Combined Wind Profiler/Polarimetric Radar Studies of the Vertical Motion and Microphysical Characteristics of Tropical Sea-Breeze Thunderstorms

PETER T. MAY
Bureau of Meteorology Research Centre, Melbourne, Victoria, Australia

A. R. JAMESON
RJH Scientific Inc., El Cajon, California

THOMAS D. KEENAN
Bureau of Meteorology Research Centre, Melbourne, Victoria, Australia

PAUL E. JOHNSTON
CIRES, University of Colorado, Boulder, Colorado

CHRIS LUCAS
Department of Physics and Mathematical Physics, Adelaide University, Adelaide, South Australia, Australia

(Manuscript received 7 June 2001, in final form 16 January 2002)

ABSTRACT

An experiment combining wind profiler and polarimetric radar analyses of intense, but shallow, tropical thunderstorms has been performed. These storms are important as they are very common along many tropical coasts and islands and are sometimes the precursors to large intense multicellular storms such as occur over the Tiwi Islands north of Darwin, Australia. All the storms sampled had a similar structure, with intense updrafts on the periphery of the cells producing significant-sized hail and downdrafts in the storm center. The hail concentrations are relatively small, but have a large effect on the radar reflectivity and polarimetric measurands because of the size (10–20 mm). It is this hail melting that produces characteristic $Z_{DR}$ columns in the polarimetric radar data.

1. Introduction

An experiment combining polarimetric radar and multiple-frequency wind profiler observations was performed at a site near Darwin, Northern Australia, during November and December 1997. The experiment was designed to study storm vertical motion and microphysical characteristics as well as quantitative precipitation measurements with polarimetric and conventional weather radar techniques using the Bureau of Meteorology Research Centre (BMRC) C-band (5-cm wavelength) dual-polarization weather radar (C-Pol: Keenan et al. 1998). In the early part of the experiment, several short-lived storms that formed on locally forced convergence lines passed over the profilers. This paper focuses on the vertical circulation and precipitation microphysics of these storms that were observed primarily with the wind profilers and C-Pol radar. Results concerning rain estimation, hail detection, and microphysical effects on the polarimetric measurements are given in May et al. (2001).

These shallow storms are interesting and important in many respects. Although they are generally short lived, they can be intense, with strong updrafts and local heavy rain associated with reflectivities in excess of 50 dBZ, but only extend about 3 km above the freezing level (FZL). They are widespread across the Tropics and are the precursor convection that triggers deep island-based convection over much of the Maritime Continent in the lead-up to the monsoon (Wilson et al. 1997). Although these storms are relatively small and short lived, there are large numbers of them along the northern Australia coast and the islands nearby. More than 40% of the cases examined by May and Rajopadhyaya (1999)
were these shallow storms and, on average, they contributed significantly to the total low-level mass flux associated with convection in the Darwin area. Similar storms observed in Florida and elsewhere have been shown to contain large quantities of ice including hail with diameters in excess of 1 cm despite extending only a short way above the FZL (e.g., Koenig 1963; Braham 1964; Jameson et al. 1996).

Although profilers were developed to provide long-term measurements of the mean horizontal wind, they are very useful tools for high temporal resolution measurements of vertical motion (May and Rajopadhyaya 1999; Cifelli and Rutledge 1994, 1998) and precipitation characteristics (e.g., Fukao et al. 1985; Larsen and Rottger 1987; Wakasugi et al. 1986; Gossard 1988; Currier et al. 1992; Chilson et al. 1993; May and Rajopadhyaya 1996). This is a result of their ability to simultaneously observe radar signals from the clear air and precipitation with high time resolution. The focus here is on 1-min-resolution vertical velocity measurements with a 50-MHz system and a collocated 920-MHz system to provide reflectivity and reflectivity-weighted fall speed information.

The profilers are located approximately 23 km south of the C-Pol radar (Fig. 1). The scanning strategy for C-Pol included a sequence of a volume scan (for other research requirements) followed by an RHI scan and a 2-min fixed-azimuth antenna position scan directed over the wind profiler site. This strategy allows the detection of the organization of the storms, boundaries associated with their initiation and their three-dimensional structure, and high spatial and temporal resolution observations over the profiler for comparison purposes.

