequently, a deeper updraft was created at the intersection points, providing additional impetus for cloud development. Furthermore, clouds forming at periodic intervals along the HCRs intensified as they propagated over the front.

During the onshore flow case, the HCR orientation was nearly parallel to the front. Extended sections of the front "merged" with the HCRs. This process strengthened the front and is explained as the merger of like-sign vortices associated with both the front and HCRs. Clouds formed along the intensified portions of the front and at the locations of periodic enhancements on the HCR, which were present prior to the merger.

Documentation of two distinct frontal boundaries is presented for the onshore flow case. The first is a kinematic sea-breeze front delineating the region of maximum near-surface convergence between the sea-breeze air and the warmer, drier environmental air. The second is a thermodynamic sea-breeze front, which delineates the location where the mean thermodynamic properties differ from the ambient air mass. It is generated by the interaction of the HCRs with the sea breeze and extends a few kilometers ahead of the kinematic frontal position.

The kinematic differences between the two cases are quantitatively illustrated. The offshore flow case exhibited stronger low-level convergence, larger vertical velocities, and larger radar reflectivity values. The source air for the clouds developing along the front originated from the ambient and moist sea-breeze air masses for the offshore and onshore flow cases, respectively.

1. Introduction

The sea-breeze circulation is one of the most extensively studied mesoscale meteorological phenomenon, with a history dating back to the seventeenth century (Jehn 1973). A myriad of theoretical and observational studies have emerged since World War II clarifying the underlying physics and general kinematic and thermodynamic structure of the circulation [see Atkinson (1981) for a review]. Its effects on coastal meteorology such as air pollution transport (e.g., Mizuma and Kakuta 1974; Bornstein and Thompson 1981; Kitada et al. 1984; Ueda et al. 1988), location and initiation of convection (e.g., Byers and Rodebush 1948; Gentry and Moore 1954; Pielke 1974; Pielke and Cotton 1977; Blanchard and López 1985; Nicholls et al. 1991), aviation safety (Watts 1955), and forest fire forecasting (Fosberg and Schroeder 1966) are recognized.

A sea-breeze front forms at the leading edge of the circulation and is structurally and dynamically similar to a density current, the horizontal flow of a fluid that is driven by the density difference between two fluids (Simpson 1987). A three-dimensional finescale structure along the front has been documented in laboratory experiments (e.g., Simpson 1969; Simpson 1972; Britter and Simpson 1978; Simpson and Britter 1979, 1980; Simpson 1982) and is created by two processes. The first is a shearing (Kelvin–Helmholtz) instability creating billows on the frontal interface that grow in size as they propagate rearward of the surface frontal position and then eventually break down into small-scale turbulence. The second process is a gravitational instability created when the denser fluid overruns the lighter one near the surface. A complex pattern of clefts and

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lobes is created in the direction parallel to the horizontal frontal orientation. Previous observational and numerical studies have not adequately resolved this horizontal alongfrontal variability due to either 1) coarse spatial resolution (e.g., Johnson and O’Brien 1973; Pielke 1974; Physick 1976; Zhong and Takle 1992), 2) the two-dimensional nature of the dataset (e.g., Hatcher and Sawyer 1947; Atlas 1960; Simpson et al. 1977; Nakane and Sasano 1986; Yoshikado 1990; Kraus et al. 1990; Reible et al. 1993), or 3) a combination of both (e.g., Sutcliffe 1937; Fisher 1960). Accordingly, a three-dimensional wind field associated with the sea breeze, with high spatial and temporal resolution explicating its horizontal alongfrontal structure, has not been documented in the literature.

The ambient flow direction (onshore vs offshore) has been shown to affect the sea-breeze front (e.g., Wexler 1946; Estoque 1962; Frizzola and Fisher 1963; Pielke 1974; Arritt 1993; Reible et al. 1993). These studies have shown that offshore flow cases tend to have a more pronounced discontinuity, are more intense, and do not penetrate as far inland. Except for laboratory experiments (Simpson and Britter 1980), detailed, quantitative observations highlighting the differences in frontal structure between onshore and offshore flow cases have not been previously documented.

The sea-breeze circulation is an atmospheric response to the differential heating across the land–sea interface (i.e., the conversion of available potential energy to kinetic energy). It interacts with the convectively active boundary layer as it propagates inland. The coexistence of sea-breeze fronts and boundary layer convection has been previously shown (e.g., Drake 1982; Mitsumoto et al. 1983; Nakane and Sasano 1986). These studies indirectly suggest that the boundary layer convection may affect the frontal structure. Moreover, studies by Crook et al. (1991) and Wilson et al. (1992) over the High Plains have shown that boundary layer convection can significantly alter the kinematic structure of convergence zones.

The factors that control the initiation of clouds along convergence boundaries are an important issue that has received relatively little attention in the literature. Wilson et al. (1992) noted that the intersection points of horizontal convective rolls (hereafter, HCRs) with the boundary were preferred locations for clouds to initiate. HCRs are a common form of boundary layer convection (e.g., Kuettni 1959; Asai 1970a,b; LeMone 1973; Brown 1980) consisting of counterrotating helices aligned nearly parallel to the mean boundary layer wind direction. Knowledge of the factors governing where and when clouds will initiate is important for making accurate nowcasts (<2 h) for thunderstorm development (Wilson and Mueller 1993). Many investigators have noted a general relationship between the convergence zone generated by the sea breeze and thunderstorm activity, however, as shown by Leopold (1949), enhanced cloud development can occur at periodic locations along the front, similar to the results discussed by Wilson et al. (1992). The mechanism creating this periodic development of clouds along the sea-breeze front is not clearly understood.

During the Convection and Precipitation/Electrification (CaPE) Experiment (Gray 1991), the NCAR (National Center for Atmospheric Research) CP-3 and CP-4 Doppler radars (Keeler et al. 1991) performed coordinated high-resolution sector scans on sea-breeze fronts. This clear-air data was able to resolve the horizontal vorticity structure of the sea-breeze front and HCRs, allowing for the synthesis using dual-Doppler techniques of the three-dimensional wind field with unprecedented spatial and temporal resolution. Brandes and Rabin (1991) have shown that synthesis of the three-dimensional wind field is possible from clear-air radar data. Previous studies using radars (e.g., Atlas 1960; Eastwood and Rider 1961; Simpson 1967; Pedgley et al. 1982; Drake 1982) and lidars (e.g., Nakane and Sasano 1986; Banta et al. 1993) have presented high-resolution reflectivity data and/or cross sections (either horizontal and/or vertical) of two-dimensional wind fields along sea-breeze fronts. No three-dimensional wind fields, however, have ever been synthesized.

In Part I of this study [Wakimoto and Atkins 1994 (hereafter, WA94)], satellite, cloud photography, and radar reflectivity data were used to show that HCRs modulated the horizontal alongfrontal structure and appeared to control the organization of clouds along the front. The study herein uses dual-Doppler and aircraft data to address the following objectives:

- To understand the complex interaction between the HCRs and the sea-breeze circulation using the synthesized three-dimensional wind field from the radar data and aircraft observations. In particular, how the HCRs affect the structural and dynamical characteristics of the sea-breeze fronts and its relation to cloud development will be explored.
- To quantitatively describe the differences in frontal characteristics between the offshore and onshore flow cases.

