Numerical Simulation of the Interaction between the Sea-Breeze Front and Horizontal Convective Rolls. Part I: Offshore Ambient Flow

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ABSTRACT

A three-dimensional, cloud-resolving model is used to investigate the interaction between the sea-breeze circulation and boundary layer roll convection. Horizontal convective rolls (HCRs) develop over land in response to strong daytime surface heating and tend to become aligned parallel to the vertical wind shear vector, whereas the land–sea heating contrast causes the formation of the sea-breeze front (SBF) along the coastline. The ability of HCRs to modulate the along-frontal structure of the SBF is examined, complementing and extending previous observational and numerical studies.

Three simulations are discussed, the first two demonstrating that the model can produce both phenomena independently. The third is initialized with offshore mean flow and vertical shear perpendicular to the coastline, and results in a sharply defined, inland-propagating SBF that encounters HCRs aligned perpendicular to it. Before the interaction takes place, the SBF is nearly two-dimensional and devoid of along-frontal variability. Its subsequent encounter with the HCRs, however, causes enhanced (suppressed) convection at frontal locations where HCR roll updrafts (downdrafts) intersect. The suppressing effect of the roll downdrafts seems particularly striking. The interaction as it relates to vertical and horizontal motion, vorticity, and the cloud field are discussed. In future work, a similar simulation with HCRs oriented parallel to the SBF will be analyzed. These results provide further evidence that HCRs can play an important role in the initiation and modulation of convection along a sea-breeze front.

1. Introduction

The sea-breeze circulation (SBC) is a mesoscale phenomenon driven by daytime heating contrasts between land and water surfaces and a subject of long-standing interest (e.g., Davis et al. 1890). It has been observed with a variety of tools including aircraft measurements (e.g., Hatcher and Sawyer 1947; Fisher 1960; Reible et al. 1993), pilot balloons (e.g., Frizzola and Fisher 1963; Yoshikado 1990), satellite imagery (e.g., Wakimoto and Atkins 1994), and Doppler radars (Carroll 1989; Intrieri et al. 1990; Wakimoto and Atkins 1994; Atkins et al. 1995). Laboratory studies (e.g., Simpson 1969, 1987; Simpson and Britter 1980; Mitsumoto et al. 1983) and analytical models (e.g., Haurwitz 1947; Schmidt 1947; Dalu and Pielke 1989) have also contributed extensively to our present knowledge. Many numerical investigations have been conducted, including those of Pielke (1974), Pielke and Mahrer (1978), Nicholls et al. (1991), Bechtold et al. (1991), Sha et al. (1991), and Arritt (1993), to name but a few. Several of these studies have shown that the SBC’s basic structure and dynamics can be captured in two-dimensional models.

The sea-breeze front (SBF) refers to the leading edge of the cool marine air, and much attention has been paid to its propagation speed and inland penetration. Average speeds of 10–20 km h⁻¹ have been reported (Clarke 1955) with an acceleration often noted late in the day (Simpson et al. 1977). The front has been known to reach as far as 300 km inland (Atkinson 1981), though 30–100-km distances are more typical. Observations have long indicated that onshore flow promotes inland penetration but results in a weaker and more diffuse front. In contrast, offshore flow can result in a sharply defined front capable of generating greater frontal lifting, even though in these cases the SBF remains restricted relatively closer to the coastline. Arritt (1993), however, showed that excessively strong opposing flow keeps the SBF from propagating inland at all, resulting in weakened frontal lifting because the SBF remains in the stably stratified marine environment.

Convergence at the SBF may cause a band of clouds to form along the frontal boundary, and SBF-associated clouds over Florida have been studied both observationally (e.g., Burpee and Lahiff 1984; Wakimoto and Atkins 1994) and numerically (Pielke 1974; Pielke and Mahrer 1978; Nicholls et al. 1991). Aside from its ob-
rious meteorological significance and impact on millions of coastal inhabitants, forecasters are concerned with the movement and strength of the SBF for its role in the distribution of pollutants (Stevens 1975), airborne allergens (Raynor et al. 1974), and crop-threatening insects (Berry and Taylor 1968).

Nearly 50 years ago, Leopold (1949) illustrated the ability of SBFs to induce towering cumuli at discrete locations along the frontal boundary. The source of this along-line variation was not understood, and an increasing amount of attention is being paid to the study of variability along convective boundary layer (CBL) convergence zones in general, and density-driven currents in particular, both of which provide favorable regions for convective initiation (Purdom 1982). Simpson and Britter (1980) showed that “clefts and lobes,” small-scale (~1 km) variations on the propagating head circulation produced by the influences of friction and inertia, could form along their laboratory-produced density currents. This type of variability along the SBF has not yet been documented through observations and there are significant differences between laboratory-produced density currents and the sea-breeze current. In the former, a dense fluid is impulsively released into a less dense fluid while the SBC develops slowly as the cross-shore density difference builds up over the course of the day.

Some studies have attempted to tie convergence zone variation to the presence of boundary layer phenomena such as horizontal convective rolls (HCRs). Mason and Sykes (1982) pointed out that HCRs can act as small obstacles to the ambient flow, thereby exciting internal gravity wave activity. Balaji and Clark (1988) showed that these waves can influence convective initiation but their effect is highly dependent on the their orientation relative to the convergence zone. If the alignment is parallel, for example, the result can be the periodic generation of convective cells, though such periodicity is not necessarily predicated upon such an interaction (Fovell and Dailey 1995). Gravity waves typically have wavelengths several times longer than that of the HCRs inducing them and vertical motions in excess of 1 m s$^{-1}$ (Crook et al. 1991). Wilson et al. (1992) observed the Denver convergence zone and its interaction with HCRs, finding that the intersection between HCR updrafts and the convergence zone resulted in discrete along-line vertical velocity maxima, which appeared to be responsible for the deepest convection observed. Others have noted variability in the form of vortices located along a gust front boundary (McCaul et al. 1987), and these local vorticity maxima have been associated with nonsupercell tornadoes (Wakimoto and Wilson 1989).