### 2. Analysis techniques

This section will very briefly discuss the basic data analysis techniques with a focus on their potential problems and limitations. May et al. (2001) also includes a much more detailed description of the analysis techniques.

#### a. Profilers

There are two wind profiler radars located at a site near Darwin airport, one operating at 50 MHz and the other at 920 MHz. The operating parameters of these systems are given in Table 1.

The 50-MHz profiler is used to determine the vertical air motion, both as an end in itself and to provide a reference for precipitation retrievals using 920-MHz data. The accuracy of these vertical motion measurements is dependent on the spectral width of the radar echoes. The width is usually such that the uncertainty in individual estimates is ~0.1 m s⁻¹. However, when intense convection is overhead, the spectral width may reach values of around 2 m s⁻¹ and the theoretical accuracy of the vertical motions is ~0.2–0.3 m s⁻¹ (e.g., May and Strauch 1989). However, visual inspection of some spectra indicates that the uncertainty may be as much as 0.5 m s⁻¹ for some of the wider spectra. Another issue is potential contamination by precipitation echoes. Our experience with these data is that the fall speeds of rain and hail echoes provide ample spectral separation between the precipitation and clear air peaks. Problems may occur when there is snow or if a brightband is evident. With a brightband, the precipitation peaks are very broad and intense. However, with the highly convective nature of the storms under discussion this was not a problem. Snow was a potential problem, but appears to be confined to heights well above the FZL in these data. Dry snow also tends to have relatively weak reflectivity, but there may be some small biases (~0.5 m s⁻¹).

The 920-MHz profiler is used as a vertically pointing Doppler weather radar to provide reflectivity estimates and to provide the fall speed spectra to derive drop and

<table>
<thead>
<tr>
<th>Parameter</th>
<th>50-MHz profiler</th>
<th>920-MHz profiler</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scan sequence</td>
<td>V(45s)E(15s)V(45s)N(15s)</td>
<td>V(45s)E(15s)V(45s)N(15s)</td>
</tr>
<tr>
<td>Height sampling (m)</td>
<td>315</td>
<td>105</td>
</tr>
<tr>
<td>Height resolution (m)</td>
<td>450</td>
<td>150</td>
</tr>
<tr>
<td>Height coverage (km)</td>
<td>1.5–20 km</td>
<td>200 m–12 km</td>
</tr>
<tr>
<td>Beamwidth (°)</td>
<td>3</td>
<td>9</td>
</tr>
</tbody>
</table>

---

**Fig. 1.** Map showing the relative locations of the C-Pol radar and the profiler site.
hail size distributions using the following relation between fall speed and diameter for rain:

\[ w = (9.65 - 10.3e^{-0.6D})(\rho/\rho_0)^{-0.45}, \]

(1)

where \( w \) is the fall speed in still air, \( D \) is the diameter (in mm), \( \rho \) is the air density, and \( \rho_0 \) is the air density at mean sea level pressure (Atlas et al. 1973). Note that the fall speed relation gives an asymptotic limit for the maximum rain fall speed at mean sea level pressure of 9.65 m s\(^{-1}\) and for hail it is

\[ w = 1.426D^{0.8}(\rho/\rho_0)^{-0.45}, \]

(2)
adapted from Pruppacher and Klett (1978).

This dataset is one of the first to include hail echoes in profiler data. How can we be confident that what we are seeing is hail, despite having no in situ data? An example of the 50- and 920-MHz Doppler spectra (Fig. 2) shows two clear precipitation peaks, one at fall speeds beyond the asymptotic limit for rain. Using Eq. (2), we estimate hail sizes of 1–2 cm. This is one of the clearest examples, but data from all the storms examined as part of this study showed clear evidence of hail at some time and such bimodal spectra were common through the dataset that will be discussed in more detail. This is independently supported by analysis of polarimetric radar data (see May et al. 2001 for details). At the heights where the microphysical retrievals are performed, below 4 km (~1 km below the FZL), it is assumed that the echoes at fall speeds less than the rain asymptotic limit are dominated by rain; that is, only the largest hail has not melted. If the reflectivity associated with hail fall speeds is less than 0.3 of the total reflectivity, the hail content is set to zero.