In this study, the following definitions will be adopted for the sea-breeze front and HCRs:

**Kinematic sea-breeze front**—The location of maximum near-surface (0.1 km AGL) convergence, as determined from the dual-Doppler analysis, separating the warmer and drier air within the convective boundary layer (CBL) from the moist, sea-breeze air.

**HCRs**—The axes of maximum convergence, which occur between counterrotating circulation(s) (roll pairs) within the CBL.

Section 2 will describe the data and methodology employed for this study. Sections 3 and 4 discuss the offshore and onshore flow cases, respectively. Finally, section 5 presents a concluding discussion and comments on future work.
2. Data sources and methodology

The CaPE experiment (Gray 1991) was a multi-agency field program that operated intensively over east-central Florida from 8 July 1991 to 18 August 1991. Most of the ground-based observing platforms deployed for CaPE are shown in Fig. 1. They include the NCAR CP-2 (S and X band), CP-3 (C band), and CP-4 (C band) Doppler radars; NCAR CLASS (Cross-chain Loran Atmospheric Sounding System) fixed and mobile soundings (Lauritsen et al. 1987); the PAM II mesonet (Brock et al. 1986); Kennedy Space Center (KSC) tower data; and four photo sites taking 35-mm and time lapse pictures of the cloud field. In addition, airborne platforms included the NCAR and University of Wyoming King Airs, NCAR sail plane, and the South Dakota School of Mines and Technology armored T-28. High-resolution (1 km) satellite data were also recorded over the CaPE network.

a. Radar data—Scanning strategies and dual-Doppler synthesis

On 12 August (6 August) 1991, a sea-breeze experiment was conducted in the north lobe (south lobe). The CP-3 and CP-4 radars recorded high-resolution PPI sector scan volumes on the sea-breeze front. Typically, the time required to complete a volume scan was about 2.5 min on 12 August and 2.9 min on 6 August. These volumes were obtained approximately every 5 min for both days since the CP-4 radar was also required to scan periodically in a 360° surveillance mode for project control and nowcasting. The data were then synthesized using the dual-Doppler techniques described in the appendix and are displayed in the Cartesian coordinate system with Δx = Δy = Δz = 300 m.

b. Aircraft data

On 6 August, the Wyoming King Air made several penetrations through the sea-breeze front. The aircraft is equipped to measure the kinematic and thermodynamic state variables at a frequency of 1 Hz. The data were calibrated by making intercomparisons with instrumented towers (tower fly-bys) and aircraft intercomparisons (AIC) with the NCAR King Air. Biases in the dewpoint temperature and vertical velocity measurements of −0.97°C and −0.22 m s⁻¹, respectively, were noted and corrected (J. C. Fankhauser 1994, personal communication). In addition, uncertainties in the pressure, temperature, moisture, and wind measurements are approximately ±1 mb, ±0.5°C, ±1°C, and ±0.5 m s⁻¹, respectively (A. R. Rodi 1994, personal communication). The data were filtered such that energies having wavelengths less than approximately 2.4 km were severely damped. This filter was chosen to be consistent with the required filter employed with the radar data. The data were processed and displayed with the ACANAL (aircraft analysis) software developed at NCAR.

Since accurate estimates of the aircraft ground-relative position were deemed necessary for proper comparison and integration with the radar data, the aircraft position, as determined by the INS (Inertial Navigation System), was renavigated. Accordingly, errors created by the Schuler oscillation, which is a low-frequency (period approximately 84.4 min) response created by the erroneous inclusion of gravity into acceleration errors (Beck 1971), accelerometer biases, and gyroscope drift, were removed. The renavigation process employed for the CaPE dataset is similar to what is discussed by Rodi et al. (1991). Using radar echoes of the aircraft, the ground-relative position is estimated to be accurate to within 300 m.

3. 12 August 1991—Offshore flow event

The ambient flow on 12 August was southwesterly, a type 3 flow regime as defined by Blanchard and López (1985). The CBL depth over the CaPE network at the primary analysis time was approximately 1.3 km. The sea-breeze front forming on this day was characterized by an easily identifiable thin line (Wilson and Schreiber 1986) in the radar reflectivity field. A temperature drop and increase of the dewpoint temperature of about −1.5° and 4.5°C, respectively, were observed by PAM 18 (see Fig. 1 for site location) as the front propagated westward into the middle portion of the north lobe. No surface pressure change was associated with the frontal passage. The depth of the cold air was approximately 440 m. Alongfrontal variability was created by the nearly perpendicular intersections of the front with HCRs in the warm air. These intersection points were also preferred locations for cloud development.

a. Three-dimensional kinematic structure

1) Horizontal cross sections

At 0.1 and 1.3 km AGL (above ground level), the horizontal winds and convergence field at approximately 1950 UTC (hereafter, all times will be UTC, UTC = EDT + 4 h) are presented in Fig. 2 and correspond to the levels of maximum convergence and divergence within the dual-Doppler domain, respectively. Note that the front is located near the center of the radar-detected thin line. At 0.1 km AGL (Fig. 2a), the dual-Doppler winds show the southwesterly flow west of the front, and the south-southeasterly flow associated with the sea breeze. Note the good agreement with the PAM and KSC data in the ambient air. The convergence along the front and HCRs is apparent and quantitatively shown in Fig. 2b. Tremendous variability in the surface convergence field is observed along the front. Undulations generally occur at the intersection points with the HCRs. Although no clear relation-
ship between the intersection points and undulation location is apparent for HCRs 3–5 in Fig. 2, analysis of 15 dual-Doppler volumes from 1930 to 2047 UTC indicated that about 70% of the HCRs intersecting the front did so at the inflection point of the undulation, similar to HCR 4 in Fig. 2. Hence, it is hypothesized that the undulation spacing along the front may be approximately the same as the HCR separation, approximately 3.6 km at this analysis time, and is consistent with the value of 3–4 km reported by Drake (1982).

Maxima in convergence along the front and HCRs are generally collocated with maxima in radar reflectivity. An average value of $1.6 \times 10^{-3}$ s$^{-1}$, as reported in Table 1, is observed along the front. At 1.3 km AGL (Fig. 2c), the wind field is greatly perturbed along the HCRs and the front, and tremendous vari-
ability in the convergence field (Fig. 2d) is observed.

2) Relationship between clouds, reflectivity, and vertical velocity

Further evidence of the alongfrontal variability is illustrated in Fig. 3, which shows the periodic nature of vertical velocity $W$ along the sea-breeze front where maxima ($\approx 2$ m s$^{-1}$) are generally located at the intersection points with the HCRs. Average reflectivity and $W$ values along the front (see Table 1) are 9.6 dBZ$_e$ and 1.2 m s$^{-1}$, respectively. The $W$ magnitudes shown in Fig. 3a are consistent with values documented with aircraft penetrations by Simpson et al. (1977) (0.7–3.0 m s$^{-1}$) and Kraus et al. (1990) ($\approx 2.0$ m s$^{-1}$). Along the HCRs, periodically spaced $W$ maxima are responsible for initiating clouds like “pearls on a string” (e.g., Kuettner 1959; Christian and Wakimoto 1989). The mechanism responsible for generating the alongroll periodicities is not well understood and is currently under investigation.

