A preliminary morphology of precipitation systems in tropical northern Australia

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SUMMARY

A preliminary morphology of convective systems observed in the vicinity of Darwin, Australia is presented. Several types of tropical convection during monsoonal and break-season flow are identified using specific examples and compared with a range of systems observed worldwide in the general context of the bulk Richardson number. A significant spectrum of convective activity ranging from low shear and low Convective Available Potential Energy (CAPE), to high shear and moderate to high CAPE, typical of mid-latitude severe storms, is identified. Significant differences between monsoonal flow and break-season systems are apparent. The formation of convective lines perpendicular to the low-level shear and the apparent balance between advancing shallow cold pools and the shear are ubiquitous features. Convective-scale downdraughts and mesoscale descent appear to be responsible for redistribution of 700 mb momentum to the planetary boundary layer (PBL).

The motion of the break-season squall lines appears to result from an equilibrium between the cold pool and the undisturbed environmental PBL flow. Deviations from this balance are hypothesized, to result in the observed 'propagating' and 'non-propagating' or slow-moving modes. Examples of these modes are given, showing the classical continuous development of new cells in a balanced state along the leading edge of long-lived squalls and, in the propagating case, the discontinuous development of new cells on the cold pool ahead of the squall.

1. INTRODUCTION

An observing station has been established at Darwin, north Australia (12°S, 131°E) to provide the monsoon rainfall data that are required for the Tropical Rainfall Measuring Mission (TRMM) of the National Aeronautics and Space Administration (NASA) and the Tropical Oceans and Global Atmosphere (TOGA) Program.

The Darwin station houses the National Oceanic and Atmospheric Administration (NOAA)/TOGA Doppler radar, and provides rawinsonde data every 12 hours, wind soundings every 6 hours, and has a mesoscale network of 26 tipping-bucket rain-gauges and a surface observing network. Details of the Darwin area, the network and observational procedures are given by Keenan et al. (1988). An outline of the area and basic observing network is given in Fig. 1(a).

The aim of this paper is to present a preliminary morphology of the precipitation systems that were observed primarily during the 1987/88 summer monsoon period. Climatological characteristics of Darwin are presented together with examples of precipitation systems observed during monsoon and drier transition and break periods. These examples are intended to provide a representative selection of the precipitation systems in this region based on the experience of the authors. The examples will provide some information on the mesoscale structure, differences between organization in break- and monsoon-season flow, mechanisms involved in the initiation and evolution, system-relative flow and propagation of the squall lines. This study does not purport to portray all modes of convection that occur at this location, or their frequency, and does not attempt to provide an in depth analysis of each case; such detailed case studies are in

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progress elsewhere. The work presented herein attempts to provide a background to these forthcoming case studies so that they can be put in context and seen as part of a spectrum of convective activity.

Observational studies and results of numerical simulations of convection, summarized by Weisman and Klemp (1986), provide substantial evidence that the growth, intensity and dynamical properties of many storms are strongly dependent upon environmental factors such as thermodynamic instability and vertical wind shear. Variability of storm types is also related to mesoscale forcing, initiation mechanisms, vertical variability in shear, moisture, stability and microphysical properties. These factors notwithstanding, there is ample evidence suggesting that the basic physics and continuum of storm types are determined by the Convective Available Potential Energy (CAPE) and vertical shear of the lower tropospheric winds. This point is discussed by Carbone et al. (1990a), where the relative location of convective circulations such as frontal rainbands, tropical circulations, severe storms and subtropical storms within this two-dimensional buoyant energy/kinetic energy coordinate system are postulated. Herein, we will adopt this basic approach to map the spectral regime occupied by the convective storms observed at Darwin, and compare the extent to which this spectrum overlaps with the continuum of convective modes observed elsewhere.

Finally, some characteristics of squall-line propagation are discussed. The speed of Darwin tropical squall lines was, in general, very close to that of an easterly jet maximum located near 700 mb. However, there were exceptions with some cases moving faster and some cases moving slower than the 700 mb flow. Observational studies in other tropical regions tend to support this conclusion, with an overall mean trend, relative to the easterly jet, for the tropical systems to propagate slightly. For examples see Betts et al. (1976) for Venezuela, Aspliden et al. (1976) for west Africa and Barnes and Sieckman (1984) for Global Atmosphere Research Program Atlantic Tropical Experiment (GATE).

The speed of motion of squall lines is influenced by cold-pool interaction with the undisturbed planetary boundary layer (PBL). In principle, this interaction can result in either transporitive or truly propagating systems, as discussed by Carbone et al. (1990b). However, PBL effects are not occurring in isolation. Numerical simulations by Nicholls et al. (1988) and Lafre and Moncrieff (1988) conclude that interaction between the storm environment and the storm circulation affects the cold-pool development and hence the potential characteristics of the storm motion. Mesoscale properties of the storm, in the form of rear-to-front flow, also affect the cold-pool development, as shown by Smull and Houze (1987) and Chalon et al. (1988). Based on the observational evidence obtained herein, other observational studies, and modelling investigations, some concepts thought relevant to the relationship of environmental flow, storm structure and the tendency for storm propagation will be presented.

2. CLIMATOLOGICAL CHARACTERISTICS OF THE SUMMER MONSOON

Darwin is situated in the maritime continent region (Ramage 1968) and experiences pronounced wet and dry seasons. The wet season usually extends from November through April and is characterized by the occurrence of over 90% of the annual rainfall. It has low-level equatorial origin westerly flow with so-called ‘transition’ and ‘break’ periods when there is a low- to mid-level easterly flow of continental origin. Transition flow occurs at the beginning of each wet season and terminates with the onset of deeper
westerly monsoonal bursts that are usually accompanied by the onset of widespread precipitation. These monsoonal bursts typically last less than a month and may occur several times during the wet season. With the breakdown of monsoonal bursts there is a return to break-period flow; this is similar to the transition-season flow. In this study, monsoonal flow is defined as occurring whenever the daily pressure-weighted surface to 3 km flow was westerly with magnitude greater than 4 m s\(^{-1}\), and break and transition flow whenever this pressure-weighted flow was easterly and greater than 4 m s\(^{-1}\). Note, these definitions are based solely on the wind flow and do not take into account the occurrence of precipitation in the regime definition, as done by Troup (1961). Precipitation does occur throughout the wet season but, as this paper will show, there are some very different characteristics evident in the monsoonal-season and break- or transition-season precipitation systems.

As shown in Fig. 1(b) the 1987/88 break-period easterly flow had a relatively uniform layer of shear extending to the easterly maximum near 700 mb. While westerly flow is normally encountered in the 400-150 mb level during the break and transition periods, the 1987/88 wet-season mean showed weak easterly flow in this layer. During the monsoon westerly flow, the low-level shear was reversed and the surface shear layer was very shallow. Maximum westerly winds occurred at 900 mb but extended to 400 mb in a deep weakly-sheared layer. Easterly flow was apparent above 400 mb during the westerly monsoon flow.

CAPE\(^1\), as defined by Moncrieff and Green (1972), is comparable during both the break and monsoon periods although the Convective Inhibition\(^2\) (CIN), as defined by Colby (1983), is slightly larger for the former. The mean precipitable-water\(^3\) content for the two regimes is large and typical of tropical situations. In comparison, they are at least twice the size of values found in the Oklahoma data of Bluestein and Parks (1983).

Note, the computations of CAPE are very sensitive to the virtual temperature and initial surface conditions. A 1 g kg\(^{-1}\) increase in mixing ratio or an increase in temperature of 1 degC can increase the CAPE by 500 J kg\(^{-1}\). In addition, the number of monsoon days available to this study was small.

We will now present some examples of convective systems and relate the occurrence of these systems to specific environmental regimes and forcing mechanisms. The examples have been selected to highlight the population of systems encountered at Darwin and to elucidate some features relevant to convective structure and propagation mechanisms. A primary purpose of this paper is to document the variability in structure and dynamical regimes encountered at this tropical location.

\[ ^1 \text{CAPE} = \int_{z_2}^{z_1} g \left( \frac{\theta_e - \theta_{env}}{\theta_{env}} \right) dz \]

where \( g \) is the acceleration due to gravity, \( \theta_e \) is the potential temperature of an air parcel having been lifted from the surface, \( \theta_{env} \) is the potential temperature of the unsaturated environment, \( z_1 \) is the level of free convection and \( z_2 \) is the equilibrium level. It is the work per unit mass done by the environment on an air parcel that rises from \( z_1 \) to \( z_2 \). In this paper, following Bluestein and Jain (1985), \( z_1 \) is determined by lifting from the surface an air parcel having the potential temperature and mixing ratio weighted by pressure over the lowest 500 m.

\[ ^2 \text{CIN} = -\int_{z_0}^{z_1} g \left( \frac{\theta_e - \theta_{env}}{\theta_{env}} \right) dz \]

where \( z_0 \) is the surface. It is the net work per unit mass required to lift a negatively buoyant parcel from the surface to the level of free convection.