Note that prior to the calculation of particle size distributions, the spectral broadening associated with turbulence, the use of radars with finite beamwidths, etc. must be removed. We use a deconvolution approach using the clear air peak as an estimator for the broadening of the precipitation peak (e.g., Gossard 1988; Rajopadhyaya et al. 1993). The main uncertainties in these retrievals arise from the effects of incomplete deconvolution. This will leave the spectra somewhat broader than a “pure” fall speed spectrum and result in overestimates of small and large drops. The small drops are of most concern as large numbers may be needed to explain the observed spectral power (because of the \( D^6 \) dependence) and cause overestimates in water content. In practice this is minimized because of our cutoff at small drop diameters. The drop size distributions (DSDs) retrieved by profilers have shown good agreement with in situ measurements (Rogers et al. 1993) and gauge data (e.g., Rajopadhyaya et al. 1998). May et al. (2001) showed good agreement between the polarimetric radar measurements of DSD (e.g., differential reflectivity, \( Z_{DR} \)) and the profiler estimates of \( D_0 \) in rain and rain rate using these data. The other great uncertainty in the rainwater volume is probably associated with calibration of the echo intensity. We have cross-calibrated the profiler with the C-Pol radar and estimate a 1 dBZ uncertainty (May et al. 2001). Overall, our estimates of the uncertainties in the water contents associated with rain and hail are ~30%.

In this paper we use the \( D_{0w} \), the diameter such that \( 1/2 \) of the water volume lies at diameters larger than \( D_{0w} \), as a descriptor of the full DSD for both rain and hail distributions. Hail and water totals are estimated by integrating the size distributions with respect to diameter. Particles less than 10 mm for hail and 0.7 mm for rain are neglected; so that these estimates are biased. However, for pure rain at least, these biases are only about 0.1 mm for \( D_{0w} \) (May et al. 2001).

b. Polarimetric radar

Applications with polarimetric weather radar have developed rapidly over the last two decades for quantitative rainfall estimation and precipitation microphysical classifications (see Zrnic and Ryzkov 1999 for a review and references within). The C-Pol radar measures the reflectivity factor, differential reflectivity \( Z_{DR} \), propagation differential phase \( \Phi_{dp} \), and the cross correlation of the horizontal and vertical polarization signals, \( \rho_{hv}(0) \), as well as velocity. The range derivative of \( \Phi_{dp} \), the specific differential phase \( (K_{dp}) \), is then calculated using an algorithm similar to that described by Jameson and Caylor (1994) to estimate the slope with backscatter effects removed. The \( K_{dp} \) turns out to be almost linearly related to rain rate and is insensitive to tumbling hail. However, large values of \( K_{dp} \) can occur when 1–2-cm-diameter hail is melting and a ring of water builds up around the hail core (Chong and Chen 1974). In rain the \( Z_{DR} \) provides an estimate of mean
drop shape (Jameson 1983) and, hence, diameter (as large drops become more oblate), but enhanced values also occur when hail particles begin melting and have a surface coating of water (Vivekandanan et al. 1990; May et al. 2001). In such cases when there is substantial mixed phase precipitation present or the scatterers are large ($\geq 5$ mm), the values of $p_{\text{prv}}(0)$ are suppressed from around 0.98 down to $\sim 0.92$. We will also be using the Doppler information from the polarimetric radar in the following analysis.