Also apparent in Fig. 3a is the correlation between upward $W$ and reflectivity enhancements. Wilson et al. (1994) have shown that the clear-air scatterers creating the reflectivity enhancements within the well-mixed boundary layer are primarily birds and insects. The density of insects during the day decreases with height and is correlated with the temperature lapse rate (Johnson 1957a,b; Isard et al. 1990). Hence, reflectivity
12 August 1991

W (ms\(^{-1}\)) Reflectivity (dBZ)

A

1950:45 - 1953:13 UTC

Clouds

W (ms\(^{-1}\))

B

2 km

Fig. 3. Horizontal cross sections showing the relationship between the clouds, reflectivity, and W fields. (a) Reflectivity field (gray) and W (contoured every 1 m s\(^{-1}\)) is plotted at 1.0 km AGL. The 0 m s\(^{-1}\) contour is not shown. (b) Vertical velocity (gray) and cloud locations with number identifiers are shown. The dashed contour is a radar echo observed above the prevailing cloud base, however, the cloud was not visually observed due to intervening clouds. The dashed contour with hatching represents a radar echo observed above the prevailing cloud base, but no cloud was associated with it.

Maxima will occur at the same location as W maxima at higher levels since more (and larger) insects would be transported upward in the stronger updraft regions.

The close association between the updraft maxima and clouds\(^1\) is apparent in Fig. 3b and supports the validity of the dual-Doppler-derived vertical velocities (discussed further in the appendix). The tendency for the displacement of the clouds to the northeast of the W maxima can be explained as follows. A parcel within an updraft at 1.0 km AGL will travel about 0.7 km vertically before reaching cloud base. Given an average horizontal wind speed between 1.0 and 1.7 km AGL of 4.7 m s\(^{-1}\) toward 41° [from a sounding launched at TICO (see Fig. 1 for site location) at 1803 UTC] and an updraft magnitude of 2.0 m s\(^{-1}\), the parcel will travel 1.6 km horizontally before reaching cloud base. This result is consistent with the displacement shown in Fig. 3b.

An interesting observation to be made regarding HCR characteristics closer to the sea-breeze front during the offshore flow case is that their mean reflectivity, convergence, and vertical velocity values increase and the HCRs also change their orientation so that the intersections with the front are nearly perpendicular. This effect is obvious in Fig. 9 from WA94 and is apparent with HCRs 3 and 5 in Fig. 2. The HCR modification occurs within 5–6 km of the front and was a common observation made on many days during CaPE. Wilson et al. (1992) showed a similar behavior for HCRs encountering the Denver convergence line (see their Fig. 23), however, this observation was not discussed. The next section attempts to elucidate the mechanism potentially responsible for this phenomenon.

3) HCR MODIFICATION BY THE SEA-BREEZE FRONT

A great deal of literature has been published regarding the characteristics of HCRs. Relevant for this case is the theoretical work by Kuo (1963) and Asai (1970a,b). In particular, Asai (1970a) has shown that, for a plane Couette flow with unstable stratification, longitudinal perturbations (perturbations where the wavenumber in the direction of the mean flow is much smaller than in the perpendicular direction, i.e., \(k_x \gg k_z\), if the mean wind direction is along the \(x\) axis) resulting from thermal instability modified by a shear flow are able to grow from the conversion of both potential energy and kinetic energy of the mean flow. The conversion of the mean flow kinetic energy to the perturbation, which is proportional to the magnitude of the boundary layer shear, occurs through the downward transport of horizontal momentum. The longitudinal mode is oriented parallel to the shear vector, a result that has been verified observationally by Ferrare et al. (1991).

\(^1\) The cloud locations and numbers in Fig. 3 and all subsequent figures correspond to those determined by WA94 using photogrammetric techniques (see their Figs. 9 and 18 for the offshore and onshore cases, respectively). Cloud depths in subsequent figures are determined in a similar manner.
### Table 1. Frontal statistics for select fields.

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<tbody>
<tr>
<td>$\zeta$ ($\times 10^{-3}$ s$^{-1}$)</td>
<td>$\zeta$ ($\times 10^{-3}$ s$^{-1}$)</td>
</tr>
<tr>
<td>Min</td>
<td>-1.1</td>
</tr>
<tr>
<td>Max</td>
<td>5.0</td>
</tr>
<tr>
<td>Mean</td>
<td>1.2</td>
</tr>
<tr>
<td>$\sigma^d$</td>
<td>2.1</td>
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* $^a$ Data measured along the front at a height of 0.1 km AGL.
* $^b$ Data measured along the front at a height of 1.0 km AGL.
* $^c$ Data measured along the front at a height of 0.7 km AGL.

Figure 4a shows the computed boundary layer shear vector$^2$ ($0.5 \times 10^{-3}$ s$^{-1}$, toward 34°) for the environmental flow and is nearly parallel to the HCR orientation, consistent with Ferrare et al. (1991). Closer to the front, however, the shear vector intensifies and rotates about 21° clockwise ($1.6 \times 10^{-3}$ s$^{-1}$, toward 55°). This angle of rotation is roughly consistent with the value of 34° calculated for all HCRs observed to intersect the front in Fig. 9 from WA94. Consequently, the HCR orientation seemingly changes with the shear vector within the zone 5–6 km ahead of the front, and their intensity increases since they are able to convert more kinetic energy from the mean flow.

The modification of the shear vector magnitude and direction within 5–6 km of the front is primarily due to the decrease of the front-normal surface winds by $\approx 1.5$ m s$^{-1}$. In Fig. 4b, the surface stations PAM 18 and K1000 clearly show the $U$ component decreasing within the zone 5–6 km ahead of the front, thus, changing the shear vector. The slowing of the front-normal ambient flow by thunderstorm outflows has been previously documented (e.g., Goldman and Sloss 1969; Charba 1974; Wakimoto 1982) and attributed to a nonhydrostatic pressure gradient generated by the collision of the warm and cold air masses (Wakimoto 1982; Droegemeier and Wilhelmson 1986). The magnitude is governed by $P_{sh} = \frac{1}{2} \rho V^2$, where $\rho$ and $V$ are the density and velocity of the cold air. Using the values in Table 1 from WA94, a nonhydrostatic pressure rise of 0.07 mb may be generated by the front and is consistent with a retrieved value (Gal-Chen 1978) of 0.1 mb from the dual-Doppler data (not shown). This small nonhydrostatic pressure rise would be difficult to detect at a surface station, consistent with PAM data shown by WA94 (their Fig. 6). In addition, assuming the flow is irrotational, frictionless, and homogeneous, the equation of motion normal to the frontal orientation [see Eq. (15) from Wakimoto (1982)] yields a decrease of approximately 1.9 m s$^{-1}$ of the front-normal ambient flow speed for a $P_{sh} = 0.07$ mb.