The observational studies that most closely apply to this modeling effort are Wakimoto and Atkins (1994; hereafter WA) and Atkins et al. (1995; hereafter AWW), both of which focused on the interaction between the SBF and pre-existing roll convection over eastern Florida during the Convection and Precipitation/Electrification Experiment (CaPE) program. AWW and WA discuss observations taken during two sea-breeze events, occurring on 12 and 6 August 1991. The mean flow was onshore and nearly parallel to the coastline on 6 August, whereas the other day had an offshore flow nearly perpendicular to the coast. During both, roll convection existed in the environment out ahead of the front, which modulated convective activity along the front as the SBF propagated into the HCR environment. When the front was oriented nearly perpendicular to the rolls (12 August), along-frontal variability occurred at distinct points along the frontal boundary. However, when the rolls were nearly front-parallel (6 August), the HCRs seemed to merge in segments along the front. The vertical motion and resulting cloud field associated with these sea-breeze events were closely tied to the interaction of the SBF with roll convection. Our goal is to simulate these events with a three-dimensional cloud-resolving model to further elucidate the role of HCRs in modulating SBF convection. The offshore case, with HCRs oriented perpendicular to the SBF, is specifically considered herein; the front-parallel case will be examined in Part II of this work.

The discussion is organized as follows. Section 2 is an overview of the numerical model used in this study. In section 3, two benchmark simulations (“sea-breeze only” and “roll only”) are introduced. In section 4, a simulation in which the model is initialized with the vertical shear vector perpendicular to the coastline is discussed in depth. A discussion of the results is the focus of section 5. Finally, section 6 provides a brief summary.

2. Model

This study employs an enhanced version of the Klemp–Wilhelmson (1978; hereafter KW) 3D cloud model, which discretizes the compressible equations of motion in a nonhydrostatic setting. The hydrostatic assumption is deemed inappropriate in any study of the SBF under conditions of intense surface heating and instability (Martin and Pielke 1983). The KW model formulation is fully discussed in Klemp and Wilhelmson (1978) and Wilhelmson and Chen (1982). Below, the model setup and initialization are discussed. A summary of the model enhancements relating to parameterizations of the radiation budget and surface fluxes is provided in the appendix.

a. Model setup

The model uses a commonly employed staggered grid arrangement (the Arakawa “C” grid) with 50 vertically stretched levels to economically enhance resolution in the lower troposphere. With the model top at 18 km and stretching parameters of $cc_1 = 0.3$ and $cc_2 = 0.7$ (see Wilhelmson and Chen 1982), the first thermodynamic
grid level is located ≈55 m above the surface and the model has seven grid points in the lowest 1 km. The model top is rigid, though testing with the Klemp–Durran (1983) radiation condition revealed no tropospheric sensitivity to the handling of this boundary. The lower boundary is flat but not free-slip.

The horizontal grid is unstretched with either open or periodic lateral boundaries, depending on the simulation and direction \((x\ \text{vs}\ \ y)\). The resolution employed is also simulation-dependent. The \(x\) and \(y\) directions are taken as west–east and north–south, respectively. When included, the coastline is linear and aligned along the \(y\) axis, with the ocean occupying the western 25\% of the domain. Specific grid details will be presented along with each simulation.

Advection is handled with the fourth-order horizontal and second-order vertical centered scheme described in Wilhelmson and Chen (1982). The KW model employs “time splitting,” which integrates the acoustically active and inactive terms on separate time steps, taken to be 1 and 4 s, respectively. The subgrid turbulence scheme is 1.5 order and based on the local value of the Richardson number, as in KW. The scheme requires specification of a mixing length, which is taken to be a function of grid resolution. The turbulent Prandtl number is taken to be 3, based on Deardorff’s (1972) findings. The reader is referred to KW for further information.

b. Model initialization and assumptions

All simulations were started at 0600 local standard time (LST), the approximate time of sunrise, with the initial temperature and mixing ratio profiles shown in Fig. 1. These profiles were adapted from early morning Florida soundings taken during mid-August and will be subject to considerable modification during the day, especially over land, due to solar heating. The temperature profile lacks a capping inversion, as this is difficult to include without even higher vertical resolution. Though this lack does not compromise the results, it does permit the CBL to deepen more during the afternoon than it would have been able to had such an inversion been initially present.

The wind profiles, shown in Fig. 2, represent offshore mean flow, perpendicular to the coastline and directed from land to sea. Note the winds possess vertical speed shear of \(5 \times 10^{-3} \, \text{s}^{-1}\) in the lowest kilometer but lack directional shear, and thus the mean flow and vertical shear vectors are parallel. Thus, all further mention of offshore mean flow also represents an offshore-directed vertical shear vector. Observational and theoretical work has shown that without sufficient shear, whether through the boundary layer (e.g., Tsuchiya and Fujita 1967; Asai 1970a,b; 1972) or confined very near the surface (e.g., Pennell and LeMone 1974; Kristovich 1991; Weckwerth et al. 1997), convection will be random and cellular rather than organized into the bands represented by the HCRs. Rolls have been known to form in both stable and unstable environments (e.g., Christian and Waki- moto 1989; Ferrare et al. 1991), but they are generally observed in nearly neutral boundary layers, presumably because the mixing they accomplish contributes toward the establishment of a well-mixed planetary boundary layer (PBL).

These profiles are applied throughout the domain, which means the land and sea surface temperatures are initially equal (see appendix). However, during the first 1–2 simulated hours the model does develop a realistic boundary layer and land–sea temperature contrast. Also, our focus is placed on period from 5 to 10 h after model start, by which time the atmosphere has evolved considerably. The radiation scheme described in the appendix utilized typical mid-August values applicable for
precipitation is deactivated. To excite fully 3D motions over the land, random multipliers (between 0.95 and 1.05) are applied to the surface temperature prognostic equation each time step. The result is a ±5% normally distributed perturbation in surface heating. Similar methods have been used in prior modeling studies (e.g., Balaji and Clark 1988). Most simulations were run out to 1800 LST (6:00 P.M.) by which time the solar forcing has greatly diminished.

To improve the efficiency of model calculations, decrease the required computing resources, and eliminate complicating factors, the following assumptions are made.