3. Storm observations

Several cases of isolated relatively shallow thunderstorms were observed over the profilers during a period of 3 weeks, but they displayed many common characteristics in their temporal development as well as their vertical motion and reflectivity structure. However, in two cases the subsequent storm development reached above 10 km. In both of these cases, there was another storm nearby. The storms were either initiated at mid-afternoon local time forming on the sea-breeze front after it had penetrated $\sim 10\text{–}20$ km inland or early in the morning just offshore on the land breeze. The storms tended to be isolated cells, typically developed rapidly, and had a total lifetime of about an hour. These occur in the region almost every day in the period prior to the monsoon onset and during “breaks” in the monsoon. A typical satellite image shows the penetration of the sea breeze with clear skies behind it and widespread cumulus congestus (Fig. 3). Two more intense cells were just forming, one immediately to the east of Darwin. As the storms develop they were advected to the west by the low-level easterly winds.

Characteristics of the soundings taken a few hours prior to the storms (the storm on 13 December occurred near the time of the sounding) are shown in Table 2. All except one have similar moderate values of $\text{CAPE}^1$ and values of about 600 J kg$^{-1}$ below 500 hPa (the latter is relevant to the vertical motions observed near the FZL, $\text{CAPE} \sim \frac{1}{2} \rho w_{\text{max}}^2$, where $\rho$ is the air density and $w_{\text{max}}$ is the maximum attainable vertical velocity). The other case seems relatively dry in the low levels but this may be an indication that the sounding passed through a region local of subsidence (cf. Weckwerth 2000). The delay in storms forming on the sea breeze until the afternoon may also be related to the time for the boundary layer to reach a sufficient depth and be sufficiently unstable due to the sensible and latent heating at the surface. The early morning storms had a similar structure as the afternoon storms. The following discussion will focus on the evolution and structure of one of the cases (9 December 1997). Although the later stage of development of this particular storm is less common, this case was chosen because it passed directly over the profiler and, at that time, had a structure fairly typical of such storms.

Figure 4 shows a series of PPI scans of reflectivity from the C-Pol radar along with automatic weather station (AWS) measurements of wind and temperature. The sea-breeze convergence line (SB) is visible passing near the profiler and had onshore wind components and temperatures $\sim 2\text{–}3^\circ$C cooler than over the mainland. At 0300 UTC a cell C1 is dissipating but the gust front (G) from it is colliding with the sea breeze along a line over the profiler. At 0312 UTC a new cell is just developing near the profiler and another to the south of the profiler. By 0325 UTC new convection has developed along the colliding boundaries including the cell over the profiler (P1) and another to the south of the profiler (C2) (cf. Wilson and Schreiber 1986). Cell P1 moved southwestward directly over the profiler at about 4 m $\text{s}^{-1}$. Cell P1 had an area of approximately 20 km$^2$ at this time, but only 12 min before was only about 2 km$^2$ in size and by 12 min after was merging with C2 and had an area $\sim 60$ km$^2$. This merger process was possibly important in continuing the storm growth (e.g., Simpson et al. 1993), although the other case studied that had echo tops above 10 km had two cells nearby, but no merger. The majority of storms normally dissipate when they reach a similar stage of development as the cell had reached at 0325 UTC.

RHI scans over the profiler show cell P1 just forming at 0309 UTC, but reaching a height of about 7 km by 0316 UTC (Fig. 5). The cell, C2 to the south is also visible. The 0309 UTC RHI shows the strongest echoes near 3–4-km altitudes indicating that warm rain processes were initially important, while by 0316 UTC there is evidence of frozen drops and hail in both the profiler and polarimetric radar data (discussed in detail later). The radial velocity shows low-level convergence and upper-level divergence in P1, while the low-level flow at C2 was directed toward P1. This is in evidence with strong flow away from the radar near the surface at a range of 23 km and flow toward at a range of 25 km. The upper-level flow shows the reverse pattern particularly strongly at about 4.5 km. There is a region of intense motion toward the radar in the middle of C1 indicative of the presence of a strong downdraft forming. The following RHI at 0334 UTC is immediately after the merger of C2 and P1 with the echo tops reaching to almost 15 km. There is now a massive upper-level region of divergence and while the low-level flow has some structure, any convergence is poorly organized. The maximum reflectivities had decreased and the area of large $Z_{\text{DR}}$ had decreased, consistent with the storm weakening. By 0346 UTC, P1 is dissipating with low-level divergence indicating strong outflows and the upper-level flow just shearing across the storm top (not shown).