\(^3\) The precipitable water is the amount of water vapour in the atmosphere and is defined as the mass of water vapour in a vertical column of unit cross-section. The saturated precipitable water is the precipitable water of the air column if saturated. The ratio of the precipitable water to the saturated precipitable water is the mean humidity.
Figure 1. (a) Basic observing network around Darwin with radar range rings shown. (b) Mean vertical structures of the 1987/88 break- and monsoon-period flows at Darwin.
3. **Monsoon Flow Systems**

A typical monsoon, as shown in Fig. 2, contains widespread coastal and oceanic precipitation including relatively disorganized and quasi-stationary weak convection. The Range Height Indicator (RHI) in the direction of 270° reveals echo tops between 10 and 14 km with very weak horizontal gradients in radar reflectivity and low vertical shear of the horizontal wind above the 0°C melting level located near 5 km. The most vigorous convection appears to be within 10 km of the radar where low-level reflectivity gradients (Fig. 2(c)) are strongest and echo tops are highest. Radial velocities (Fig. 2(b)) confirm westerly flow below 5 km and easterly flow above. Patches of westerly momentum above 5 km are likely the result of convective momentum transport. A nearly saturated and probably storm-affected sounding (Fig. 2(d)) is evident with a relatively small CAPE of 450 J kg⁻¹ and weak 0–3 km low-level shear of 1.9 × 10⁻³ s⁻¹. The region of precipitation is quasi-stationary throughout the event.

The evolution of a second monsoon system in an environment of moderate CAPE (1700 J kg⁻¹) and moderate shear (3 × 10⁻³ s⁻¹) is shown in Fig. 3. The higher CAPE in this monsoon situation results essentially from the high surface moisture and a deep middle-level layer where the sounding indicates air slightly cooler than moist adiabatic values.

![Diagram](image)

**Figure 2.** Radar structure of a monsoon band on 19 December 1987 as observed at Darwin by the NOAA/TOGA radar: (a) PPI effective reflectivity (dBZₑ) scan at 0.8° elevation at 0540 LST; (b) storm relative radial velocity (m s⁻¹) and (c) RHI effective reflectivity (dBZₑ) along azimuth 270° at 0555 LST; and (d) Darwin rawinsonde sounding. Note, calm conditions are represented by ⌀, winds <2.5 m s⁻¹ are indicated by directions without a barb, each half barb represents 2.5 m s⁻¹ and each full barb 5.0 m s⁻¹. A wet-bulb potential temperature, $\theta_w$, value is shown.

Note, in this paper, the terminology such as weak, moderate and strong for such things as CAPE, shear and echo strength are used relative to mean values. In the case of CAPE and shear, the observed distributions are given in Fig. 15.
Initially, at 23:05 LST 30 December 1987 (LST = GMT + 9.5 h), a series of cells formed parallel to the north-east coast. Data presented by Keenan et al. (1988) show that such diurnal development between the mainland and the islands is quite frequent and this suggests that interaction between land breezes and environmental flow was a probable cause of convection. In this case, the low-level flow was westerly with shear of magnitude
Figure 3. Evolution of a monsoonal rainband on 30–31 December 1987 as observed at Darwin by the NOAA/TOGA radar: (a)–(c) PPI effective reflectivity (dBZ<sub>e</sub>) scan at 0.8° elevation at 2305, 0408 and 0750 LST; (d) an RHI cross-section of reflectivity (dBZ<sub>e</sub>) and (e) storm-relative radial velocity (m s<sup>-1</sup>) along azimuth 235° at 0442 LST and (f) Darwin rawinsonde sounding. See Fig. 2 for sounding details.

7 × 10<sup>-3</sup> s<sup>-1</sup> in the lowest kilometre. Relatively weak convection appears to have been initiated along the convergence line between the land breeze and the monsoonal flow. Lifting, necessary for the release of conditional instability, was generated parallel to the coast and, therefore, parallel to the low-level shear. The relative magnitude and orientation of the shear in the monsoonal and land-breeze flows are factors related to the
Figure 3. Continued.
strength of initial lifting. The data available herein do not permit quantitative assessment of these factors.

Embedded within the band are squall-like regions of enhanced precipitation with a suggestion of a 50 km wave pattern, e.g. at 23:05 LST the cells C_1, C_2, C_3 and C_4 indicated in Fig. 3(a). The regularity in the spacing is suggestive of a preferred horizontal scale of instability within the confluence region. These enhanced regions of precipitation translated eastward down the band, perpendicular to the low-level shear at speeds between 9–14 m s\(^{-1}\). In addition, each area was associated with a westerly velocity maximum (not shown) indicating downward momentum redistribution and possibly a spreading cold pool. Development of a symbiotic gravity-wave disturbance may be a factor in scale selection of these embedded squall-like disturbances.

Evolution from the initial zonally-forced convective state towards meridionally-oriented squall lines was a common feature, as will be shown in some of the other examples to be presented.

A similar band formed later along the south-west coast and by 4:08 LST 31 December major stratiform areas persisted with embedded convection in relatively disorganized patches and short-line segments (C_5 and C_6 in Fig. 3(b)) moving along the band from the south-west at about 6 m s\(^{-1}\) perpendicular to the low-level windshear. The effective reflectivity exceeds 50 dBZ\(_e\) in the more intense convective cores. The band itself was quasi-stationary but there is some tendency for slow southward movement with a gradual trend toward more linear organization in the convective areas.

By 4:42 LST the entire system was embedded in westerly flow throughout the 10 km depth of precipitation echo as evidenced in Fig. 3(d) which is an RHI along 235° azimuth. The scan plane passes through C_5 and C_6, the two major convective line segments mentioned above that were moving from 200° azimuth at 6 m s\(^{-1}\) within the band of stratiform precipitation. The melting-level is evident from increased reflectivity near 5 km and the depth of convection is seen to range from 10 to 15 km in the near and far bands, respectively. Doppler velocity structure (Fig. 3(e)) reveals downward redistribution of
south-westerly momentum (about 12 m s\(^{-1}\) storm relative or 18–20 m s\(^{-1}\) absolute rear-inflow) in the flow from the 4 km level at 70 km range descending to less than 2 km at 12 km range underneath the convective bands. This low-level westerly jet structure was part of the strong confluence field at the base of deep convection. By inference from reflectivity and velocity structures, we conclude that updraughts are tilted 30° to 40° (from zenith) toward the south-west in a downshear direction (above 3 km). This organization is reminiscent of GATE squall lines as depicted by Houze (1977).

At 7:51 LST a much higher degree of organization evolved when a major south-west to north-east oriented convective band was evident. The precipitation region has reversed motion towards a northerly direction such that the major band was now located off-shore. A lower amplitude, periodic transverse structure in the form of 20–50 km spaced ‘fingers’ (F in Fig. 3(c)) stretching ahead of the line was also evident to the west. This degree of convective organization is unusual for monsoon situations.

4. BREAK- AND TRANSITION-SEASON SYSTEMS

(a) Continental convection

Continental convection is observed regularly during the afternoon and evening periods of the break and transition season. The characteristic evolution consists of scattered isolated cells later aggregating and often forming weak squall lines. Such structure is depicted in Fig. 4. Typically, as shown in Fig. 4(b), the cells are vertically erect and deep with tops above 15 km. Continental cells form under a wide range of conditions. For this example, the CAPE was 920 J kg\(^{-1}\) and the low-level shear was \(2.9 \times 10^{-3}\) s\(^{-1}\).