#### b. Cloud development along the sea-breeze front

1) HCR TILTING BY THE SEA-BREEZE FRONT

The vertical cross sections (see Fig. 3a for their locations) in Fig. 5 show that the HCR circulations are strengthening and being tilted by the frontal updrafts, having a slope of approximately 20° from the horizontal. Note that in $(DD')$, the counterrotating circulations associated with HCRs 3–5 are evident and centered 0.5–0.7 km AGL. Closer to the front $(EE')$, however, the negative horizontal vorticity $\omega$ maxima associated with HCRs 3 and 4 and the positive $\omega$ maximum with HCR 3 have shifted upward to 1.0–1.2 km AGL and have intensified (with the exception of the HCR 4 negative circulation). Behind the front $(FF')$, only the positive circulations associated with HCRs 3 and 4 exist and are centered vertically at 1.0–1.3 km AGL atop the frontal boundary.

The tilting of the HCR circulations is first observed to occur 1–1.5 km ahead of the front. Only the positive circulations appear to extend rearward of the front after they have been tilted and lifted by the frontal updrafts. In doing so, they mix the ambient and sea-breeze air masses on the sea-breeze front head, similar to the effect created by Kelvin–Helmholtz (K–H) billows on the head of laboratory-produced (Britter and Simpson 1978) and atmospheric (Droegemeier and Wilhelmson 1986; Mueller and Carbone 1987; Sha et al. 1991; Weckwerth and Wakimoto 1992) density currents. Furthermore, the positive roll circulations create reflectivity and $W$ maximum at upper levels along the roll axis. This effect is also shown in Fig. 3a along HCR 4. Maxima in the reflectivity and $W$ fields extend approximately 2 km behind the frontal position, however, due to the lack of scatterers in this region, it is not known if they extend farther. Reflectivity minima are also cre-
ated between the maxima along the front due to the descending portion of the roll circulations bringing down relatively insect-free air from the upper portions of the CBL. Indeed, an alternating pattern of reflectivity minima and maxima can be seen along the front. It is not clear if the negative circulations extend behind the front, as this observation is not conclusively resolved in the dual-Doppler analysis. If so, they have been weakened since the horizontal low-level flow direction of the negative circulations is approximately north-northwesterly, opposite to the south-southeasterly flow associated with the sea breeze.

Further evidence that the HCRs are being tilted by the frontal updrafts can be seen in Fig. 6 (the data location in Figs. 6a,b is shown in the box in Fig. 2a). Note that the plus-minus horizontal vorticity $\omega$ maximum associated with the HCR 4 circulations terminate just ahead of the front and at the same location as the plus-minus vertical vorticity $\zeta$ maximum at 1.3 km AGL, consistent with the hypothesis that the horizontal roll vorticity is tilted. Given an angle of 20° and $\omega$ magnitudes of $-6 \times 10^{-3}$ and $10 \times 10^{-3}$ s$^{-1}$ for the negative and positive circulations, respectively (see Fig. 5), this will generate $\zeta$ maxima at upper levels with magnitudes of $-2.1 \times 10^{-3}$ and $3.4 \times 10^{-3}$ s$^{-1}$, respectively. This is in good agreement with the observed $\zeta$ values shown in Fig. 6a. The vertical stretching term does not appear to be important for creating the $\zeta$ maxima at upper levels since they are not collocated with updrafts.

A different $\zeta$ structure exists at lower levels and is shown in Fig. 6b. A $\zeta$ maximum is located at the intersection point along the front and extends from the surface to 0.7–1.0 km AGL (not shown) and appears to be generated in the following manner. Behind the front, near the location of the intersection with an HCR, the low-level flow is influenced primarily by two factors. The first is the south-southeasterly flow associated with the sea breeze. The second is due to the south-southeasterly flow associated with the positive roll circulation. By adding these two wind velocity components together, an enhancement and counterclockwise turning of the low-level flow is locally generated and is evident in the horizontal winds (Figs. 6b and 2a). The rearward extent of the positive roll circulation approximately coincides with the region of perturbed winds behind the front. This enhancement and turning of the low-level winds locally increases the shear across the frontal interface. The vorticity structure along the front is schematically summarized in Fig. 6d.

2) CLOUD DEVELOPMENT MECHANISMS

There are two mechanisms responsible for cloud development at the intersection points along the sea-breeze front. The first is due to the lifting of the HCR axes by 0.5–0.6 km, as they are tilted by the frontal updrafts. This has the effect of creating a deeper updraft along the HCR axes over the front and, hence, a greater potential for initiating clouds. Moreover, the water vapor mixing ratio is relatively larger in updraft regions due to the upward transport of moist surface air (Brümmer 1985). Note that the observations suggest that the positive roll circulation has a much greater effect of creating the deeper updraft since it is much stronger than the negative circulation as it propagates over the front.

The second mechanism involves the intensification of an existing cloud located over a periodic enhancement along an HCR. As the W maximum on the HCR and associated cloud move toward the front, the $W$ maximum is enhanced and becomes deeper as it collocates with the front, similar to the first initiation mechanism. The lifted enhancement will then intensify the cloud associated with it, creating a much deeper cloud.

These two mechanisms are supported by the observations shown in Fig. 7. Each panel shows a cross section (see Fig. 3b for the location) along HCR 4 for 10 consecutive dual-Doppler volumes. Cloud 37 supports the first mechanism since it initiated along the front.

![HCR Turning into the SBF](image_url)

**Fig. 4.** (a) Calculated shear vectors and the reflectivity field (gray) are shown. The dash-dot lines show the data positions for the space-time-adjusted (relative to the sea-breeze front) surface stations, PAM 18, and KSC tower K1000. (b) Plotted is the $U'$ component ($\text{m s}^{-1}$) of the winds from PAM 18, KSC tower K1000, and the north–south domain average dual-Doppler winds at 0.4 km AGL. The spatial scales are identical in both panels.
Cloud 13 supports the second mechanism since it intensifies as it propagates over the front. Furthermore, a detailed trajectory analysis (not shown) suggests that the source air for the clouds along the front originates in the warm air, consistent with the two initiation mechanisms. This result is corroborated by the fact that cloud base was at the same height for clouds along the front and HCRs. As the positive roll circulations mix the ambient and sea-breeze air, it is believed that some sea-breeze air is reaching cloud base, however, this fraction appears to be small.

The two initiation mechanisms illustrated in Fig. 7 differ from the scenario discussed by Wilson et al. (1992) for HCRs intersecting a preexisting conver-
Fig. 6. Vorticity structure along the sea-breeze front is shown. (a) $\zeta$ at 1.3 km AGL and $\omega$ (in gray with magnitudes greater than $\pm 4 \times 10^{-3}$ s$^{-1}$ shaded) at 0.7 km AGL are contoured every $2 \times 10^{-3}$ s$^{-1}$. The black dots denote the locations of the $\pm \zeta$ maximum. (b) $\zeta$ (in gray with magnitudes greater than $\pm 4 \times 10^{-3}$ s$^{-1}$ shaded) contoured every $2 \times 10^{-3}$ s$^{-1}$ and horizontal winds (m s$^{-1}$) at 0.4 km AGL. The heavy, black solid and dashed lines are the location of the $\omega = \pm 1 \times 10^{-3}$ s$^{-1}$ contours and approximately delineate the horizontal dimension of the HCR 4 circulations, shown schematically in (c). The thick, black circle denoting the location of the $\zeta$, is also shown in (a). (d) Schematic diagram illustrating the three-dimensional vorticity structure observed along the sea-breeze front.