Figure 6 shows the profiler time–height cross sections of vertical motion ($w$), reflectivity, and microphysical characteristics as the cell passed over. Some of the fea-

---

1 The $\text{CAPE}$ is calculated using a parcel that has been well mixed over the lowest 50 hPa of the sounding.
Fig. 3. GMS satellite images showing widespread cumulus congestus over northern Australia (a) at 0230 UTC and the development of storms moving across Darwin (D) (b) by 0430 UTC 9 Dec 1997.
tures that are evident in these cross sections include an overhanging reflectivity maximum on the edge of the cells with intense upward motion driving through it (e.g., around 0314 and 0324 UTC). These overhangs are also seen in C-Pol RHI scans (Fig. 5). The overhang regions were associated with intense updrafts, usually exceeding the terminal fall speed of large rain drops. The updrafts typically extend up to the echo top or just above (~500 m). The reflectivity shows the storm top at about 7–8 km, extending approximately 4 km above the FZL allowing a mixed-phase region, but not demonstrating an acceleration above the FZL, which characterizes cross sections through many squall lines (e.g., May and Rajopadhyaya 1996). The magnitude of the vertical motions is large, and often exceed 10 m s$^{-1}$. This is greater than the maximum velocities observed at the 90th percentile level in oceanic convection across many basins and many storm conditions (cf. Lucas et al. 1994). Downdrafts are seen both within the convective cell and in the region adjacent to the updrafts on the outside of the cells. This example does not show it clearly, but there is often a small reflectivity maximum on the outer boundary of the updraft that probably marks the cloud edge. The downdrafts surrounding the cloud

---

**Table 2. Sounding data for storm cases.** The 13 Dec storm was in late morning while the others occurred in the afternoon.

<table>
<thead>
<tr>
<th></th>
<th>2300 UTC</th>
<th>2300 UTC</th>
<th>2300 UTC</th>
<th>2300 UTC</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAPE (J kg$^{-1}$)</td>
<td>2577</td>
<td>1775</td>
<td>404</td>
<td>3412</td>
</tr>
<tr>
<td>CAPE at 500 hPa (J kg$^{-1}$)</td>
<td>618</td>
<td>393</td>
<td>93</td>
<td>653</td>
</tr>
<tr>
<td>CIN (J kg$^{-1}$)</td>
<td>-8</td>
<td>-36</td>
<td>-86</td>
<td>-3</td>
</tr>
<tr>
<td>Lifting condensation level (hPa)</td>
<td>934</td>
<td>933</td>
<td>904</td>
<td>939</td>
</tr>
<tr>
<td>Level of free convection (hPa)</td>
<td>880</td>
<td>843</td>
<td>697</td>
<td>915</td>
</tr>
</tbody>
</table>

---

*Fig. 4. Reflectivity fields at an elevation of 0.94° at (a) 0248, (b) 0300, (c) 0312, and (d) 0325 UTC on 9 Dec 1997. The C-Pol radar is located at the origin. The X marks the profiler location. Wind and temperature data from AWSs are plotted. The outflow from C and G interacted with the sea-breeze front (SB) to initiate the cell that moved over the profiler (P) and a nearby cell that merged with P (C2).*
is a feature also observed with shallow cumulus (e.g., Kollias et al. 2001).

The time–height cross section of $w$ shows a second region of enhanced vertical motion and reflectivity at 0337 UTC (not shown). However, examination of the weather radar data shows that this is not a separate cell. Rather, $P_1$ is growing as it moves past the profiler and this growth extends the edge of the cell back over the profiler at that time. This illustrates some of the difficulties associated with using a single vertically pointing instrument to observe the characteristics of convective cells and the need to have corresponding weather radar data for confidence in the interpretation of the data. Again, the persistence of the strong updraft on the periphery of the cell is seen. The storm growth is also evident in the reflectivity time–height section with significant echoes extending above 10 km after 0325 UTC.