(b) Downburst event

The occurrence of a downburst event in the Darwin tropical environment is shown in Fig. 5. At 16:44 LST a small divergent velocity couplet is evident from the thunderstorm 10 km north-east of the radar. Approaching velocities up to 12.5 m s\(^{-1}\) coupled with receding velocities of 5.0 m s\(^{-1}\) on a horizontal scale of 4.0 km give a radial horizontal shear of \(4.3 \times 10^{-3}\) s\(^{-1}\). At 16:49 LST (not shown) the approaching (receding) velocity maximum increases to 17.5 (−10.0) m s\(^{-1}\) over 5 km, i.e. a horizontal shear of \(5.7 \times 10^{-3}\) s\(^{-1}\). By 16:59 LST (not shown) the couplet maxima are 7.5 km apart giving a horizontal shear of \(3.9 \times 10^{-3}\) s\(^{-1}\). The horizontal shear values, the spatial scale and temporal evolution of the divergent couplet are consistent with that of microbursts in mid-latitude convection, as described by Fujita (1981) and quantified by Wilson et al. (1984).

Compared with most microbursts studied to date, this event occurred in an environment of moist, stable subcloud air with relatively dry mid levels, as shown by the sounding in Fig. 5(c). Dew-point depressions ranged from 5 degC at 900 mb to 1 degC at the surface. In comparison, dew-point depressions of 20 degC are present in the very dry, deep adiabatic subcloud environment of U.S. plains-type microbursts (McCarthy and Wilson 1984). According to Srivastava (1985) and Wakimoto (1985) this dry layer enables considerable evaporative cooling below cloud base and is significant in maintaining the microburst. Given the moist (precipitable water twice that of U.S. plains-type microburst producing storms), stable and shallow PBL observed in the Darwin sounding, the role of water loading may assume relatively greater importance. Examination of vertical cross-sections through this storm indicates radial convergence into the region of maximum precipitation associated with the microburst. This is consistent with drag induced by precipitation loading, although evaporational processes would also
produce this effect. In this example, the CAPE is moderate (1021 J kg$^{-1}$) and the shear is large ($5.8 \times 10^{-3}$ s$^{-1}$).

(c) **Maritime continent thunderstorms**

Thunderstorms characteristic of the maritime continent are observed over the islands to the north of Darwin on 65% of days during the transition and break seasons. These storms exhibit a variety of forms including both single- and multiple-cell types. A
Figure 5. Microburst on 22 January 1989 as observed at Darwin by the NOAA/TOGA radar: (a) PPI effective reflectivity (dBZ_e) scan at 0.8° elevation and (b) the Doppler derived radial velocity field (ground relative m s^{-1}) for 1644 LST; and (c) Darwin rawinsonde sounding. See Fig. 2 for sounding details.
multicellular type is shown in Fig. 6. The CAPE on this day was 1560 J kg$^{-1}$ and the shear was moderate at $4.5 \times 10^{-5}$ s$^{-1}$. The initial cells at 14:33 LST were aligned in the centre of Melville Island parallel to the shear vector from the surface to 700 mb. The formation was consistent with cell development along a collision zone of two sea breezes. The RHI at 14:26 LST along 355° azimuth (Figs. 6(d) and (g)) is through the cell IT$_1$ highlighted in Fig. 6(a) and indicates convergence to 2.5 km in a 12.5 km high cell with an updraught tilted approximately 30° from zenith in a downshear direction with respect to the flow above 3 km. At this stage, rainfall rates in excess of 100 mm h$^{-1}$ are apparent from the low-level reflectivity structure. The 30 dBZ$_e$ echoes to 9 km indicate considerable ice concentrations within the developing cell.

By 15:20 LST (not shown) the alignment of the developing cells had changed, starting to become perpendicular to the low-level environmental shear. Both storm complexes in Fig. 6(b) are aligned north-west to south-east. At 16:42 LST (Fig. 6(c)) the remaining complex was aligned north–south, perpendicular to the low-level environmental shear exhibiting a squall like appearance and moving to the west. This evolution (from zonally-oriented) relative to the environmental shear is consistent with the dynamics of cold pools, as advanced by Rotunno et al. (1988). The RHI at 15:20 LST along 350° (Figs. 6(e) and (h)), through the leading storm cell IT$_2$ highlighted in Fig. 6(b), indicates a deep updraught extending from a height of 2.5 km to 15 km with an approximately 40° average slope from zenith in the downshear direction for that layer. Weak downdraught inflow and descent from 2.5–5 km are evident behind and under the updraught. The sounding (Fig. 6(j)) shows that the environmental air at this level was relatively dry, a factor that can assist the downdraught development. At this stage, echoes in excess of 38 dBZ$_e$ were observed at 12.5 km ($T = -52 ^\circ C$) again suggestive of an extremely high ice concentration, indicating the suspension of large ice crystals by strong updraught motion.

At 16:30 LST the RHI along 340° azimuth (Figs. 6(f) and (i)) through cell IT$_3$ indicates the island thunderstorm was in a mature stage. One remaining north–south oriented squall is evident moving to the west at 5 m s$^{-1}$. 
The Island Thunderstorm Experiment, described by Keenan et al. (1989), was undertaken during November and December 1988 to obtain experimental data on these systems.

(d) Sea-breeze storms

An example of sea-breeze initiated convection on the mainland and the subsequent generation of a squall line is evident in the high to moderate shear/moderate CAPE environment of 14 January 1988 in Fig. 6. At 14:33 LST an initial east–west oriented line of convection (SB1 in Fig. 6(a)) was evident approximately 15–20 km inland from the coast, roughly 50 km to the north-east of Darwin. Evidence for convection of this type being associated with local sea-breeze circulations is provided by Keenan et al. (1988). At 14:33 LST SB1 was essentially parallel to the low-level environmental shear, as found for the island thunderstorm case discussed previously. The RHI at 14:42 LST along 68°, through SB1 and shown in Figs. 7(a) and (d) indicates that cell tops were reaching 17 km with 30 dBZc echoes extending to 14 km (T = −65°C), i.e. high ice concentrations. The updraught in Fig. 7(d) is sloping upshear away from the radar at 65° from zenith and contains two cores; one at approximately 2.5 km and a second at 7 km. Weak storm-relative descending rear inflow with horizontal momentum almost equal to the speed of the storm is evident undercutting the updraught below 5 km. There is evidence for rearward spreading of the downdraught in the PBL from 45 to 55 km range in the lowest

Figure 6. Evolution of an island thunderstorm on 14 January 1988 as observed at Darwin by the NOAA/TOGA radar: (a)–(c) PPI effective reflectivity (dBZe) scans at 0.8° elevation at 1433, 1537 and 1642 LST; (d)–(f) RHI effective reflectivity scans at the azimuths (AZ) indicated; (g)–(i) corresponding RHI storm-relative Doppler derived radial velocity fields (m s⁻¹) at times corresponding to (d)–(f) at 1426, 1520 and 1630 LST; and (j) Darwin rawinsonde sounding. See Fig. 2 for sounding details.
Figure 6. Continued.
1.5 km and a forward spreading cold pool undercutting the lower maximum in the updraught and extending out ahead of the cell to 30 km range. The spreading pool was probably responsible for the initiation of a new updraught or the lower maximum described above. Maximum reflectivity is evident at 40 km, below the more mature updraught. This updraught was associated with maximum Doppler-observed confluence in the 5–8 km layer.

By 15:37 LST (Fig. 6(b)) the convection (SB₂) was concentrated at the downstream end of a 30 km north–south oriented squall line that was moving perpendicular to the lower-level shear. The RHI at 15:42 LST along 73°, through SB₁ and shown in Figs. 7(b) and (e), indicates an eastward-sloping updraught oriented approximately 70° from zenith in the 2.5 km to 9 km height range. A storm-relative rear inflow maximum of approximately 7.5 m s⁻¹ is evident, descending from 3 km height to form a low-level PBL outflow. Maximum velocities in this spreading pool are approximately 15 m s⁻¹. The maximum updraught is above the region of maximum precipitation. The formation of an upper-level 10–15 km anvil is evident at this time, with maximum outflow directly above the mature updraught.

At 16:42 LST, as shown in Fig. 6(c), the squall (SB₃) generated by the sea-breeze front was 15 km west of the radar and still aligned in the north–south direction, i.e. perpendicular to the low-level shear. An arc-shaped leading edge was apparent, and Doppler data indicate a maximum in receding radial velocities concentrated within the bulge, i.e. a region of enhanced descending rear inflow is apparent. The RHI at 16:41 LST along 270° through SB₃, shown in Figs. 7(c) and (f), indicates the spreading cold pool with maximum storm-relative velocities of approximately 2.5–7.5 m s⁻¹. These velocities are almost equal to the 700 mb wind speed. Again a sloping updraught starting at a height of 2.5 km is apparent above the forward spreading cold pool. An anvil between heights of 10–15 km also extends ahead of the squall.