1) The Coriolis effect is neglected. Although known to shift the sea-breeze winds a few degrees by afternoon (Simpson 1994) and reduce the circulation’s horizontal extent and intensity (Niino 1987; Dalu and Pielke 1989), the Coriolis effect is not seen to be crucial to the SBC–HCR interaction, though it should be incorporated in future work.

2) Topography and coastline variation are ignored. Local topography and coastline curvature can clearly influence the sea-breeze flow as well as introduce along-frontal variability (e.g., Findlater 1971; Pielke 1974; Simpson 1994), but we wish to isolate the variability that is generated independently of these effects.

3) Precipitation is deactivated. Precipitation can play a significant role in sea-breeze events. For example, it is known that evaporational cooling and water loading can have marked effects on the pressure gradient across the front (e.g., Nicholls et al. 1988). Though the model domain is designed to accommodate both shallow and deep convection, if the latter develops, evaporative cooling of hydrometeors can obliterate the SBF as well as cause a considerable change in boundary layer stability (Nicholls et al. 1991). To eliminate the deleterious effects precipitation can have on the surface features with which we are most concerned, the production of rainwater is deactivated.

3. Benchmark simulations

Herein, we briefly examine two benchmark simulations, representing sea-breeze only (SBO) and roll only (RO) cases. These simulations not only demonstrate the model’s ability to realistically produce each phenomenon, but also are useful for assessing the effects each circulation has on the atmospheric state in isolation. The latter point will be shown to be useful in section 4. For the SBO run, the model is run in its two-dimensional (2D) configuration, thereby preventing the development of HCRs. The RO simulation is 3D but lacks an ocean surface, thereby allowing rolls to develop without marine influence.

The principal goal here is to produce an intense sea-breeze front as well as vigorous convective rolls so that when these two phenomena are combined (in section 4, below) the interaction between them is brought into high relief. This is accomplished by giving the land surface urban characteristics, that is, with little moisture available (see appendix). With evaporational cooling over land restricted, the land surface temperature can get hot and the PBL can become quite deep. In both model geometries, the inland surface heat flux over land well removed from the sea breeze (if present) reaches a maximum value of ≈186 W m⁻² around noon, a not unreasonable value for this location and time of year (T. M. Weckwerth 1998, personal communication). Further details concerning these simulations may be found in Dailey (1996).

a. SBO benchmark

For this SBO run, the model domain is 141 grid points wide with 1-km horizontal resolution and open lateral boundaries. Figures 3 and 4 depict the evolution of the SBC at 1-h intervals during the period from 0900 to 1200 LST. Recall that the land and sea surfaces start with the same temperature at sunrise (0600 LST). Subsequent solar heating of the land surface makes the marine air relatively more dense, but it takes time for this air to overcome the offshore mean winds and flow inland. By 0900 LST, the SBF may be detected in the perturbation potential temperature field (Fig. 4a) but it is very weak and located offshore. In the next hour, onshore flow appears but is primarily confined to the immediate coastal area (Fig. 3b). By this time, the intensifying horizontal convergence has established a forced updraft at the kinematic frontal boundary (labeled “KF”). This boundary, defined by the location of zero ground-relative cross-shore flow (i.e., $u = 0$) is located in close proximity to the front’s thermodynamic boundary (Fig. 4b), as expected. Subsidence has developed on the seaward side of the forced updraft.

By 1100 LST, the sea-breeze horizontal flow has reached about 6 m s⁻¹ and the front has propagated almost 15 km inland, though the strongest onshore flow is still confined to the coast (Figs. 3c, 4c). Both the frontal forced updraft and the seaward subsidence have continued to intensify and expand, and the altitude of its maximum lifting has risen. By this time, the frontal lifting has caused the establishment of a shallow cloud above the SBF; the base of which is just under 3 km above the ground. By noon (Figs. 3d, 4d), the onshore flow has developed an elevated head behind the kinematic boundary and (though not shown) the cloud over
the still intensifying SBF has become ≈ 2 km deep. Due to surface friction, the strongest onshore flow immediately behind the SBF is actually located about 150 m above the ground (see “×” on Fig. 3d), rather than right at the surface; this location rises as time progresses (not shown). At this time, the SBF is propagating at 11 km h⁻¹, and an acceleration (to >13 km h⁻¹) does occur later in the afternoon (not shown). Both the propagation speed and the inland penetration of the modeled SBF appear reasonable given the range of observations cited in the introduction.

This simulation is somewhat analogous to Arritt’s (1993) “moderate opposing flow” regime, which produced the strongest lifting along the SBF. He states that frontal lifting should be most intense when the mean flow allows it to reside in the neutrally stratified at-
mosphere just ahead of the front. Though the return flow is embedded within the offshore mean wind, it is clear that air from the east (land) side of the domain is forced up and over the SBF. This is especially true in this strictly 2D simulation since along-coastal flow cannot develop. The marine air is moderated as it passes over the radiatively heated land surface. The SBC continues intensifying over the next several hours (not shown), weakening as the land surface heating wanes during the afternoon. The SBF-induced cloud never develops into deep convection. Finally, it is noted that the SBO simulation was not found to be sensitive to the domain width chosen; further expansion of either the land and/or sea subdomains had virtually no effect on the structure and evolution of the model SBC.

b. RO benchmark

Horizontal convective rolls (HCRs), depicted in Fig. 5, are convectively induced vortex couplets aligned
along a horizontal axis in the boundary layer that act to mix an otherwise unstable atmosphere by lifting warm conductively heated air near the surface and subsiding potentially cooler air from aloft. Thermally forced convective rolls typically develop in the presence of vertical wind shear and align themselves with the low-level vertical shear vector (cf. Houze 1993, 167), which in this case (see Fig. 2) would be west–east. Lifting along the updraft intersection often results in the formation of cumulus bands or “cloud streets,” whose spacing reflects the cross-roll wavelength of the HCRs. The roll activity is typically most intense around the time of maximum solar heating of the surface (1430 LST herein) but quickly weakens thereafter and the rolls have usually vanished by sunset. Thus, HCRs are a daytime, essentially afternoon, phenomenon.