Radiosonde soundings taken within a few hours of each of the storms were examined. In particular the available CAPE up to the 500-hPa level was calculated. This provides an upper limit to the theoretically attainable vertical motion (CAPE $\sim 1/2\rho w^2$, where $\rho$ is the air density and $w$ is the vertical velocity of an undilute parcel), which in this case was in the range of 6–16 m s$^{-1}$. While there will be a significant fluctuations in the CAPE depending on the exact situation with respect to local circulations that the sondes ascended through (e.g., Weckwerth et al. 1996), it is clear that the maximum updraft speed within these storms is very close to the maximum velocity of an undiluted parcel. However, the updrafts decelerate quite markedly above the FZL and do not extend much beyond about a height of 7.5 km over the profiler. The continued rate of storm growth was consistent with similar updrafts extending to near the 15-km echo top at later times. In other cases there was no development above 10 km despite the radiosonde profiles showing large potential buoyancies above this.

Intense downdrafts such as are seen in the center of

![Figure 5. RHI scan over the profiler site (range marked as the vertical line at 23.8 km) of the $Z$ with a contour of $Z_{dr} = 3$ dB overlain, and radial velocity at 0309, 0316, and 334 UTC 9 Dec 1997. The cell motion has been subtracted from the velocity fields.](image-url)
the cell occur in all the storms. This is an indication of the effect of precipitation loading, evaporation, and the descending cool air penetrating the boundary layer and subsequent outflow probably cuts off the inflow of high-$\theta_e$ air into the storm leading to its subsequent decay. The intensity and timing of the downdraft is consistent with the divergence seen in the 0334 UTC RHI. There is also an upper-level area of strong downward motion over the center of the storm. This was only seen in this case and its origin is uncertain.

The high-reflectivity areas (>50 dBZ) near or within the updrafts invariably contained significant contributions to the echo power from 1–2-cm diameter hail. These were seen as low as 2.6 km despite the quite long residence times for the particles to fall from the FZL to this height (~4 min). Note that only dense hail particles have sufficient fall speed to overcome the vertical motion, and the intense flanking updraft can be imagined as an efficient hail-creating engine as small water drops are lofted up to the suspended large hail particles and efficient riming occurs. In this case we have a cross section through the center of a cell, so that updrafts are present on both sides of the strong echo region. In this example the hail distribution is scattered throughout the cell, but again the highest concentrations are visible near the updraft boundary.

There were large median drop sizes near the reflectivity maxima and near the storm edge (Fig. 5e). The latter is usually seen at the surface and is a result of velocity sorting. There is always deep descending motion within the storm cell indicating the effects of evaporation and drop loading that result in the decay of the cells as the inflow into the updrafts is interrupted. The profiler estimates of total water volume are very large in these storms with water densities exceeding 3–4 gm m$^{-3}$.

As noted, the intense updrafts on the edge of the storms (radar echoes) produce a flux of supercooled drops above the FZL that allows the production of hail through a drop freezing and riming process with di-

FIG. 5. (Continued)
ameters of 1–2 cm (e.g., Braham 1964). This has a profound effect on the polarimetric signatures as these large particles fall out of the upper levels reaching down to heights of only 1–2 km above the ground (Vivekanandan et al. 1990; May et al. 2001). The $D^6$ dependence of the reflectivity means that a relatively few particles dominate the reflectivity and fall speed spectrum creating difficulties for the quantitative estimation of DSDs from fall speed spectra. Since the water volume scales as $D^3$ and rain rate as $\sim D^{3.67}$, the presence of such large wet particles also contaminates the resulting estimates. The total mass of the hail was small compared with the rain mass below about 4 km. However, the contributions to the radar reflectivity by the melting hail is large and in some cases dominant because of the relatively large hail size ($\sim$10–20 mm). Our retrievals will not differentiate hail smaller than about 10 mm from rain, but it is a reasonable assumption that the bulk of the smaller hail will melt at heights significantly higher than the height where the larger hail melts.