They stated that clouds formed at the intersection points due to the enhanced low-level convergence created by the HCR intersecting the convergence line. However, as stated by the authors, the HCR locations were subjective, often difficult to identify and, hence, the relationship between the intersection points and initial cumulus field was tenuous.

It should be noted that significant, deep convection did not occur along the portion of the front and at the times discussed in Figs. 2–7. The cloud depths in Table 2 are shallow, with the exception of most clouds found along the front (e.g., numbers 14, 18, 19, and 23). If the environmental conditions had been more conducive for the development of deep convection, it would have happened at the intersection points, with even deeper convection occurring when an enhancement along an HCR was collocated, being lifted by the front.

The results discussed in this section are schematically illustrated in Fig. 8. The HCR $\omega$ vectors are turning clockwise with the shear vector as they approach the front and are tilted by the frontal updrafts. The cumulus field along the HCRs and front is also shown. Along the front, three clouds are located at the intersection points. Two clouds are experiencing enhanced growth due to the second initiation mechanism, while the shallower cloud is created by the first mechanism.

4. 6 August 1991—Onshore flow event

In contrast to the offshore flow case, the ambient flow on 6 August was southeasterly, a type 1 flow regime as defined by Blanchard and López (1985). The CBL depth at the primary analysis time was approximately 1.1 km. The sea-breeze front was characterized by a diffuse, at times difficult to detect, radar thin line. It was observed thermodynamically at the surface only by a gradual increase in moisture (dewpoint increase of about 2°C) as the temperature and pressure fields did not change noticeably during frontal passage. A wind shift of 50° counterclockwise was noted. Alongfrontal variability was observed, as it appeared that extended portions of the front had merged with HCRs oriented near parallel to the front. These sections were preferential regions for cloud development.

a. Three-dimensional kinematic structure

1) HORIZONTAL CROSS SECTIONS

An interesting feature of the satellite imagery shown by W/A94 is the banded structure of the cloud field at about 1710 UTC (their Fig. 11c). This structure is also evident in the radar reflectivity data shown in Fig. 9. Enhancement of the reflectivity field (0.3°) along three
extended sections of HCRs is observed. Fortunately, HCR 1 was located within prime dual-Doppler area in the south lobe so that the relationship between the front and enhanced bands could be determined and is shown best in the 3.0° scan. The enhanced reflectivity along a portion of the kinematic sea-breeze front and HCR 1 is apparent. Note that the frontal position is located along the enhanced portion of HCRs 1 and 2. It is believed that the front continues north and is also located along the enhanced portion of HCR 3. However, since the radars were not performing coordinated scans in the north lobe, the viewing angle from CP-4 was near parallel to the HCR 3 orientation, and since CP-3 was not operating in a surveillance mode, the kinematic frontal position cannot be accurately determined in this region.

In Fig. 10a, the sea breeze and ambient flows are evident in the horizontal wind field at 0.1 km AGL, the level of maximum convergence. The environmental flow to the west of the front is southeasterly, parallel to the HCR orientation while the sea-breeze air has a more easterly component. Note the good agreement with the surface PAM data. Similar to the offshore flow case, the kinematic frontal position is approximately centered on the radar-detected thin line. Alongfrontal variability in the convergence field is observed in Fig. 10b. The average near-surface convergence along the
Table 2. Cloud depths as determined from cloud photogrammetry.

<table>
<thead>
<tr>
<th>12 August/offshore flow (1950 UTC)</th>
<th>6 August/onshore flow (1715 UTC)</th>
</tr>
</thead>
<tbody>
<tr>
<td>cloud base = 1.7 km AGL</td>
<td>cloud base = 1.1 km AGL</td>
</tr>
<tr>
<td>Cloud no.</td>
<td>Depth (km)</td>
</tr>
<tr>
<td>-----------</td>
<td>------------</td>
</tr>
<tr>
<td>1</td>
<td>0.7</td>
</tr>
<tr>
<td>2</td>
<td>0.8</td>
</tr>
<tr>
<td>3</td>
<td>0.5</td>
</tr>
<tr>
<td>4</td>
<td>0.3</td>
</tr>
<tr>
<td>5</td>
<td>0.3</td>
</tr>
<tr>
<td>6</td>
<td>0.6</td>
</tr>
<tr>
<td>7</td>
<td>0.5</td>
</tr>
<tr>
<td>8</td>
<td>0.3</td>
</tr>
<tr>
<td>9</td>
<td>0.3</td>
</tr>
<tr>
<td>10</td>
<td>0.6</td>
</tr>
<tr>
<td>11</td>
<td>0.4</td>
</tr>
</tbody>
</table>

* Clouds on or just behind the sea-breeze front.

Front is $1.3 \times 10^{-3}$ s$^{-1}$ as reported in Table 1 and is slightly less than the value of $1.6 \times 10^{-3}$ s$^{-1}$ observed for the offshore flow case. At 1.0 km, the divergent flow along the kinematic front is clearly visible (Fig. 10c). Local minima (Fig. 10d) of approximately $-4 \times 10^{-3}$ s$^{-1}$ are located along the front, similar in strength to the HCRs. Curiously, within the zone between the kinematic and thermodynamic fronts, the HCRs, as defined by convergence and reflectivity, are ill-defined.

2) RELATIONSHIP BETWEEN CLOUDS, REFLECTIVITY, AND VERTICAL VELOCITY

The alongfrontal variability in the vertical velocity field is shown in Fig. 11a. Mean reflectivity and W values of 6 dBZ, and 0.9 m s$^{-1}$, respectively, are observed along the front and are smaller than those for the offshore case (see Table 1). The clouds in Fig. 11b (with the exception of cloud 0) are located along the portion of the front that has merged with HCR 1. Cloud depths valid at the analysis time are reported in Table 2.

3) VERTICAL STRUCTURE OF THE HCRs AND SEA-BREEZE FRONT

The vertical structure of the front and HCRs for the onshore case is presented in Fig. 12 (cross-section lo-
cisely known but is thought to be within a few minutes of 1700 UTC. The fronts continue to propagate westward at 1714, with the kinematic front merging with more of HCR 1 in the northern portion of the dual-Doppler domain at 1732 UTC. The mechanism that is responsible for apparently intensifying the front as the merger process occurs is discussed next.

(ii) Merger of like-sign vortices

Figure 14 shows three vertical cross sections near the area where the front and HCR 1 are merging (see Fig. 11a for cross-section locations). In Fig. 14a, negative and positive $\omega$ maxima are observed to the left and right of the reflectivity maxima, respectively, associated with the kinematic front and HCR 1. The negative circulation with HCR 1 has intensified slightly and grown larger in $JJ'$ while a larger region of positive $\omega$ exists behind the kinematic front, however, two distinct maxima are visible, similar to $II'$. Only one positive and negative circulation exists in $KK'$ and both are much larger spatially than those in $II'$. It is visually apparent in these cross sections that the like-sign circulations associated with the front and HCR 1 are merging to create larger circulations. After the merger process has occurred, the frontal circulation is comprised of a positive and negative circulation located behind and ahead of the kinematic frontal location, respectively. The aspect ratio of the frontal circulations in $KK'$ is approximately 6.4, more than twice the value of 3.0 for the HCRs. These values are well within the range reported by previous investigators (Miura 1986).