For the roll-only simulation, the model is fully 3D but lacks an ocean surface. The horizontal domain extent is 25 km × 25 km, with 500-m horizontal resolution in the north–south (cross roll) direction and a 1-km spacing in the east–west (along roll) direction. The lateral boundaries on the western and eastern sides of the domain are open, whereas the north–south boundary is periodic. To produce HCRs in a numerical model, high spatial resolution perpendicular to the roll orientation is required, with a grid spacing being 500 m or less (e.g., Balaji and Clark 1988). Simulations performed using lower cross-roll resolution (=1000 m) failed to generate recognizable roll structures. Balaji and Clark (1988) pointed out that enforcing periodicity in the cross-roll direction can “quantize” the number of rolls that develop, thereby affecting their horizontal cross-roll wavelength. This is not a concern herein as we desire to examine the dynamical impact of the rolls, irrespective of their specific wavelengths. The relatively coarser along-roll resolution does not appear to degrade the results and adding additional grid points in this direction had relatively little effect on the results. The model is run for 12 h (to 1800 LST), by which time the generated rolls have dissipated.

The initial convection that appears after sunrise is highly unstructured but the offshore mean flow slowly organizes it into east–west bands (HCRs), with the first suggestion of roll convection appearing around 1130 LST (not shown). The rolls are most active between 1400 and 1500 LST, around the time of the most intense surface heating. Figure 6 shows vertical y–z cross sections of vertical motion (w), cross-roll wind (v), and horizontal divergence taken across the rolls at the center of the x domain at 1400 LST. Five roll updrafts and five downdrafts may be seen, the axes of which extend normal to the page, nearly parallel to the ambient flow and vertical shear. Each updraft (downdraft) is collocated with low-level horizontal convergence (divergence) in the cross-roll flow.

Note that the vertical velocities weaken above the ≈2-km deep CBL until they have disappeared completely. The average cross-roll wavelength is ≈5 km, though
there is some variation in intensity and wavelength among them. The exchange of energy between simulated HCRs is known to cause some degree of lateral expansion over time (Balaji and Clark 1988). At the time shown, the HCRs possess relatively little along-roll variability (not shown) and are also reasonably stationary, the latter owing to the lack of mean flow in the cross-roll direction. The HCRs appear to meander mostly after 1600 LST when surface heating is declining rapidly.

In many respects, the simulated HCRs appear realistic. Their gross structure compares favorably to those simulated by Mason and Sykes (1982) except for the general absence of gravity wave activity occurring in the layer above the somewhat artificially deep CBL. Both the absence of wave activity and the excessive CBL depth may be partly due to the absence of an initial capping inversion in the model sounding. The ≈5-km average roll wavelength falls within the range typically found in observations. A review of observational work reveals typical HCR aspect ratios (i.e., the ratio of cross-roll wavelength to boundary layer depth) fall in the range from 2 to 4. Kuettner (1971) theoretically determined that the wavelength should be given by

\[ l = 2(2)^{1/2} z_i, \]

where \( z_i \) is the boundary layer depth, yielding an aspect ratio of about 2.8. At 1400 LST, the rolls’ aspect ratio is about 2.5. This ratio decreases slightly in the afternoon when the rolls expand more vertically than they do horizontally.

Observations (e.g., LeMone 1973; Chou and Ferguson 1991; Kristovich 1991) have shown that HCRs, along with smaller-scale turbulent motions, are very efficient in vertically redistributing inequities in heat, moisture, and momentum in the boundary layer, resulting in a well-mixed layer by afternoon. Recent observations by Weckwerth et al. (1996) confirm the critical importance of HCRs in the thermodynamic structure of the CBL. Since they mix the CBL in an organized manner they can establish horizontally aligned thermodynamic variations. Figure 7 shows perturbations from the horizontal means of potential temperature (\( \theta' \)), water vapor mixing ratio (\( q' \)), and along-roll horizontal velocity (\( u \)) for the same portion of the domain as in Fig. 6. Tropical observations from LeMone and Pennell (1976) noted temperature perturbations of about 0.1 K within one roll wavelength, whereas perturbations as high as 0.5 K (Reinking et al. 1981; Atlas et al. 1986; LeMone and Pennell 1976; Weckwerth et al. 1996) have been observed. There can be a great deal of vertical variation in this perturbation (Atlas et al. 1986). In Fig. 7a, perturbations in potential temperature across one roll wavelength are as high as 0.1 K in the CBL and even higher at the rolls’ tops, where the stable atmosphere is acting to suppress the boundary layer motions.

The roll updrafts (downdrafts) act to locally moisten (dry) the boundary layer. Observations of the moisture variability due to roll circulations suggest a range of 0.08–0.50 g m\(^{-3}\) (e.g., LeMone and Pennell 1976; Kristovich 1991), or roughly 0.1–0.5 g kg\(^{-1}\), within one roll wavelength, whereas values as high as 2.5 g kg\(^{-1}\) have also been observed (Weckwerth et al. 1996). Figure 7b shows that the moisture perturbations in the models’ HCRs fall within the observed range. Cross-roll wind speed variations of 0.4–1.4 m s\(^{-1}\) have been noted, and Fig. 7c shows perturbations of up to ±1 m s\(^{-1}\) in the simulated rolls. Momentum is primarily affected in the center of the roll circulations between 1.0 and 2.0 km. Although the strength of the rolls appears realistic, their updrafts generate little condensation. Certainly, no deep convection develops at any time.

The across-roll thermodynamic perturbations induced by the roll circulations cause variations in convective instability to develop, with parcels located in the roll updrafts being relatively more unstable than their counterparts in the downdrafts. Indeed, Weckwerth et al. (1996) noted that soundings must be launched from within roll updrafts to accurately assess the potential

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*Beyond this range, lower and upper limits appear to be 200 m (Ferrare et al. 1991) and 20 km (Asai 1966).*
instability. Figure 8 shows how the across-roll mean and standard deviation of parcel convective available potential energy (CAPE) varies with time in the RO simulation. Responding to surface heating, the mean CAPE increases swiftly after 1100 LST, though across-roll variations do not commence until the rolls become well established (after 1400 LST). Mean CAPE subsequently decreases, but remains relatively higher for parcels located on the roll updraft axes, due primarily to their higher moisture contents (Fig. 7b). Some residual CAPE variation remains even after the rolls disappear. The roll circulation’s effect on convective instability will be seen in section 5a to play a role in establishing along-frontal variations on the SBF.