The mean environmental flow at 700 mb for this case is an easterly of approximately 14 m s⁻¹, almost identical with the 15 m s⁻¹ observed in the spreading cold pool. The
Figure 7. Vertical structure of the development of a squall line initiated from sea-breeze-front convection east of Darwin on 14 January 1988 as observed by the NOAA/TOGA radar. See Fig. 6 for PPI scans and sounding data; (a)–(c) RHI effective reflectivity scans and (d)–(f) corresponding RHI storm-relative Doppler derived radial velocity fields (m s\(^{-1}\)) at the azimuths (AZ) indicated.
Figure 7. Continued
Figure 7. Continued
overall motion of this system, however, averages only 6.5 m s\(^{-1}\), i.e. the system on average moved slower than any environmental wind. The environmental shear in this case is relatively strong at 4.5 \(\times\) 10\(^{-3}\) s\(^{-1}\). It should be noted that the sounding shown in Fig. 6(j) indicates the presence of dry middle-level air. It is expected that entrainment of this environmental air into the storm in the rear inflow jet would create very favourable conditions for the development of downdraughts through evaporation and melting of hydrometeors. Hence, as observed, a means exists for effective redistribution of the middle-level momentum to the PBL. Possible mechanisms responsible for slow motion of the squall relative to the environmental winds will be discussed later.

The interaction between a sea-breeze front and a gust front spreading from pre-existing convective activity is shown in Fig. 8 in a low CAPE (331 J kg\(^{-1}\)) and high shear (5 \(\times\) 10\(^{-3}\) s\(^{-1}\)) situation with low-level stability in the PBL. At 18:15 LST, Doppler data (not shown) indicate a region of receding velocities extending in a radial arc from 30 km to the east to 30 km south of the radar behind the weak 6–14 dBZ, bow-shaped sea-breeze-front echo (SBF) shown in Fig. 8. The evolution of this system away from the coastal regions plus wind-trace and thermographic data from Darwin suggest this echo is the inland penetration of a sea-breeze front. The scatters in this case are expected to come primarily from insects (see Pratte and Keeler 1986).

To the east, a similar 6–14 dBZ, bow-shaped echo (GF) is apparent approximately 3–5 km to the west of a decaying storm. This latter echo is interpreted as a clear-air gust-front signature in the manner described by Wilson and Carbone (1984). At this time, the two bow-shaped echoes are approximately 7 km apart and converging towards each other at a relative speed of approximately 14 m s\(^{-1}\).

Figure 8. Initiation of convection resulting from the interaction between a south-east-moving sea-breeze front (SBF) and north-west-moving gust-front outflow (GF) from a pre-existing thunderstorm on 21 January 1989 as observed at Darwin by the NOAA/TOGA radar: (a)–(c) RHI effective reflectivity (dBZ) scans at 0.8° elevation at 1815, 1821 and 1846 LST and (d) Darwin rawinsonde sounding. See Fig. 2 for sounding details.
At 18:21 LST the gust front and sea-breeze front collided approximately 32 km southeast of the radar (Fig. 8(b)). At the point of intersection there are no echoes from precipitating systems.

At 18:46 LST a major storm with maximum echo strength of 54 dBZ developed approximately 10 km downstream of the point of collision (Fig. 8(c)). The interpretation offered is that the colliding clear-air boundaries facilitated the initiation of the convection. The time lag between the collision and the detection of the storm is typical of that found by Wilson and Schreiber (1986). The downstream displacement is consistent with the 16 m s\(^{-1}\) 700 mb wind speed. As this paper will demonstrate, most storms in the Darwin environment move at a speed very close to the 700 mb flow. In this example, the forcing provided by the interaction between the sea-breeze front and storm outflow boundaries is one mechanism by which development could commence in a potentially unfavourable low CAPE, high shear environment.

(e) Squall lines

Case 1: The classical tropical squall of 18 December 1987. An example of a classical tropical squall line is given in Fig. 9. This system is very similar to those reported from the GATE by Houze (1977), Zipser and LeMone (1980) and Barnes and Sieckman (1984) among others. Such squall lines are noted for continual development of new cells along the leading edge of the convective line; a rear-to-front, descending mid-level inflow jet; and pronounced trailing stratiform precipitation. Often there is a precipitation ‘trough’ between the rear flank of the convective line and the stratiform area. In Fig. 9, these features are evident in the reflectivity and Doppler velocity fields. CAPE is 920 J kg\(^{-1}\) and low-level shear is 1.1 \(\times\) 10\(^{-3}\) s\(^{-1}\). These conditions are very similar to those experienced in the GATE (Barnes and Sieckman 1984). Of special interest is the wave-like pattern in Doppler velocity behind the convective line. The pattern is suggestive of a 20–30 km gravity wave where the first depression (at 28 km) is coincident with the rear flank ‘trough’. Increased vertical shear and vertical mixing is also suggested by the echotop pattern at 15 km range and 10 km height. The rear (easterly) inflow jet is located
at 3 km height and has a storm-relative magnitude of 2.5–7.5 m s⁻¹. The undisturbed environmental flow at 3 km was 5 m s⁻¹, i.e. the storm-relative rear inflow was faster than the environmental wind and of course greater than the speed of the squall line.

Case 2: Squall of 4 January 1988 exhibiting discontinuous propagation. In Fig. 10 a time sequence of squall-line evolution on 4 January 1988 is shown. We distinguish this situation from that shown in Fig. 9 because propagation is clearly discontinuous. Close examination of Figs. 10(a)–(d) reveals a succession of three small convective lines (L₁, L₂ and L₃) which initiated 10–30 km ahead of the previous line. While direct thermodynamic evidence is not available, Doppler velocities at low levels behind each convective line suggest that a rapidly forward-spreading cold pool was responsible in each instance. While each individual convective line translated from 7–10 m s⁻¹, the aggregate convective system propagated at 16.5 m s⁻¹ eastward. This rate of propagation exceeds the environmental easterly winds at any level (12 m s⁻¹ maximum). Only a shallow PBL cold pool transports mass at a rate which equals or exceeds the aggregate propagation rate. Conditions in this case have weak CAPE = 618 J kg⁻¹ and moderate low-level shear of 3.9 × 10⁻³ s⁻¹. In
Figure 10. Time sequence for a squall line on 4 January 1988 at Darwin exhibiting discontinuous propagation: (a)–(d) the effective reflectivity (dBZ$_e$) scan at 0.8° elevation for 1655, 1727, 1742 and 1757 LST and (e) Darwin rawinsonde sounding. See Fig. 2 for sounding details.
this case the sounding (Fig. 10(e)) indicates considerable stability in the PBL. However, it is a morning sounding and by the afternoon the stability in the PBL would have been reduced considerably.

Case 3: Squall of 23 March 1988 showing new cell generation on the cold pool. A relatively shallow squall line in a low shear \((2 \times 10^{-3} \text{ s}^{-1})\), low CAPE \((126.1 \text{ kg}^{-1})\) environment with a stable PBL is shown in Fig. 11. In this case the nearest available sounding (Fig. 11(b)) was in the morning, but with the onset of surface heating the observed PBL stability may well have been reduced by the time this system developed. The squall was first tracked at 14:42 LST from a position approximately 80 km east of Darwin and moved to the west at 10 m s\(^{-1}\). By 17:45 LST the squall was 30 km west of Darwin.

At 15:46 LST an RHI through this system (not shown) indicated the presence of a 16.3 m s\(^{-1}\) PBL velocity maximum emanating from beneath the leading edge precipitation maximum. By 15:59 LST a -2 to 6 dBZ\(_e\) bow-shaped echo (not shown) was apparent approximately 5 km ahead of the squall. This bow-shaped line (BSE) is evident in Fig. 11(a), 26 km east of the radar and approximately 4–5 km ahead of the main body of the squall. The structure of this echo is typical of that associated with spreading cold pools, as described by Mueller and Carbone (1987).