The behavior and interaction of like-sign vortices has been studied extensively in numerical fluid dynamical studies (e.g., Fornberg 1977; McWilliams 1984, 1990; Overman and Zabusky 1982; Melander et al. 1987a,b; 1988) and laboratory experiments (e.g., Fujiwhara 1921, 1923; Freymuth 1966). Within the atmosphere, vortex merging has recently been hypothesized to be important for the evolution and movement of tropical cyclones (Holland and Lander 1993; Lander and Holland 1993; Ritchie and Holland 1993; Holland and Dietachmayer 1993), the formation of hurricane spiral bands (Guinn and Schubert 1993), and the evolution of HCR spacing and intensity (Sykes et al. 1988). It is well known that when two like-sign vortices of similar size and intensity come close enough to each other, they will merge, coalesce to form a single larger vortex. The resulting circulation will be larger, however, the magnitude of the resultant vorticity will remain approximately the same. This is consistent with the observations in Fig. 14.

It is believed that the merging process of like-sign vortices is the primary mechanism for why the sea-breeze front intensified along extended sections during the onshore flow event. Figure 15 presents a cross sec-
Fig. 9. Surveillance scans of radar reflectivity (dBZ) with values greater than 4 and 8 dBZ, shaded gray and black, respectively, from the CP-4 radar. The left panel also shows the winds (full barb—5 m s\(^{-1}\), half barb—2.5 m s\(^{-1}\)), temperature (°C), and dewpoint temperature (°C) from PAM stations at 1707 UTC. The long, dashed lines delineate HCRs. The right-hand panel shows the location of HCR 1 and the kinematic sea-breeze front as determined from the dual- and single-Doppler analysis. PAM data at 1708 UTC are also plotted.

A moment along the length of HCR 1 and the kinematic sea-breeze front (see Fig. 11a for cross-section location). Also shown are the calculated values of circulation for the positive and negative circulations associated with HCR 1 and the front. In this study, the circulation of a vortex is defined as

\[ \Gamma = \bar{\omega} A, \]  

where \( \Gamma \) is the circulation, \( \bar{\omega} \) and \( A \) are the mean vorticity and area of the vortex, respectively. The area is encompassed by the \( \pm 1 \times 10^{-3} \text{ s}^{-1} \) contour of the component oriented parallel to the front or HCR 1. The above definitions may seem arbitrary, however, it is believed that they will not affect the major conclusions since it is desirable to note the relative change in circulation along the cross section.

It is clear that the merged portion of the front is more intense since the reflectivity and convergence fields show an increase from the values along HCR 1. The vertical velocity magnitude increases slightly and the depth of the updraft increases since the average maximum height of the \( W = 0 \text{ m s}^{-1} \) contour, the approximate vertical extent of the updraft, was approximately 400 m higher along the merged portion of the front than along HCR 1 (not shown). The larger circulations induce stronger and deeper updrafts between the vortices, providing additional impetus for cloud development. It should be noted that cloud 0, the only cloud not on the merged portion of the front in Fig. 15, exhibited the least vertical growth of all clouds triangulated at 1715 UTC. This result was confirmed by following the life history of the clouds with video taken at the Photo South site. Further theoretical verification that the larger circulations will intensify the front is now presented.

(iii) Irrotational flow theory and the complex potential

Given a 2D irrotational flow of a nonviscous incompressible fluid, it is possible to describe certain flow
Fig. 10. Results of the dual-Doppler synthesis at approximately 1715 UTC. Plotted in all four panels are the reflectivity field in gray at a height of 0.7 km AGL along with the kinematic (solid barbed) and thermodynamic (dashed barbed) frontal positions. The dash-dot line delineates HCR 1 shown in Fig. 9. (a) Horizontal winds (m s⁻¹) and (b) convergence at 0.1 km AGL contoured every 2 × 10⁻³ s⁻¹. The 0 contour is not plotted. (c) Horizontal winds and (d) convergence at 1.0 km AGL. Wind data from the PAM stations are also plotted in (a).

fields in terms of an analytic function, the complex potential \( w(z) = \phi - i\psi \) within the region of the complex plane \( z = x + iy \) (e.g., Batchelor 1967) where \( \phi \) is the velocity potential and \( \psi \) is the stream function. For a distribution of \( j \) point vortices with strengths \( \Gamma_j \) at the locations \( Z_j \), the complex potential can be written as

\[
w = -\frac{i}{2\pi} \sum_j \Gamma_j \ln(Z - Z_j).
\]  
(2)

From the complex potential, the complex velocity \( \bar{q}(z) \) at some point in the fluid is given by

\[
\bar{q}(z) = \frac{DW}{DZ}.
\]  
(3)

Substituting (2) into (3), the complex velocity can be expressed as

\[
\bar{q}(z) = \frac{1}{2i\pi} \sum_j \frac{\Gamma_j}{Z - Z_j}.
\]  
(4)

For two counterrotating point vortices near a lower boundary, as illustrated in Fig. 16, this equation can be used to determine the induced velocity of the fluid by the vortices, which will be a maximum directly between the circulations. Furthermore, making the assumptions that the boundary layer circulations associated with the HCRs and sea-breeze front during the onshore flow case are two dimensional, the flow is irrotational and incompressible, and ignoring the fact
that the fluid is stratified, Eq. (4) can be used to roughly estimate the relative increase in vertical velocity along the cross section $HH'$ shown in Fig. 15 due to the frontal–HCR merger. It is recognized that these assumptions grossly oversimplify the real flow in the boundary layer, however, the important processes should not be greatly affected. The input parameters for this calculation and corresponding results are shown in Table 3.

Fig. 11. Horizontal cross sections showing the relationship between the clouds, reflectivity, and vertical velocity field. (a) Reflectivity (gray) and vertical velocity (contoured every 1 m s$^{-1}$) fields at 0.7 km AGL are plotted. (b) Vertical velocity (gray) and cloud positions with number identifiers are shown.

Fig. 12. Vertical cross sections showing the reflectivity (dBZ) and winds in the plane of the cross section. Frontal and cloud positions are also depicted.
It is apparent that the calculated increase in vertical velocity using (4) is consistent with the increase observed in the reflectivity, convergence, and vertical velocity fields, further evidence that the observed intensification of the front is due to the merger of like-sign vortices.
2) Periodic Nature of the Cloud Development

Although the intensified portions of the front are preferred sections for cloud development, variability is observed as clouds initiated at periodic locations along these sections (e.g., Fig. 11). The periodicity is quite similar to what is observed to occur along the HCRs. Indeed, the wavelength of the periodicities (or enhancements) as observed in the radar reflectivity field along HCR 1 was about 3.6 km along the unmerged portion of the HCR. This is consistent with the values of 3.5 and 3.6 km for the reflectivity and convergence maxima, respectively, along the merged section of the front in the south lobe. Hence, the clouds along the front appear to initiate at the location of the periodic enhancements on the HCR, which intensified during the merging process.

c. The thermodynamic sea-breeze front

A feature that has been consistently analyzed at the westward edge of the negative circulation ahead of the kinematic front in Figs. 10–14 is the thermodynamic sea-breeze front and is defined as follows:

Thermodynamic sea-breeze front: the location within the CBL where the mean thermodynamic properties of the air begin to differ from those in the undisturbed ambient air over land. Similar frontal structure where the thermodynamic and kinematic boundaries are spatially displaced from each other has not been previously documented in association with laboratory-produced (Simpson 1987) or atmospheric density currents, such as gust fronts (e.g., Charba 1974; Wakimoto 1982; Mueller and Carbone 1987; Weckwerth and Wakimoto 1992). Reible et al. (1993), however, have
shown aircraft penetrations through a developing sea-breeze front under calm synoptic flow conditions documenting the existence of a transition zone between the ambient flow and sea-breeze air. This zone (approximately 4–6 km wide) possessed thermodynamical properties representative of a mixture between the two different air masses and was thought to be created by the turbulent mixing occurring in the convective boundary layer. However, other than one aircraft penetration, detailed observations of this zone were not presented.