The reflectivity field observed with C-Pol shows a spatial structure that is similar to the time–height cross section (e.g., Fig. 4). Of more interest is the further insight into the microphysical structure that can be gained from combining the profiler and polarimetric radar data. For example, the large values of $D_0$ correspond closely to the vertical regions of enhanced differential reflectivity ($Z_{DR}$) extending to near the FZL—so-called $Z_{DR}$ columns such as that seen at a range of about 22.5 km in Fig. 5a. Note that these do not necessarily correspond to strong updrafts, illustrating some of the complexity and the transient nature of many of the features in the radar images. The $Z_{DR}$ columns correspond to both regions of large $D_0$ for the rain and wet hail, both of which are known to contribute to the $Z_{DR}$ enhancements (Vivekanandan et al. 1990; May et al. 2001). The de-
Fig. 6. A time-height cross sections of the vertical motion, radar reflectivity (from 920-MHz profiler), and microphysical characteristics of a the 11 Dec storm passing over the profiler. The intense updraft region is marked by the arrows in the reflectivity and vertical velocity frames. The microphysical retrievals have been summarized as median volume diameters for hail and rain, and estimates of the water content for the two species (see section 2 for details). The shaded areas are for the rain content and median volume diameter. Contours for hail mass content exceeding 0.005 gm m$^{-3}$ are superimposed over the rain content and for hail diameter exceeding 10 mm over the rain median volume diameter.

Increasing $Z_{DR}$ after 0316 UTC indicates the vertical motions within the storm, at least near the freezing level, were decreasing along with the production of large drops and hail. In this case, the $K_{DP}$ values are small, so that the column at a range of 23.8 km is predominantly associated with melting, more spherical, hail.

4. Conclusions

This study illustrates both the complex microphysics in relatively shallow tropical thunderstorms that form on sea-breeze fronts and the power of the combined wind profiler and polarimetric radar measurements. A conceptual picture of the evolution of these storms is shown in Fig. 7. The growth from a shallow cumulus stage with an updraft core and downdrafts surrounding the cloud to a deeper stage as was sampled here. By then the updrafts are penetrating the freezing level allowing freezing to occur and hail production through riming. Precipitation loading and evaporation of the rain produce a downdraft core in the center of the storm that ultimately produces an outflow that cuts off the inflow of high-$\theta_e$ air and the storm decays. In the example discussed in detail this decay stage is delayed by a merger with an adjacent cell allowing deeper development still. The intensity of the updrafts is high, exceeding 10, and often reaching 15, m s$^{-1}$, close to the maximum possible updrafts based on nearby soundings.

The question of what limits the growth of these cells to near the freezing level on most occasions, rather than continuing to accelerate, is an interesting one. It is possible to speculate that the limit to growth is actually reached when the central downdraft reaches the surface and the divergent flow cuts off the inflow. That is, there is a race between the updraft and downdraft so that unless there is additional forcing, such as during the
merger in the case studied in detail, the cold pool production provides the limit to storm growth in these low shear environments (see, e.g., Xu and Moncrieff 1994).

Enough hail is generated in these shallow storms for it to dominate the reflectivity, even though the actual hail concentrations may be quite modest. This accompanies the most intense updrafts of the storm system. These results are consistent with those of Jameson et al. (1996) where it was shown that storms forming on the sea breeze in Florida that became electrically active have important implications for radar measurement of rainfall and the interpretation of "ZDR columns."

These observations supply the detailed circulation and microphysical structure of storms and provide strong tests for cloud resolving models. The analysis techniques described in this paper are now being applied to a number of storms and are being extended by the addition of 50-MHz Radio Acoustic Sounding System (RASS) data to provide temperature measurements and hence buoyancy fields through up to heights above the FZL. A trial experiment has been conducted in northern Australia and further experiments are planned including observations in the midwest of the United States.

Acknowledgments. This work has been supported by NSF Grant ATM-9419523.

REFERENCES