4. Interaction between the SBF and HCRs

In this section, we examine a 3D simulation having both land and sea subdomains that permits both the sea-breeze circulation and horizontal convective rolls to form. Again, the mean flow is vertically sheared and directed offshore. The primary focus is on the temporal evolution of these features and the spatial variability of the SBF that is induced during the period in which both exist and interact. The model domain, combining characteristics of the SBO and RO simulations, has 141 grid points with 1-km resolution in the along-roll ($x$) direction and 25 grid points with 500-m resolution in the across-roll ($y$) direction, with boundary conditions as employed in the RO run.

a. Formation and evolution of the SBC and HCR circulations

Figure 9 presents horizontal cross sections of the vertical motion and horizontal divergence fields taken at

\[ \text{CAPE (J/kg)} \]

\[ \text{std dev(CAPE)} \]

Fig. 8. Time series showing across-roll mean (solid line) and standard deviation (dashed line) of convective available potential energy (CAPE) for the RO simulation. Note the highest variation in CAPE occurs well after the time of maximum mean CAPE.

The CAPE values were computed by lifting parcels originating at the first thermodynamic grid point above the surface and located along $x = 12.5$ km, the center of the $x$ domain.

the 1-km height level for a portion of the model domain at 1300, 1400, 1500, and 1700 LST. Note the subdomain depicted lies entirely over the land surface. Prior to 1300 LST, the 3D model’s SBC (not shown) is extremely similar to that seen in the 2D simulation presented in section 3a. Even at 1300 LST (Fig. 9a), little along-frontal variation in vertical motion is detectable, and the SBF lifting (with maximum intensity of $\approx 4$ m s$^{-1}$) has established a nearly homogeneous band of shallow cloudiness along the front (not shown). Unlike the 2D simulation, the 3D SBF at this time is propagating across a land surface on which heating perturbations have been applied, but these appear to be too small to have had much effect on along-frontal structure or the frontal propagation speed.

As in the RO simulation, roll convection is developing over the land at this time, but is not yet strong enough to disturb the SBF. Although the timing of roll development remains very similar to that seen in the RO run as the simulation progresses, the spatial distribution and intensity of the rolls differ somewhat. By 1400 LST in the RO simulation, the HCRs were well developed and this also occurred in the present run (Fig. 9b). Again, five rolls of unequal intensity were generated. However, roll convection over land is clearly suppressed in the immediate vicinity of the marine air ($80 \leq x \leq 95$ km), which represents a transition zone between the SBC and HCR environments. In contrast, the roll updrafts (downdrafts) developing farther inland ($x > 95$ km) are slightly stronger (weaker) than their counterparts in the RO run. This is attributed to the SBC’s cross-shore mesoscale circulation in which general upward motion is induced over the land surface (aiding the updrafts and weakening the downdrafts) owing to the stronger heating occurring there than over the sea. This could not occur in the RO simulation, which lacked a sea subdomain.

By 1500 LST (Fig. 9c), just prior to the time of maximum surface heating, the transition zone between the two environments has disappeared and the SBF is encountering the now well-developed and intense HCRs. Slight bending of the HCR updraft axes can be noted in the immediate vicinity of the SBF, though, unlike the situation in AWW, the deflection is neither substantial nor systematic near the front. In contrast to earlier times, there is now a clear variability in $w$ along the SBF, which reflects the wavelength and spatial positioning of the HCRs. This can be more easily detected in the right-hand panels of Fig. 9, which focus narrowly on the vicinity of the SBF. At this height level, upward motion exists all along the SBF, though the lifting is clearly stronger (weaker) where the roll updrafts (downdrafts) have intersected the front. This will be discussed in depth in section 5. Lifting in both zones has reached over 3.0 m s$^{-1}$ at the height level shown, but the roll updrafts are not producing lifting as intense as that associated with the SBF. At 1600 LST (not shown), the situation closely resembles that of Fig. 9c.
As in the RO simulation, the HCRs weaken rapidly as the solar heating diminishes. By 1700 LST (Fig. 9d), the rolls have largely vanished and so too has the previously identified variability along the SBF. By 1800 LST (not shown), roll convection has disappeared and the SBF, itself weakening as the cross-shore density difference declines, has once again become an essentially 2D entity. Therefore, in the present simulation, along-frontal variability in the SBF existed solely due to the presence of HCRs.6

b. Vorticity and horizontal motion

The panels of Fig. 10 show the three vorticity components, the horizontal vorticities about the x axis ($\eta_x$) and y axis ($\eta_y$) and the vertical vorticity ($\zeta$), respectively, at 1500 LST for the same subdomain and height level depicted in Fig. 9. The stippled region indicates vertical motion in excess of 3.0 m s$^{-1}$ and marks the SBF. The roll circulations are clearly identifiable in the $\eta_x$ component (Fig. 10a); note that at this time they extend up to, but not beyond, the SBF, which obliterates the rolls as they come in contact with it. The $\eta_y$ component is being generated by baroclinic horizontal (eastwest) density gradients across the SBF and within the marine air. These gradients are concentrated in two locations: at $x = 92$ km, immediately behind the SBF, and also farther seaward, at $x = 80$ km, beyond which location relatively unmodified marine air is present.