The structure of the thermodynamic front is revealed by the data recorded onboard the Wyoming King Air. One of the flight tracks as the aircraft flew westward at an altitude of 0.54 km AGL is shown in Fig. 17. As the aircraft penetrated through the kinematic sea-breeze front, a slight clockwise wind shift occurred (Fig. 18). In Figs. 18b and 18c, a smooth transition of the ther-
modynamic properties from the cool, moist sea-breeze air to the warm, drier ambient air occurs between the kinematic and thermodynamic fronts. Interestingly, $\theta_v$ remains essentially constant. The region between the kinematic and thermodynamic fronts is called the thermodynamic frontal zone and is comprised of a mixture of the ambient and sea-breeze air masses. This zone extends approximately 14 km to the west of the kinematic front along the flight track in Fig. 17.

The flow structure within the thermodynamic frontal zone is shown in Fig. 18a (see Fig. 17 for cross-section location). Similar to Fig. 12, a portion of the sea air is propagating westward of the kinematic sea-breeze front. It is transported vertically by the frontal updrafts and becomes part of the negative circulation ahead of the front. The sea air will travel farthest to the west approximately the same distance as the westward extent of the frontal negative circulation and mixes with the air ahead of the kinematic front. This distance is approximately 13 km, apparent in the horizontal vorticity and wind fields, and is consistent with the westward extension ($\sim$14 km) of the thermodynamic frontal position observed in the aircraft data. Hence, the thermodynamic sea-breeze front is located at the westward position of the frontal negative circulation, along the portion of the front that has merged with HCR 1. It is recognized that surface flux variability, which may be induced by heterogeneity in land use, soil moisture, and/or irrigation practices, may create thermodynamic fluctuations similar to those shown in Fig. 18. However, penetrations through the thermodynamic frontal zone by the Wyoming King Air at different times revealed that the frontal zone moved westward at approximately the same velocity as the kinematic front, consistent with the hypothesis that the frontal zone is generated by the interaction of the sea breeze and HCRs.

Another observation to be made in Figs. 17 and 18 is the absence of well-defined HCRs in the frontal zone. The negative circulation ahead of the front seemingly is dominant within this region. It is hypothesized that other scales of convection are occurring in the thermodynamic frontal zone as suggested in the reflectivity field and shown by Walter and Overland (1984), however, the signal is not of sufficient magnitude to be adequately resolved by the dual-Doppler technique. Perhaps Doppler lidar or higher-resolution radars such as the FM-CW radar discussed by Eaton et al. (1993) would elucidate the detailed kinematic structure of the thermodynamic frontal zone.

The vertical extent of the thermodynamic front is analyzed only to the height of the aircraft track in Fig. 18. Although it is hypothesized that it extends to the top of the CBL, no aircraft data exists at these levels to confirm this idea.

The results discussed in this section are schematically illustrated in Fig. 19. A portion of the kinematic front has merged with part of HCR 1. Deeper cloud development is occurring along this enhanced section of the front at the locations of the periodic enhancements, which were present along HCR 1 prior to the merger. The thermodynamic front is located at the westward edge of the frontal negative circulation. The thermodynamic frontal zone is located between the kinematic and thermodynamic fronts.

![Fig. 17. Horizontal cross section showing the UW King Air flight track (black, solid line) and aircraft-measured horizontal winds (length scale is the same as for the dual-Doppler winds). Radar reflectivity field is plotted in gray along with the dual-Doppler-derived winds at 0.4 km AGL at approximately 1750 UTC.](image)
5. Summary

The three-dimensional structure of two sea-breeze events during CaPE has been documented. Alongfrontal variability was observed and created by the interaction with the HCRs existing in the ambient air, contrary to previous studies on HCRs involving a frontal boundary that have focused on the HCRs forming in the cold air. Indeed, during the onshore flow case, HCRs were observed in the moist air behind the sea-breeze front as shown in Fig. 20. This result is consistent with a laboratory experiment performed by Matsuato et al. (1983) where HCRs formed in the cool air behind the front, which was propagating in a convectively active boundary layer. HCRs behind the front during the offshore case were not prominent since, assuming that they are confined to the depth of the cool sea-breeze air (∼440 m), their circulations would be much shallower than for those observed behind the front during the onshore case where the depth of the cool air is approximately 950 m at the time in Fig. 20. The HCRs behind the front for the onshore event do not appear to affect the frontal structure since no intersection points or mergers with the front were clearly observed.

During the offshore flow case, nearly perpendicular intersections with the HCRs were observed with undulations in the front generally occurring at the intersection points. The HCRs appeared to change their orientation and intensify within 5–6 km of the front. This observation was consistent with the apparent change in orientation and intensification of the shear vector close to the front, being altered by the slowing of the low-level winds normal to the front by nonhydrostatic pressure effects.

The HCRs were tilted 20° from the horizontal by the frontal updrafts and lifted 0.5–0.6 km over the front, having the effect of creating deeper updrafts and clouds at the intersection points. Furthermore, clouds forming
at periodic intervals on the HCRs were intensified as they propagated over the front. It appeared that only the positive roll circulations propagated rearward of the

surface frontal position to a maximum observed distance of about 2.0 km. The HCR circulations mixed the two air masses on the sea-breeze front head, similar to

Fig. 19. Schematic diagram showing the interaction between the sea-breeze front and HCRs and how it relates to cloud development on 6 August 1991. The kinematic and thermodynamic sea-breeze fronts are delineated by the heavy, solid and white, barbed lines, respectively. The positive head circulation is lightly shaded. The horizontal vorticity vectors associated with the roll circulations are shown. Clouds along the HCRs and kinematic front are shaded gray. Low-level winds are shown by the white, 2D arrows. The approximate aircraft track position discussed in Figs. 17 and 18 is also depicted.

Fig. 20. Horizontal cross section showing the existence of HCRs ahead of and behind the kinematic sea-breeze front. Shown in both panels are the reflectivity field (gray) along with the kinematic sea-breeze front location. (a) The short- and long-dashed lines lie along the HCRs behind and ahead of the front, respectively. (b) Dual-Doppler-derived horizontal winds (m s\(^{-1}\)) at 0.1 km AGL, UW King Air horizontal winds at approximately 0.12 km AGL, and wind data from the PAM stations are plotted.
the effect created by K–H billows on the head of laboratory-produced and atmospheric density currents. As a result, the turbulent wake region behind the sea-breeze front may be created by both K–H billows and HCRs atop the frontal boundary. Due to the lack of scatterers, it is not known if K–H billows formed along the sea-breeze front during the offshore flow event.