New cell development within the squall system is linked directly to this spreading cold pool. As shown in Fig. 11(a), along the southern flank of this gust-front echo, new cells of up to 46 dBZ\(_e\) were developing approximately 10 km ahead of the previous line position. By 16:30 LST (not shown) these cells formed the main convective elements within the squall. As the gust front continued to move to the west at approximately 10.7 m s\(^{-1}\) new cell generation always occurred on this gust front. The gust-front signature was even discernible at 17:09 LST (not shown) approximately 15 km to the west of the radar over the ocean. Note, the 700 mb easterly flow was 11.3 m s\(^{-1}\), only slightly faster than the mean speed of both the gust front and the squall-line system.
Case 4: Squall of 13 January 1988 with mid-latitude characteristics. A squall line on 13 January 1988 that exhibited characteristics similar to mid-latitude systems is shown in Fig. 12. The sounding for this day (Fig. 12(e)) indicates very dry air above 1.5 km. The plan view of reflectivity (Fig. 12(a)) shows slight bowing of the shape of a vigorous convective line together with a weak trailing precipitation region. The reflectivity and storm-relative radial velocity in the RHI, along azimuth 63° (direction indicated in Figs. 12(a) and (b)), reveal a well-developed leading anvil and an updraught which is tilted upshear at about 55° from zenith. This is a consequence of unusually deep and strong easterlies aloft which exceed the squall-line translation speed by about 5 m s⁻¹. Convective cells developed continually at the leading edge as predicted by the theory of Rotunno et al. (1988). An advancing surface cold pool is implied in the PBL beneath the precipitation cores with very strong approaching Doppler velocities up to 2.5–7.5 m s⁻¹ in excess of the storm motion near the surface. Intense vertical shear of the horizontal wind was evident at 43 km range, between 1.5 and 2 km, where shear exceeds 2 × 10⁻² s⁻¹. CAPE is quite large, 2050 J kg⁻¹, a value comparable with that found in mid latitudes. Environmental low-level shear was moderate at 4.2 × 10⁻³ s⁻¹. Convective line movement was toward the west at 12.5 m s⁻¹. This speed is faster than low- to mid-tropospheric easterlies and slower than easterly momentum generated by the hypothesized spreading cold pool.

Case 5: The intense squall of 16 January 1988. An RHI of a squall system with a relatively large CAPE of 2100 J kg⁻¹ and moderate low-level shear of 3.2 × 10⁻³ s⁻¹ is shown in
Figure 12. A squall line with the characteristics of a mid-latitude system on 13 January 1988 as observed at Darwin by the NOAA/TOGA radar: (a) PPI effective reflectivity (dBZ) scan at 0.8° elevation and (b) PPI Doppler derived storm relative velocity field (m s\(^{-1}\)) at 0248 LST; (c) RHI effective reflectivity along azimuth 63° and (d) corresponding RHI of Doppler derived storm-relative velocity (m s\(^{-1}\)) for 0241 LST and (e) Darwin rawinsonde sounding. See Fig. 2 for sounding details.
Fig. 13. The RHI is at 68° azimuth and shows a stratiform region ahead of the westward-moving squall line. The stratiform region was, in fact, the product of a preceding squall line with the new convective line overtaking it from the rear. Note the elevated reflectivity maximum between 5 and 9 km height, the rear-flank weak echo vault between 9 and 14 km height and the strong, mid-tropospheric confluence several kilometres below, with an associated weak echo or notch. The depth of convection extended to nearly 20 km.

Case 6: The two-dimensional squall of 16 December 1987. An intense and very two-dimensional convective line that occurred on 16 December 1987 is shown in Fig. 14. The squall was oriented zonally and moving northward in an environment where the lower tropospheric flow was westerly. The two-dimensional organization developed after the merging of two cellular convective lines, one of which was propagating northward. Soon, as shown in Fig. 14(b), the organization reverted to the more typical, cellular convective line with trailing stratiform region. Thermodynamic soundings are not available for this case and wind soundings do not reveal unusual vertical shear in the lower troposphere.

Clearly there is a diversity of squall-line structures in the Darwin area which sometimes differ markedly from tropical squall organizations observed in the GATE area and in west Africa.

5. Convective organization and mesoscale environment

The aim of this section is to consolidate ideas by examining the relationship of convective organizations observed near Darwin with the variety of convective systems observed elsewhere. In addition, some concepts relevant to squall-line motion will be discussed.

(a) Characteristic features of the convection

(i) Monsoon/Break differences. The monsoon and break-season systems exist in significantly different shear regimes. Monsoon systems develop in deep oceanic westerly
flow where shear is concentrated in the lowest 0–1.5 km. Break-season systems exist in a shear regime dominated by the 3 km easterly wind maximum.

The extreme vertical development, the 10–20 dBZ limits higher reflectivities above the melting level, and lack of large stratiform decks (except for squalls) are generally the major differences that characterize break-season systems of continental origin relative to monsoon or oceanic origin systems. Large reflectivities (>40 dBZ) do occur in the monsoon systems but these high reflectivities are typically confined to below the melting level. The monsoon system structure is consistent with active warm rain coalescence processes, weak updraughts incapable of supporting large drops, and generally glaciated conditions above. Oceanic systems observed by Jorgensen and LeMone (1989), Gamache (1990) and Rutledge et al. (1991) have similar properties. The many examples of break-season storms suggest intense updraughts and a well developed mixed phase region. The comparisons of the GATE and continental thunderstorms undertaken by Zipser and LeMone (1980) and Szoke et al. (1986) also showed much larger mean updraught and downdraught motion in the continental rather than oceanic systems. These differences in structure are felt to be responsible for the increased electrical activity during break-season periods, as described by Rutledge et al. (1991).

CAPE does exist in monsoon periods, as indicated by the period averages presented in Fig. 1, but the amount is variable. Monsoon CAPE is characteristically manifested over deep layers where the air is slightly cooler than moist adiabatic values as shown by the examples presented in Figs. 2 and 3. When CAPE exists under monsoon conditions, e.g. the case in Fig. 3, significant convective and squall-like features can develop within the monsoonal flow and within the existing and extensive stratiform decks.
Figure 14. An intense squall undergoing a transition from a two-dimensional structure on 16 December 1987 as observed at Darwin by the NOAA/TOGA radar. (a)–(b) PPI effective reflectivity (dBZ_e) fields at 0.8° elevation at 1020 and 122418ST.
(ii) **Initiation and evolution of convective line features.** Initial forcing, both in monsoon and break-period flow, tends to be dominated by coastal land/sea differences. Sea and land breezes are important as initiation mechanisms and often interactions between existing storms and these local circulations generate storms, as shown in Fig. 8. In the case of break-season squall lines, initiation tends to occur on the Oenpelli escarpment region approximately 300 km east of Darwin, on ranges approximately 100–200 km southeast of Darwin or on sea-breeze convergence zones. As a result of the initial forcing, convection can be aligned parallel to the low-level shear (see $C_1$, $C_2$, $C_3$ and $C_4$ Fig. 3(a) for a monsoon case and $IT_1$, $SB_1$ in Fig. 6(a) for a break-season case). This study has shown that these initial convective features tended to evolve towards a linear system oriented perpendicular to the low-level shear. This evolution was observed consistently with all types of convection and was associated with the development of rear inflow and a middle-level downdraught resulting in momentum exchange to the cold pool. Examples of this process are provided in the RHI of monsoonal system convection, shown in Figs. 3(d) and (e) and the RHI through the sea-breeze storm shown in Fig. 7(e). New cell growth within the storms then occurs during the interaction of the cold pool with the environment. Examples of new cell growth over the cold pool are given in Figs. 7(b), 11(a), 12(c) and (d). The result is a tendency to develop convective systems aligned perpendicular to the shear with forward-moving squall-like characteristics.

(iii) **Periodic structure in convection.** Periodic convective structure within lines has been noted in several examples. In the monsoon case, presented in Figs. 3(a) and 3(b), the occurrence of convective elements within pre-existing bands was noted. These squall-like convective elements tended to move down the bands and were associated with areas of enhanced rear inflow, downward transport of momentum and increased surface convergence, i.e. the convective elements had evolved into the perpendicular-to-shear squall structure mentioned above. The reason for this periodic scale selection is not known.

Another feature, often observed on the leading edge of squall lines, were the ‘finger’ type echoes (see Fig. 3(c)) advancing down shear perpendicular to the line. These finger echoes, spaced laterally by 20–50 km, appeared to represent new cells initiating up to 50–60 km ahead of the main line and may represent some gravity-wave disturbance excited ahead of the advancing cold pool.