The SBF lifts the rolls it encounters, resulting in the transformation of the rolls’ original vorticity ($\eta_z$) into vertical vorticity, as shown in Fig. 10c. At the updraft intersection points, where the SBF lifting has been locally enhanced, vertical vorticity couplets are seen bisecting the vertical velocity maxima, as expected. The horizontal circulations associated with these $\zeta$ couplets introduce a small but systematic along-frontal variation in the SBF’s kinematic position; this will be more closely examined presently. Vorticity couplets can also be seen immediately behind the SBF, in the range 78 km $\leq x \leq 91$ km, where the rolls descend in the subsidence region seaward of the front. This is indicated by the fact that these (albeit weaker and more elongated) couplets have the opposite sign of their counterparts at the SBF’s leading edge, which is considered proof that the rolls are physically lifted by the SBF even though the rolls are considerably weakened in the process. Indeed, it was noted earlier that the roll circulations vanish soon after passing into the marine air; this is indicated yet again by the lack of horizontal structure in the $\zeta$ field in the marine air ($x < 78$ km).

Figure 11 shows the $u$ field (flow orthogonal to the coastline) with an overlay of horizontal airflow vectors at 1500 LST for a still smaller subdomain, taken at the 50-, 1000-, and 1500-m height levels representing the surface layer (lowest thermodynamic level) and the upper portions of the frontal and inland boundary layers, respectively. The frontal area is again marked by stippling. The near-surface flow is directed offshore (toward the SBF) ahead of the front and inland behind it. Convergence and divergence signatures owing to the rolls’ circulations may be seen in the offshore flow, which begins weakening as far as 5–10 km out ahead of the front as it becomes subjected to lifting by the SBC’s mesoscale circulation. The marine air behind the SBF is fairly undisturbed. Although the maximum horizontal temperature gradient is located at the 50-m level, the strongest onshore flow immediately behind the SBF at this time is found at 400 m above the ground (not shown). As noted previously in section 3a, this is caused by surface friction, and the height level of this local maximum rose during the afternoon.

At 1000 m, the $u$ contours in the immediate vicinity of the SBF possess marked along-frontal undulations, the result of the horizontal circulations established by the HCRs as they are tilted vertically over the SBF. Positive perturbations, representing increased marine inflow and/or suppressed offshore flow, are located at the updraft axes, whereas negative perturbations coincide with the downdraft axes. The thermodynamic position of the SBF drawn into Fig. 11b illustrates that there is along-frontal thermodynamic variability, which closely mimics the modulation seen in the $u$ field. The SBF and its frontal updraft tilt westward (seaward) with height, as do the $u$ perturbations, which are most pronounced at 1000 and 1500 m (Figs. 11b and 11c), above the friction layer. In Fig. 11a, marine flow near the surface (kinematically defined at the zero contour of $u$) has reached $x = 94$ km (59 km from the coast), 2 km farther inland than the frontal position at 1000 m.

c. Cloud field

Figure 12 shows the three-dimensional 2 g kg$^{-1}$ iso-surface of the cloud water mixing ratio at 1420 LST, which falls between the times shown in Figs. 9c and 9d. The surface positions of the SBF and HCR updraft axes are also depicted. There are three distinct regions of cloudiness, representing HCR, frontal, and post-frontal cumuli, respectively. Significant cloudiness overlies only two of the five HCRs (those located at $y = 6.5$ and 16.5 km), those having the strongest roll updrafts. In contrast, five distinct areas of enhanced cumuli can be detected along the SBF. These cumuli occur at the intersection points of the HCR updrafts with the SBF, and thus even the weaker rolls have helped to locally intensify the SBF convection. Even more striking is the absence of appreciable convective activity at the locations where the HCR downdrafts intersect the SBF (located in between the indicated HCR updraft axes). Recall that prior to the development of the HCRs, as well

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6 Note again that in order to isolate and highlight this phenomenon, it was deemed prudent to remove other potential sources of along-frontal variability, as discussed in section 2b.
as after their demise, cloudiness along the SBF is nearly uniform. Therefore, HCRs have clearly acted to modulate convection along the SBF. The postfrontal cumuli shown represent clouds that formed over the SBF prior to the time shown and subsequently propagated seaward (rearward of the SBF) in the rising offshore airflow.

The cloud field is shown at 1420 rather than at 1500 LST because by the latter time one of the frontal cumuli (located where the strongest HCR updraft intersected the SBF) grows into a deep thunderstorm. The impact of this thunderstorm on the simulated circulation is limited owing to its relatively short lifetime and the fact that precipitation has been deactivated, as explained in section 2b. Yet, it is important to point out that deep cumuli do not develop in either the SBO or RO benchmark simulations. Thus, the presence of the sea-breeze front and its interaction with HCRs together provide the lifting sufficient to initiate deep convection.
Fig. 10. Horizontal cross sections of vorticity (5 × 10⁻⁴ s⁻¹ contours) parallel to the (a) x, (b) y, and (c) z axes. Cross sections are taken at a height of 1 km at 1500 LST. The stippled region indicates upward motion in excess of 3 m s⁻¹.

5. Discussion

a. Spatial modulation of convection related to the SBF–HCR interaction

Figures 9 and 12 clearly reveal a periodic modulation of vertical motion and cloudiness along the SBF, which stems directly from the merger of HCR updrafts and downdrafts with the SBF. The spatial variation, which persists only as long as the HCRs exist and are in contact with the front, has a wavelength that precisely matches that of the HCRs. To facilitate closer examination, Figs. 13 and 14 present vertical cross sections of vertical motion and cloudiness, taken at locations identified in Fig. 12. Figure 13 shows x–z cross sections through two roll updrafts and one roll downdraft at 1420 LST. The first panel is aligned along y = 6.0 km and passes through the strongest roll updraft and the cloud it produces, whereas the other two panels are located at y = 8.6 and 11.2 km and capture a roll downdraft and a weaker, cloud-free roll updraft, respectively. Also, the effect of the vertical tilting of the rolls as they approach the front can be seen in the figure’s vertical velocity fields. In Fig. 14, the y–z cross sections shown pass through the rolls located well ahead of the SBF (at x = 115 km; Fig. 14a) and along the SBF (at x = 85 km; Fig. 14b).