The HCRs were oriented nearly parallel to the frontal orientation during the onshore flow case. Consequently, extended sections of the front merged with HCRs existing in the ambient air and these sections intensified, being preferred locations for cloud development. This intensification was a result of the merger of like-sign vortices associated with the front and HCR. Clouds initiated along these sections at the locations of the periodic enhancements on the HCRs. The results shown for the onshore case are similar to numerical simulations by Sykes et al. (1988) who showed that HCR spacing and intensity increase after two HCRs merged together.

Documentation of a thermodynamic front extending up to 14 km ahead of the kinematic front was presented. It delineates the position where the mean thermodynamic properties of the air differ from the ambient flow and is created by the interaction of the HCRs and sea breeze. The region between the two fronts is called the thermodynamic frontal zone and has thermodynamic properties consisting of a mixture of the ambient and sea-breeze air masses. The width of the frontal zone varied, being the widest along the sections of the front, which had merged with the HCR.

The kinematic characteristics between the two sea-breeze events were also discussed. The offshore flow case exhibited stronger low-level convergence, and larger reflectivities and vertical velocities than the onshore flow case. The source air creating the clouds along the front during the offshore (onshore) flow day originated in the warm (cool) air. Furthermore, only the offshore flow case exhibited a kinematic frontal structure similar to laboratory-produced density currents.

Future work will concentrate on a climatological survey of sea-breeze frontal characteristics under different flow regimes (i.e., onshore, offshore, and parallel flow). A modeling effort is also being pursued to further understand the structure and dynamics of sea-breeze fronts forming under different flow regimes and their interaction with boundary layer convection.

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APPENDIX

Dual-Doppler Methodology and Uncertainties in the Wind Field

Each radar scan was carefully edited using the NCAR Research Data Support System (RDSS) interactive computer software (Oye and Carbone 1981) to remove ground clutter, sidelobe contamination, and second-trip echoes. A careful ground clutter check was also performed to ensure the ground-relative position of the data is as precise as possible. This was accomplished by plotting the radar reflectivity data for each gate recorded by a low-level scan onto a topographic base map. The locations of high-tension power lines were confirmed using this method. Uncertainties in the corrected data are 75 m in range and 0.2° in azimuth.

On 6 and 12 August the PRF (pulse repetition frequency) was set to 1250 s⁻¹ for both radars, and the number of samples used for each beam estimate was 64. The scanning rates for both radars on 6 and 12 August were 7.98° s⁻¹ and 8.95° s⁻¹, respectively. This yields an azimuthal radar resolution of 0.41° and 0.46° for 6 and 12 August, respectively. As the beamwidth for both radars was 0.9° and the elevation angle step was greater than or equal to 0.5°, it was decided to interpolate the data onto a Cartesian grid with spacing of Δx = Δy = Δz = 300 m using a Cressman (Cressman 1959) filter with a radius of influence equal to or less than 400 m (depending on the distance from the radar). No space–time correction was made to compensate for the frontal or HCR motion since they were moving slowly and at different velocities. On 12 August, the front was propagating at 1.2 m s⁻¹ toward 270° while the HCRs were propagating at about 2.5 m s⁻¹ toward 40°. On 6 August, the front propagated at a speed of 1.5 m s⁻¹ toward 237° while the HCRs moved toward 320° at 3.2 m s⁻¹. Hence, it was decided to present analyses for both days in the frame of reference relative to the ground.

The data were synthesized using CEDRIC (custom editing and display of reduced information in Cartesian space). According to Carbone et al. (1985), energies

<table>
<thead>
<tr>
<th>Analysis time (UTC)</th>
<th>Number of clouds collocated with a vertical velocity maximum</th>
</tr>
</thead>
<tbody>
<tr>
<td>12 August 1950</td>
<td>8/9 (89%)</td>
</tr>
<tr>
<td>1911</td>
<td>12/14 (86%)</td>
</tr>
<tr>
<td>6 August 1715</td>
<td>8/9 (89%)</td>
</tr>
<tr>
<td>1900</td>
<td>18/19 (95%)</td>
</tr>
</tbody>
</table>
with wavelengths less than 2.4 km are not resolvable in this study. Consequently, a two-step Leise filter (Leise 1982) was applied so that energies with spatial scales of 1.2 and 2.0 km are eliminated and severely damped, respectively.

Vertical velocities were derived by integration of the horizontal convergence field with the anelastic mass continuity equation and using a variational adjustment scheme (O’Brien 1970). The boundary conditions chosen were $w = 0 \text{ m s}^{-1}$ at the lower and upper boundaries. A fractional lower boundary condition (e.g., Kingsmill and Wakimoto 1991) was assumed since the lowest grid level was chosen at 0.1 km AGL. Setting $w = 0 \text{ m s}^{-1}$ at the upper boundary may not necessarily be accurate, however, it was deemed the best objective choice. A density weighting function was employed (Foote and du Toit 1969) using a scale height of 10 km.

There are several sources of error that contribute to uncertainties in the derived three-dimensional wind field and have been described in detail (e.g., Miller and Strauch 1974; Doviak et al. 1976; Miller and Kropfli 1980; Wilson et al. 1984). Wind fields derived from clear-air echoes have only recently been attempted (e.g., Brandes and Rabin 1991) and the uncertainties in the vertical velocity estimates have been discussed by Wilson et al. (1994).

For the two cases described herein, vertical velocity uncertainties may be as large as $3.5 \text{ m s}^{-1}$, due to statistical errors in the radial velocity estimates. As these errors are random, the actual error may be much less. Geometrical errors will be small since elevation angles for the scans used were less than 10°.

A better assessment of the three-dimensional wind field accuracy may be determined by making detailed comparisons with photogrammetrically determined cloud positions and aircraft flights. Table A1 shows the cloud–W relationship at the four analysis times presented by WA94 (their Figs. 9 and 18). Tabulated are the number of clouds that were associated with a local W maximum. Clouds were eliminated from the analysis if it was not observed within the time of the next volume scan since it was considered to be decaying and not associated with an updraft. The results in Table A1 show a close relationship between W maximum and cloud locations.

The aircraft and dual-Doppler comparison is shown in Table A2 for two different flights on 6 August by the UW King Air. The correlation coefficients for the three components of the wind field are shown. A high correlation exists for all three components. Hence, the results presented in Tables A1 and A2 suggest that the synthesized wind field from clear-air echoes may be used for meaningful interpretation of boundary layer convection and the sea breeze.

**Table A2. Aircraft/dual-Doppler wind field comparison.**

<table>
<thead>
<tr>
<th>Analysis times (UTC)</th>
<th>Altitude (km AGL)</th>
<th>Correlation coefficients</th>
</tr>
</thead>
<tbody>
<tr>
<td>1744:00–1751:00</td>
<td>1750:34–1753:30</td>
<td>$\rho_u$, $\rho_v$, $\rho_w$</td>
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<td>1756:40–1759:00</td>
<td>1906:35–1909:50</td>
<td>0.54, 0.79, 0.93, 0.80</td>
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<tr>
<td>1902:40–1909:00</td>
<td></td>
<td>0.85, 0.74, 0.83, 0.82</td>
</tr>
</tbody>
</table>

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