(b) **Bulk Richardson number depiction**

A convenient method of classifying the spectrum of organized convective systems is through the CAPE and kinetic energy associated with the vertical shear of the flow, e.g. Moncrieff and Miller (1976), Weisman and Klemp (1982), Rasmussen and Wilhelmson (1983), Carbone et al. (1990a) and Ray (1990). Such a classification provides a basis for incorporating the roles played by buoyant energy of the storm and that of the environmental kinetic energy with which the storm interacts.

The variation in low-level shear and CAPE in the manner of Rasmussen and Wilhelmson (1983) from all available rawinsonde data, under various flow regimes from the 1987/88 season, is given in Fig. 15(a). The scatter is considerable, with CAPE ranging from zero to greater than $3700$ J kg$^{-1}$. The vast majority of the sample, however, have CAPE values less than $2500$ J kg$^{-1}$ and shear between $0$ and $6 \times 10^{-3}$ s$^{-1}$, as the outline in Fig. 15(a) indicates. There is association between CAPE and shear, albeit weak, with a correlation coefficient of only 0.27.

Monsoon cases tend to reside at the low-shear end of the spectrum but exhibit considerable variability in CAPE. Some of the largest CAPE values encountered are for
Figure 15. (a) Variation of buoyant energy and low-level shear during 1987/88 wet season at Darwin. The data are stratified as monsoon flow (pressure-weighted surface to 1.5 or 3 km flow >4 m s⁻¹ westerly), well defined break or transition easterly flow (pressure-weighted surface to 3 km flow >4 m s⁻¹ easterly) and other periods of intermediate flow (pressure-weighted surface to 3 km flow <4 m s⁻¹); (b) storm occurrence as a function of buoyant energy and low-level shear. See text for a description.
the monsoon regime. Low shear values are a consequence of a very shallow shear layer which is typically 1.5 km deep. Using the lowest 1.5 km to represent the monsoonal shear (see Fig. 1), the monsoon shears represent the largest values in the sample. However, the strong shear in these cases is a reflection of the shallow layer and is not considered as dynamically significant as a deeper shear layer of similar magnitude. In general, the monsoon is characterized by rather strong, but very shallow, shear and wide variability in CAPE. There is no clear distinction in ranges of CAPE and shear between break-or transition-season flow and the so-called intermediate flow, i.e. flow not consisting of well defined surface to 700 mb easterly or westerly flow.

The association of storm types with CAPE and shear is given in Fig. 15(b) for the limited sample available to this study. In the context of this paper, low-level shear is taken to be representative of the depth of the naturally occurring shear layer. Thus various levels have been used to calculate the low-level shear, but all range over the lowest 20–35% of the depth of convection. For example, for hurricane and cold-frontal rainband data we employ shear layers 1–2 km deep, for Darwin we use the 0–3 km shear, and data taken from Rasmussen and Wilhelmson (1983) use the 0–4 km shear.

The data presented in Fig. 15(b) are based on the following:

1. Tornadic storms (T), non-tornadic storms with a mesocyclone (R), and various other North American storms (●) as given by Rasmussen and Wilhelmson (1983);
2. Darwin systems: squall lines (S), monsoon systems (M), the example of continental convection (C), the microburst storm (MB), a mean for island thunderstorm days (I) from Keenan et al. (1989), the monsoon mean (M) using a 0–3 km shear layer, the monsoon mean (MI) using the 0–1.5 km shear layer, the mean of all Darwin squall lines (S) and the break mean (B) as defined above;
3. GATE fast- (Gₚ) and slow- (Gₛ) moving squalls as given by Barnes and Sieckman (1984);
4. Mean values representative of hurricane rainbands (H) using values given by Barnes et al. (1983) and Barnes and Stossmeister (1986);
5. Frontal rainbands (Fₛ and Fₚ) as presented respectively by Carbone (1982) and Hobbs and Persson (1982); and

The variability in CAPE at Darwin from day to day and within each day can be large. Care should therefore be exercised in the interpretation of Fig. 15(b) especially for the individual examples plotted therein. In the context of this paper, Fig. 15(b) is intended to illustrate the general range of conditions in which storms occur. Differences in CAPE less than 500 J kg⁻¹ are not considered significant for individual storms. A basic problem is that Fig. 15(b) represents only a two-dimensional representation of a multi-dimensional problem. As discussed by Carbone et al. (1990a) and others, the vertical distribution of shear and stability, the mode of cell initiation and the vertical distribution of moisture, may influence the mode of convection. In this case, compared with drier non-tropical locations, the increased water loading will likely effect the net buoyancy in convective updraughts. This difference, on average, could shift the Darwin CAPE values toward conditional neutrality by approximately 300 J kg⁻¹. Another potentially significant difference between a typical Darwin sounding and mid latitudes is the absence of large directional shear in the lower troposphere.
The most significant feature evident in Fig. 15(b) is the considerable range of buoyant energy and wind shear over which this limited sample of Darwin storms extend. They range from the expected tropical conditions of relatively low shear and buoyant energy to that expected from mid-latitude severe storms. The centroid of the Darwin squall cluster (S) has somewhat less CAPE, but more shear, than the mean representative of the environment of North American plains-type non-severe squall lines of Bluestein et al. (1987). From Fig. 15(a), the dynamic range of the spectrum of systems encompasses that of the Oklahoma severe squall lines presented by Bluestein and Jain (1985) and extends into the realm of systems exhibiting mesoscale cyclonic rotation and possibly even tornadic storms. That such phenomena could occur or be postulated to occur from such a small sample of cases is surprising, but is supported from Bureau of Meteorology reports of tornadic damage on occasions in the Darwin area.

The "typical" tropical squall line, i.e. as discussed by Houze (1977), is somewhat dissimilar to many of the examples presented herein. The location of this storm in Fig. 15(b) is toward the low-shear extreme of Darwin conditions at (920, 1.1 × 10⁻³).

As Darwin is affected by tropical cyclones, it would be expected that low buoyant energy/average shear range of hurricane rainbands shown in Fig. 15(b) would also be experienced in Darwin. Only the very high-shear/low-buoyancy regions of frontal rainbands and high-shear extreme buoyancy regimes are not represented in the Darwin cases examined.

As discussed above, the monsoon systems also demonstrate a significant range of possible modes. The case in Fig. 3 has a CAPE of 1700 J kg⁻¹ with only a small departure from the almost moist-adiabatic profile expected for monsoon conditions. It also contains various convective line structures both in the early and late stages of its lifetime. During these periods, the PBL shear is as strong as 5 × 10⁻³ s⁻¹. The mean monsoon location using the 0–1.5 km shear (M) reflects these properties. It has considerable CAPE and a low-level shear greater than found for the squall mean (S). Of course, the monsoon shear is very shallow and therefore does not produce the severe storm types found with the deeper transition and break-season flow.

The relationship of Darwin data to those of the much examined GATE systems raises questions about representing the tropics. The GATE fast movers of Barnes and Sieckman (1984) are very near the break-period Darwin mean (B) and have a lower environmental shear than for the mean of the Darwin squalls (S). The GATE slow movers are at the extreme lower edge of the Darwin spectrum. The monsoon mean (M) is well removed from the mean of these GATE cases except if the Darwin monsoon is represented by the somewhat inappropriate 0–3 km shear (M). That differences exist between the mean GATE population and those observed herein has important implications in the use of the GATE derived vertical convective heating profiles. The observed differences in the dynamical characteristics of the convective systems imply both larger-amplitude heating and higher heating in the troposphere than in the GATE region. Recent results from the Australian Monsoon Experiment of Frank and McBride (1989) offer support that the vertical profile of convective heating in the Australian region is indeed different from the GATE. Conversely, many similarities may be noted to west African tropical squall systems as examined by Sommeria and Testud (1984), Chong et al. (1987) and Roux (1988).

(c) Propagation

During the course of preliminary flow-field evaluations, it became apparent that transition and break-season squalls translate at a speed and in a direction very similar to the 700 mb easterly jet. Without exception, the squall systems were oriented orthogonal
to the shear vector in the lower troposphere. Moreover, there is redistribution of 700 mb horizontal momentum to the base of the convective line by the descending rear-to-front flow. This feature has been noted in several of the systems for which Doppler data have been examined (e.g. Fig. 3(d) and Figs. 7(d)–(f)).