Figure 14a again shows that the intensity of the roll updrafts varies (by ~1.5 m s⁻¹) and that only two of

Recall from section 4a that the roll axes tend to turn slightly in the immediate vicinity of the SBF, out of the planes shown.
the five rolls are strong enough to produce significant cloudiness at this time. Where they intersect the SBF, the roll updrafts appear to augment the frontal lifting whether or not the rolls themselves are strong enough to generate a cloud. The average maximum lifting along the SBF is ≈4.4 m s$^{-1}$, but it is somewhat enhanced (to >5 m s$^{-1}$) at the roll updraft intersection points (Figs. 13a and 13c) and suppressed (≈4 m s$^{-1}$) at the downdraft intersection (Fig. 13b). As expected, the strongest frontal cumuli (in terms of liquid water content) and most intense SBF lifting occur where the strongest roll has intersected the front (cf. Figs. 13a and 13c). However, though stronger, the cloud developing above the front is not deeper than that above the strongest roll.
updraft (Fig. 13a), at least at this time. As noted earlier, the frontal cumulus shown in Fig. 13a develops into a deep thunderstorm shortly after the time shown. The combination of forcings associated with the strongest roll and the SBF have accomplished what neither could do independently.

Especially striking to us is the absence of cloudiness above the SBF at the downdraft intersection points. At the location shown in Fig. 13b, the SBF frontal updraft is capped by subsidence associated with the roll downdraft it is acting to lift above the front. Parcels being lifted above the SBF at that point are barely able to reach their lifting condensation level, resulting in a very small patch of saturated air above the frontal boundary. Thus, the spatial variability of convection along the SBF appears to be as much due to the influence of roll downdraft suppression as roll updraft enhancement. Recall that before the HCRs develop and after they decay, the SBF is able to generate nearly uniform, if shallow, cloudiness along the front.

We can identify at least two (somewhat complementary) explanations for the origin of the enhanced (suppressed) convection at the roll updraft (downdraft) intersection points. First, the extra lifting of the roll updrafts contribute to the SBF could perhaps be sufficient to encourage stronger convection at the updraft intersection locations; this represents a dynamic effect. Second, the rolls’ very presence has acted to alter the SBF’s upstream environment, introducing thermodynamic variability in the cross-roll (i.e., along-frontal) direction, as was demonstrated for the roll-only simulation (section 3b; Figs. 7 and 8). As trajectory analyses (not shown) confirm that parcels reaching the SBF updraft and downdraft intersection points tend to move along the roll axes,9 the air feeding into the updraft (downdraft) intersection locations has relatively greater (lesser) convective instability in addition to stronger (weaker) lifting.

The enhanced frontal cumuli located above the updraft intersection points are positively buoyant. Thus, whatever their origin, once the enhanced frontal cumuli are established they generate local circulations that act to produce subsidence on their flanks (e.g., Fovell and Tan 1998), that is, above the roll downdraft intersection points. That is, once variability along the front is significant, there is a positive convective feedback owing to cloud-scale circulations that serves to maintain that

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9 These parcels experience little along-frontal motion because v tends to vanish along the roll axes. See also Fig. 6b.
variability, at least as long as the rolls retain appreciable strength. This positive feedback would not exist in the absence of roll convection.

A simple schematic of the entire process is presented in Fig. 15. In Fig. 15a, the frontal lifting and the cloud band it produces are homogeneous prior to the development of the HCRs. This represents frontal convection prior to 1300 LST in this simulation and is qualitatively similar to the two-dimensional SBO benchmark detailed earlier. By the time of Fig. 15b, the HCRs have appeared and are already affecting the atmosphere in the roll environment, increasing the convective instability of parcels residing along the roll updraft axes relative to those at the downdraft axes. Finally, in Fig. 15c, the SBF has
reached the HCRs and has begun to evince substantial along-frontal variation, with enhanced (suppressed) cloudiness occurring at the roll updraft (downdraft) intersection points. The circulations generated by the positively buoyant, enhanced frontal cumuli further suppress convection at the downdraft intersection points. Such suppression continues until the rolls weaken as the CAPE in the roll environment declines. After this time, the SBF returns to the state shown in Fig. 15a, being capped with shallow, nearly homogeneous cloudiness.

b. Kinematic and thermodynamic considerations

Beyond the more obvious interaction related to vertical motion, vorticity associated with the HCRs interacts with the sea-breeze front. Vorticity parallel to the $x$ axis ($\eta_x$) is tilted in the frontal zone, resulting in a conversion of vertical to horizontal motion and, in turn, variability in $u$ along the sea-breeze front. A schematic representation of this effect is shown in Fig. 16, while Fig. 17 shows a horizontal slice of the vertical vorticity.
Fig. 15. Schematic representation of the interaction between the sea-breeze front (SBF) and horizontal convective rolls (HCRs). The SBF is propagating inland, from left to right, into the environment in which HCRs are developing. See text for further details.

Field taken at a height of 1 km with an overlay of the $u$ field. In the latter, roll updraft axes are depicted as solid lines.

It is clear that there is an alternation in the vertical vorticity field along the front as HCR updrafts and downdrafts approximately bisect regions of positive and negative vorticity. Wilson et al. (1992) showed that similar rotations occurring along a strong convergence line can develop into weak tornadoes. Here, however, the rotation is not very strong. Also, we can see that along the line of HCR updrafts there is an increase in westerly (onshore) flow and a corresponding reduction in onshore flow along roll downdrafts. Because westerly flow brings with it cool, moist marine air, vorticity tilting of HCRs plays a role in both kinematic and thermodynamic along-frontal variation. (This correlation can be seen in
the $w$ contours and thermodynamic position of the front in Fig. 11b.) In this simulation, periodic temperature perturbations occurring along the front have a magnitude of up to 0.6 K. Presumably, the stronger the front, the more lifting and tilting it will produce, and the greater the potential thermodynamic modulation would be.

6. Summary and future work

A 3D cloud model has been used to investigate the mutual interaction of the sea-breeze front (SBF) and horizontal convective rolls (HCRs). Our primary objective was to comment on the spatial and temporal evolution of the sea-breeze front as it interacts with HCRs and to detail the impact of the interaction on vertical motion and the resulting cloud field. The study was restricted to cases in which the mean flow was initially directed offshore, with unidirectional shear confined to the lowest kilometer above ground level. The initial thermodynamic profiles reflected mid-August conditions in Florida.