Table 1 summarizes the translation speed of 19 transition and break-period squall systems on 17 days and compares this speed with the maximum component of environmental flow orthogonal to each line. The flow maximum is located between 3.1 and 3.5 km in all instances. Table 1 reveals that the movement of most squall systems was similar to the easterly jet flow speed. Five systems translate at a rate significantly greater than the maximum wind while only three systems translate significantly slower. A departure of 3 m s\(^{-1}\) is defined as significant somewhat arbitrarily because it was consistent with deviations larger than those expected from observational errors and it is approximately the standard deviation of all storm speed departures from the environmental flow.

For the purposes of the following discussion, systems which move 3 m s\(^{-1}\) slower than the maximum environmental wind speed will be called 'non-propagating'. Systems which move 3 m s\(^{-1}\) faster than the maximum environmental wind will be called 'propagating' and systems which move at a speed within 3 m s\(^{-1}\) of the maximum environmental winds will be referred as being in a near ‘critical’ or a ‘balanced’ state.

The propagating cases which translate faster than the maximum environmental winds are near average in their shear characteristics or have a low 700 mb wind maximum, as indicated in Table 1. Some tend to have above average stability in the PBL (e.g. the discontinuous propagation case of 4 January 1988 shown in Fig. 10).

### Table 1. Comparison of Squall System Motion and Environmental Flow

<table>
<thead>
<tr>
<th>Date</th>
<th>Direction (°)</th>
<th>Speed (m s(^{-1}))</th>
<th>U 700 mb</th>
<th>Propagating speed (m s(^{-1}))</th>
<th>u 950 mb (m s(^{-1}))</th>
<th>(\Delta T_\text{c}) (degC)</th>
<th>CAPE (J kg(^{-1}))</th>
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<td>330</td>
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<td>2.9</td>
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<td>4.7</td>
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<td>618</td>
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**Average/Standard deviation:**

- All cases  
  - 283/21: 11.2/3.0: 10.8/3.9  
  - U = 0.4/3.7  
  - 1.5/1.1  
  - 3.3/2.2  
  - 1083/546
- Propagating  
  - 273/19: 13.6/1.9  
  - 9.7/1.5  
  - 3.9/0.5  
  - 1.1/1.5  
  - 5.1/2.3  
  - 1070/705
- Non-propagating  
  - 285/13: 10.4/3.4  
  - 17.3/3.3  
  - -6.9/1.6  
  - 1.3/0.8  
  - 2.9/2.0  
  - 1187/495
- Critical  
  - 285/24: 10.4/3.0  
  - 9.6/3.2  
  - 0.8/1.5  
  - 1.7/0.9  
  - 2.6/2.0  
  - 1060/532

*U* is the component of the 700 mb wind parallel to the squall direction of motion, *u* the component of the 950 mb wind parallel to the squall direction of motion and \(\Delta T_\text{c}\), the virtual temperature difference between environment and the cold pool calculated from Eq. (1) using *U* and *u*. 


CONVECTIVE SYSTEMS AROUND DARWIN

The non-propagating systems which translate slowest with respect to the maximum environmental winds are the highest shear cases with the strongest easterly jet speeds (e.g. the island thunderstorm squall and the sea-breeze generated squall of 14 January 1988 shown in Figs. 6 and 7).

The tendency for propagation or otherwise of the systems to propagate relative to the environmental flow thus seem to be partly reflected in the strength of the environmental 700 mb wind maximum. Buoyancy, as indicated by the CAPE values in Table 1, does not seem to be a factor determining the potential for propagation or non-propagation.

Although the number of cases presented herein is small, the results compare favourably with those obtained elsewhere.

Aspliden et al. (1976) showed that west African disturbance lines, in an environment characterized by a storm-relative shear profile similar to the transition-season squalls studied here, typically move at 10–20 m s⁻¹, very close to the speed of the African easterly jet (AEJ). Similarly, the GATE fast-moving squalls studied by Barnes and Sieckman (1984) that are again comparable with the squall systems studied here moved at a speed close to that of the 700 mb jet.

Squall lines studied in Venezuela by Betts et al. (1976) and Miller and Betts (1977) existed in a wind regime similar to the transition-season data of Fig. 1, i.e. easterly flow below 300 mb with westerly flow above and an easterly wind-speed maximum near 700 mb. Storms travelled westward at speeds between 10–16 m s⁻¹ comparable with or greater than the maximum easterly flow. Like the storms in Darwin, data presented by Betts et al. (1976) show those Venezuelan storms that propagated fastest relative to the environmental wind tended to have the lowest mid-level (about 700 mb) wind speed.

Observational data presented by Chong et al. (1987) and Calon et al. (1988) indicate that although the squalls of 22 and 23 June 1981 observed during the Convection Profonde Tropicale in 1981 (COPT81) moved at different speeds their respective speeds were very close to that of the AEJ along the direction of storm motion.

The two-dimensional numerical simulations presented by Nicholls et al. (1988) offer further support to the observational data presented here. The control squall simulation, in an environment with a 11.7 m s⁻¹ jet profile similar to the break-season profile in Fig. 1, moved at 11.4 m s⁻¹, i.e. at a speed close to the maximum environmental wind speed or the critical condition noted above. Squalls modelled with a strong 17 m s⁻¹ jet profile similar to that in Fig. 1, moved slower than the maximum wind speed by up to 3–6 m s⁻¹. These non-propagating systems also produced the weakest cold pools as indicated by the PBL cooling (about 1 degC). Squalls modelled with a weaker jet profile propagated relative to the environmental flow by up to 3 m s⁻¹ and produced average cooling in the PBL (3–6 degC). Buoyancy had a weak and variable effect on the system motion relative to the environment.

Modelling experiments on the COPT81 23 June 1981 squall presented by Lafore and Moncrieff (1989) indicate similar results. Their simulation U with the AEJ decreased to 10 m s⁻¹ and showed squall propagation; i.e. storm motion 2–3 m s⁻¹ faster than the maximum environmental wind. With the AEJ increased to have a 24 m s⁻¹ maximum, their simulation Z produced a non-propagating storm that moved approximately 6 m s⁻¹ slower than the environmental wind maximum.

The studies of Chong et al. (1987), Chalon et al. (1988), Roux (1988) and Lafore and Moncrieff (1989) also noted the importance of the rear-to-front flow associated with the storms. As found here, the storms were oriented perpendicular to the low-level shear with rear-inflow sloping downward to the surface and taking on the characteristics of the density current in the forward half of the storm. Mass transport to the cold pool was
estimated to come from both convective and mesoscale downdraughts. The rear-to-front flow, its association with squall-line maintenance and possible mechanisms for its generation have previously been studied by Smull and Houze (1987).

Translation speeds in excess of environmental winds are defined as propagating systems as established by Moncrieff and Miller (1976). Production of a shallow, spreading cold pool by means of diabatic cooling on the rear flank is a feature of many propagating squall systems. Rearward tilt of updraughts in the convective line and/or upper-level winds of smaller magnitude than the low-level easterly jet assure the sedimentation of precipitation in the rear flank and thereby provide negative buoyancy production by means of evaporation and melting below 4 km. From our observations, these mechanisms plus horizontal and vertical pressure gradients behind the convective lines appear to be quite effective in the redistribution downward of 700 mb momentum as part of cold-pool formation. It follows, to a first approximation, that the stagnation point at the leading edge of the cold pool has horizontal momentum very similar to the undisturbed 700 mb flow. In the extreme case, where unmixed horizontal momentum is transported to the PBL, it is necessary to include the divergent component of the spreading cold pool beneath the convective downdraughts. By mass conservation, one would expect this term to be 1–5 m s\(^{-1}\), thereby potentially explaining the speed of many propagating systems.

The results obtained cannot discriminate all the factors that are important, but are offered as observational support for the concepts presented by Rotunno et al. (1988) and Lafore and Moncrieff (1989) among others.