To facilitate this study, we performed two benchmark simulations to verify the formation of the sea-breeze circulation and roll convection independently. The sea-breeze benchmark (SBO) was run in 2D. The circulation developed quickly, but onshore flow did not appear for several hours due to the impinging mean flow. Once the front formed and moved inland, it propagated at a speed consistent with prior studies. Maximum vertical motion was seen at the forward flank of the circulation above the sea-breeze front and shallow clouds appeared in the afternoon along the frontal boundary. The HCR benchmark (RO) was run in 3D but without a land–sea boundary. Motion in the $y$ direction was excited by random perturbations ($\pm 5\%$) applied to the surface heating. HCRs, aligned with the mean flow (east–west), developed in the afternoon during intense surface heating and
were rather stationary and homogeneous. The HCRs had a wavelength and aspect ratio consistent with observation and theory. Perturbations resulting from roll-induced boundary layer mixing were in good agreement with observations as well, and these perturbations resulted in relatively enhanced (suppressed) convective instability along the updraft (downdraft) axes. The HCRs decayed quickly as surface heating waned late in the afternoon.

Having simulated the daytime evolution of the SBF and HCRs independently, we discussed in depth a 3D simulation containing a land–sea boundary. The domain was made sufficiently large to accommodate the formation and evolution of both the sea-breeze circulation and roll convection, allowing each to develop independently but eventually interact. Early in the simulation, the sea breeze looked much like its 2D counterpart in run SBO. There was little along-frontal variability as the front first pushed inland despite the application of random perturbations. Although roll development was suppressed near the coastline, farther inland the HCRs appeared much as they did in the land-only run RO.

However, once the SBF and HCRs met, the situation was quite different. It was first noted that the SBF–HCR interaction did not take place suddenly. Eventually, however, distinct perturbations occurred along the frontal boundary having a wavelength consistent with that of the well-developed HCRs. Vertical motion and convection was strongest where the roll updrafts intersected the front and weakest at the roll downdraft intersection points. This could have occurred due to the dynamical effect (including locally enhanced lifting and roll updraft tilting) and/or thermodynamical effect (relatively enhanced convective instability) of the roll updrafts. It was noted that compensating downdrafts located on either side of the deep cumulus clouds (located above HCR updrafts) acted to further suppress shallow cloudiness (located above HCR downdrafts). Thus, modulation of both vertical motion along the SBF and the size and depth of the resulting along-frontal cloud field could be attributed to the presence of roll convection in the frontal environment.

Further analysis of the offshore simulation revealed the effect of the HCRs on the vorticity field along the SBF. Vertical vorticity associated with the tilting of HCRs could be seen in the lifting zone ahead of the front. Horizontal motions along the front were modulated by the roll vorticity effect as well. It was hypothesized that the enhancement of onshore flow induced along a roll updraft could be associated with a local influx of marine air along the updraft axis. A corresponding perturbation occurred at the downdraft axis.

Nearly 100 years of sea-breeze investigation has made it clear that the marine circulation is not strictly a 2D phenomenon. With the exception of laboratory tests, inhomogeneities occurring along the front have, however, been largely ignored in the literature. Wakimoto and Atkins (1994) and Atkins et al. (1995) have provided some of the first observational evidence of along-frontal modulation resulting from the presence of roll convection. Here, we have attempted to verify the ability of HCRs to produce spatial variation in a sea-breeze front. Further, we have attempted to provide additional details relevant to the SBF–HCR interaction. In Part II of this study, we will perform a similar analysis as it relates to a simulation initialized with mean flow directed along (rather than across) the coastline. This singular change in the model setup produces a vastly different frontal evolution.

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APPENDIX

Model Enhancements

A planetary boundary layer parameterization was added to the Klemm–Wilhelmson (1978) cloud model, largely following the implementation detailed in Anthes et al. (1987). In this approach, the land surface temperature, \( T_s \), is predicted via a surface energy budget equation based on Blackadar’s (1976) “force–restore” slab model (see also Zhang and Anthes 1982) and given by

\[
C_s \frac{dT_s}{dt} = R_s - R_l - H_s - H_s - L_v, \tag{A1}
\]

where \( C_s \) is the thermal capacity of the slab per unit area, \( R_s \) is the net solar (shortwave) radiation, \( R_l \) is net terrestrial (longwave) radiation, \( H_s \) is the heat flow into the substrate, \( H_s \) is the sensible heat flux into the atmosphere, \( L_v \) is the latent heat of vaporization, and \( E_s \) is the surface moisture flux. The thermal capacity of the slab \( C_s \) varies by ground surface type, and the land surface was taken to have urban characteristics [see Anthes et al.’s (1987) land use table]. More details regarding the implementation of this parameterization may be found in Dailey (1996).

The parameterization of the spectrally integrated solar radiation \( R_s \) is based on Bird and Hulstrom (1980), a model found to be superior by Iqbal (1983), particularly with regard to its the handling of diffuse radiation (radiation from the atmosphere into the
ground). The effects of clouds on radiative transfer were incorporated using Benjamin’s (1983) three-category approach, which divides clouds into low, medium, and high clouds, following the Anthes et al. (1987) implementation of this scheme. The substrate heat flow term \( H_s \) is handled as described in Anthes et al. (1987), whereas the surface sensible heat and moisture flux terms \( H_e \) and \( E \) were computed using Dardoroff’s (1972) bulk aerodynamic model, which related the fluxes to the vertical gradients between the ground surface and the lowest model grid point. A higher resolution PBL [similar to the multilevel scheme in Anthes et al. (1987)] was tested and abandoned because it tended to artificially duplicate the mixing action of HCRs and thus impeded their development.

At model start, the ground, substrate, and near-surface air temperatures are set equal. This is necessary in order to obtain a stable solution (Carlson and Boland 1978). Based on the net heat flow from all of the above constituents into or out of the surface, the time derivative of ground temperature is calculated. Over land, as the zenith angle of the sun changes over the course of a simulation, the primary thermal forcing (solar radiation) results in a diurnal variation in ground temperature and moisture. Over the ocean, the sea surface temperature is held constant, a reasonable assumption given the ocean’s tremendous heat capacity.

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