Figure 16(a) shows composite ‘pre-storm’ and ‘post-storm’ environmental wind profiles for the ‘critical’ cases from Table 1 obtained by compositing Darwin rawinsonde data. A schematic representation of redistribution of easterly-jet momentum and cold-pool formation is also included. Consistent with the observations, the system is assumed to translate at roughly the maximum wind speed, \(U\). Relative flow in the shallow cold-pool layer may then be defined as \(U - u\), where \(u\) is the average near-surface flow speed over the depth of the cold pool, \(h\). If the system translation is in a quasi-steady state, then an approximate balance with density-current propagation is achieved by means of the following expression (von Kármán 1940):

\[
U - u = (2g' h)^{0.5}
\]

where \(g'\) is the reduced gravity, \(g(\delta T_v/T_v)\). \(T_v\) is defined as virtual temperature or air density of the undisturbed environment and \(\delta T\), the virtual temperature difference between the environment and the cold pool. For typical values observed \((U = 10 \text{ m s}^{-1}, u = 2 \text{ m s}^{-1}, h = 500 \text{ m})\), the implied cold-pool balance is achieved with a density difference of 1.0% or 3 degC temperature difference. While surface thermograph and barograph records are quite incomplete, this magnitude of surface-temperature and pressure changes are typical of gust-front passages near Darwin. The observed squall system translation speeds of approximately \(U\) confirm the theory of Rotunno et al. (1988) and as suggested by Thorpe et al. (1982) for long-lived squall lines in the special case where \(u\) is near zero in Darwin. This is powerful evidence of long-lived squall systems maintaining a balance between low-level horizontal velocity \((U - u)\) and density-current propagation. Furthermore, the rear flank downdraught initiated redistribution of the ‘momentum’ from 700 mb to the surface layer is consistent with diabatic processes and minimizes accelerations required to establish a cold-pool equilibrium speed.

There are many factors that can account for the observed deviations in translation speed. Tilt of the updraught affects the distribution of diabatic heat sources and sinks, as noted by Dudhia et al (1987) and Nicholls et al. (1988), and this modifies horizontal pressure gradients near the surface which ultimately control production and acceleration
Figure 16. Composite pre-storm and post-storm environmental wind profiles (storm relative) and schematic depiction of the interaction between the environmental flow, the redistribution of easterly momentum and the formation of the cold pool: (a) the 'critical' or 'balanced' state in which the storm moves at the speed of the 3 km flow; (b) the 'non-propagating' case where the storm moves slower than the 3 km flow and (c) the 'propagating' case where the storm moves faster than the 3 km flow.
of the cold pool. Tilt of the updraught is affected strongly by mid-tropospheric shear profiles. Lafore and Moncrieff (1989) discuss such factors and provide an example of how the shear above the jet maximum can affect the production of the cold pool. Vertical distribution of moisture, particularly dryness at the 2-4 km level, controls potential negative buoyancy production in the descending rear-flank inflow. Finally, stability of the PBL affects the horizontal vorticity balance implied by Rotunno et al.'s theory where lifting occurs at the convective line. Stability also introduces the likelihood of gravity-wave excitation, in which case squall propagation can be controlled by complex wave/cold-pool interactions as discussed by Crook (1988) and Carbone et al. (1990b).

In this data set, squall translation speeds which deviated significantly from $U$ had potential imbalances resulting from one or more of the aforementioned causes. It is difficult to assess the impact of these factors on the examples presented at this stage. For instance, in the non-propagating case of 14 January 1988 the observed rear inflow at 15:42 LST (Fig. 7(e)) was greater than or equal to the storm motion and any environmental wind. The propagating case of Fig. 12 had only weak descending rear inflow. In addition, both the propagating and non-propagating cases had evidence for PBL cold-pool velocity maxima greater than the mean storm motion. As only very limited sampling of the storm structure was undertaken herein the following inferences will be based on the composite environmental data and information available from other observational and modelling studies.

The imbalanced non-propagating-case composite environmental wind profiles and proposed schematic flow are given in Fig. 16(b). The large environmental shear and environmental rear-to-front flow suggests erect or forward-tilted draught structure. As a result, limitations in negative buoyancy production indicative of lower bulk Richardson number conditions are proposed. In this case, the latent-heat distribution from the erect updraughts negates the cold-pool pressure rise slowing the overall motion. In addition, with precipitation falling into the updraught inflow, the disturbed inflow environment results in an overall weaker updraught with a less well developed or non-existent cold pool. The situation depicted is similar to the unicellular case of Dudhia et al. (1987) or the case of greater than optimal shear discussed by Rotunno et al. (1988). Another factor that may be important is a less well developed cold pool because of a limited potential for evaporative cooling due to environmental conditions.

The imbalanced propagating cases, represented in Fig. 16(c), suggest insufficient shear relative to cold-pool production and strongly rearward-tilted updraughts, i.e. the case discussed by Rotunno et al. where the shear is less than optimal to balance the buoyancy production from the cold pool. Here the cold pool dominates and moves faster than the wind above, producing a highly slanted system. This system is not as intense as the critical case, with the newly triggered cells being left behind the fast-moving cold pool. Precipitation and resultant evaporative cooling enhance the cold-pool intensity. Increased PBL stability may also have led to hybrid gravity-wave/density-current propagation mechanisms such as internal bores (e.g. Simpson 1987) which are not depicted in Fig. 16(c). When large imbalances of this type occur there is the potential for discretely propagating modes as discussed in section 4(e) for the 4 January 1988 squall.

6. Summary

Characteristics of the initiation, structure and motion of a limited sample of tropical convective storms observed during the summer monsoon at Darwin during 1987/88 have been identified.
This paper has shown that the monsoon flow regime is characterized by generally weak convection embedded in low-level westerly flow with strong but shallow shear concentrated in the 0–1.5 km layer. The monsoon regime convection can be associated with large CAPE values. As shown, some of the largest CAPE values encountered were from the monsoonal regime. Squall-like structures are generated in the monsoon often within large mesoscale stratiform decks.

During the break- and transition-period flow, a very different spectrum of deep convective activity is observed. The dynamical regime is dominated by the 700 mb easterly wind maximum and the associated shear in the 0–3 km layer. This paper has presented a limited selection giving examples of deep (15–19 km high) continental thunderstorms, sea-breeze initiated convective squall lines, maritime continent island thunderstorms and long-lived westward-moving continental squall lines. These latter squall lines can be very similar to mid-latitude squalls but in the mean seem to be located in a dynamical regime very similar in several respects to that experienced by non-severe squalls studied by Bluestein et al. (1987). However, the data presented indicate an overlap with severe storms and possible rotating draught structure in some cases. In addition, the Darwin systems develop in a regime with slightly larger low-level shear than found for both the non-severe and severe US squalls.

A ubiquitous feature in both the monsoon and break-period flow was the tendency to evolve squall-line structures oriented perpendicular to the low-level shear vector. Local topographical forcing can result in initial convection oriented parallel to the low-level shear vector but in the cases examined the natural evolution seems to be towards a system oriented perpendicular to the low-level shear. This evolution includes the development of descending rear inflow or middle-level downdraughts transporting momentum into a forward-propagating PBL cold pool. The subsequent interaction of this cold pool with the environment appears to determine both the longevity, structure and motion of the evolving system.

The spectrum of convective activity and the comparison with the GATE data in the context of the bulk Richardson number classification indicate similarities and significant differences. The Darwin break-season squalls seem similar to the GATE fast-moving squalls except that the GATE systems were essentially oceanic, not the continental systems studied here. The GATE slow-moving squalls and the Darwin squall with the characteristics of the Houze (1977) squall were at the low-shear end of the Darwin spectrum. In addition, the aggregate of the two GATE systems are at the low-shear, low-CAPE end of the spectrum observed at Darwin. Such differences have important implications in the context of using the GATE data as globally representative, as found by Frank and McBride (1989).

The motion of break-season continental squall lines is in the mean at a speed very close to that of the 700 mb jet. Downward redistribution of near 700 mb easterly momentum within a downdraught, to a first approximation, gives a spreading cold pool in balance with the environment and moving at the velocity near that of the 700 mb flow. However, examples presented indicated small but important differences in terms of continuous and discontinuous development of new cells at the leading edge of the squall line. It is suggested that observed small deviations from this 'balanced' or 'critical' state may result in 'propagating' and 'non-propagating' moving systems. Variation in the Froude number, the tilt of updraughts, the potential for downdraught formation and the environmental shear are considered important in achieving this balance consistent with the concepts of Rotunno et al. (1988) and Leflore and Moncrieff (1989). PBL stability where hybrid gravity-wave/gravity-current propagation mechanisms occur may also be important for 'propagating' systems.
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