Boundary Layer Development over a Tropical Island during the Maritime Continent Thunderstorm Experiment

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ABSTRACT

Data collected during the Maritime Continent Thunderstorm Experiment (MCTEX; Keenan et al. 2000) have been used to analyze boundary layer development and circulations over two almost flat, tropical islands. The two adjacent islands have a combined length of about 170 km from east to west and 70 km from north to south. Intense thunderstorms formed over these islands every day of the field campaign. The boundary layer depth, temperature, and circulation over the island have been measured over the full diurnal cycle using a multiple radar analysis combined with surface and radiosonde measurements. On average, the island boundary layer depth reaches 1.5 km by early to midafternoon coinciding with the development of the deep convection. Thus, the island boundary layer is significantly deeper than the typical tropical oceanic boundary layer. In the midafternoon, thunderstorm outflows and their associated cold pool stabilize the lower boundary layer, suppressing late convection. This is followed by a period of partial boundary layer recovery for 1–2 h. After sunset, cooling leads to a deepening ground-based inversion below a residual mixed layer. Near the island center, the residual mixed layer of island-modified air is replaced by air of oceanic origin by about 2300 LST (local standard time) that then persists until sunrise the next day. The advection of boundary layer air of oceanic origin over the islands every evening resets the boundary layer development cycle. It is shown that much of the variation in the diurnal temperature profile is a result of thunderstorm activity, radiative processes, and the advection of island and oceanic boundary layer air.

1. Introduction

Data describing the evolution of the boundary layer over a tropical island were collected during the Maritime Continent Thunderstorm Experiment (MCTEX; Keenan et al. 2000). It was conducted from 10 November to 10 December 1995, during the buildup to the Australian monsoon on Bathurst and Melville Islands (Northern Territory, Australia). During this period, island thunderstorms are a dominant form of convection in the region with thunderstorms almost every afternoon. The storms are typically very deep with tops as high as 20 km, reach maximum intensity in midafternoon, and have largely decayed by sunset. The Island Thunderstorm Experiment (ITEX; Keenan et al. 1989) was the first experiment to investigate these storms in detail. MCTEX built on this earlier experiment and included much more detailed and extensive measurements of the boundary layer circulation and development.

MCTEX focused on convection occurring on tropical islands during the build up to the monsoon. This convection is strongly influenced by coastal and diurnal
effects. An objective of MCTEX, was to examine the life cycle of convection from the initiation process through to the mesoscale organization of the deep precipitating system. Achieving this objective depends on an understanding of the boundary layer development and associated circulation.

The diurnal island-based nature of the convection suggests that convergence in the island boundary layer is the main cause of thunderstorm initiation. A similar situation is found on the Florida peninsula, where colliding sea breezes generate convergence during the day, with the subsequent initiation of thunderstorms (e.g., Wakimoto and Atkins 1994; Kingsmill 1995). Enhanced surface convergence can result from a number of boundary layer features including gust fronts, horizontal convective rolls, sea breeze fronts, and topographically induced convergence lines (e.g., Kingsmill 1995).

Documentation of the island boundary layer growth, circulations and convergence is a critical component for the understanding of the life cycle of these strongly diurnally modulated convective systems. For example, during MCTEX, the earliest convection was often observed to occur on the east coast of Melville Island, where the east coast sea breeze convergence was enhanced (and the front steepened, e.g., Liu and Moncrieff 1996) by an opposing low-level westerly flow that was itself related to the heat low over the continent to the south. Later thunderstorm activity was influenced by gust front and sea breeze interactions (e.g., Carbone et al. 2000). On a few occasions, the convection formed when the sea breezes from the north and south coasts collided. Early convection was usually relatively suppressed on these days and the deep afternoon storms tended to be less intense. In contrast to convection during the Convection and Precipitation/Electrification experiment, convection was not organized in cloud streets. Instead, the convection was organized by interactions of the environmental wind, sea breeze, and gust fronts.

This is the first time the boundary layer structure has been documented over islands in the Maritime Continent through the full diurnal cycle. The diurnally modulated storms that form over these islands are among the deepest and most intense in the world and it is the boundary layer structure that ultimately controls the development of them and their diurnal modulation (Carbone et al. 2000). The deployment of a network of boundary layer wind profilers, Doppler weather radar, regular radiosonde launches, and an extensive automatic weather station (AWS) network, has provided the opportunity to study the development and circulations within a tropical island boundary layer. Advection of air across the island and from over the ocean will be shown to impact directly on the diurnal boundary layer structure.

This study focuses on the mean boundary layer evolution over the islands during MCTEX. The next section will outline the observation network and data collection and quality issues. This will be followed in section 3 by a description of the large-scale environment. The principal boundary layer development phases will be outlined in section 4. These phases are 1) the daytime boundary layer growth, 2) stabilization of the boundary layer resulting from thunderstorm activity, and 3) the development of a nocturnal boundary layer with a residual mixed layer of oceanic origin above. The variations in the principal boundary layer development phases are outlined by examining the diurnal cycle on 2 days during the experiment that occurred in different thunderstorm environments (section 5). Finally the conclusions and a summary of this work are presented in section 6.

2. The experiment and data

Bathurst and Melville Islands are relatively flat islands, with the highest point only 170 m above sea level. The east to west length of the islands is about 170 km and the north to south dimension is about 70 km (Fig. 1). A ridge runs in an east to west direction on the southern part of the islands. A shallow channel separates Bathurst and Melville Islands. Since only a narrow channel separates the two islands, and the majority of the instrumentation is on Melville Island, the two islands are treated as a single island. The vegetation consists of open forest and grassland. As a result of the relatively flat topography and small horizontal scale of the islands, coastal influences such as sea breeze and the advection of oceanic air affect the boundary layer over most of the island.

Keenan et al. (1994b), describe the full observation network and component objectives of MCTEX. Figure 1 shows the deployment of instrumentation during MCTEX. The primary instruments of interest in this study are the radar network, radiosonde at Maxwell’s Creek, and AWS stations. Three 920-MHz (33-cm wavelength) boundary layer wind profiler radars and a 5-cm wavelength Doppler weather radar were used to examine the circulations and structure of the island boundary layer. The wind profiling radars were positioned in a northwest to southeast line across Melville Island and together with the C-Pol Doppler radar at Nguiu (Keenan et al. 1998a) a triangle of wind observations was obtained. Radio acoustic sounding systems (RASS; e.g., May et al. 1990) operated with the wind profilers at Maxwell’s Creek and Pickertaramoor deriving virtual temperature profiles from about 100 m to about 2000 m above ground level. The RASS mode operated every 15 min for a duration of 3 min. The 3-min duration generated about five virtual temperature profile measurements. The wind profilers were configured to make radial velocity measurements in three beam directions, vertical and 15° off vertical to the east and north. Two height resolutions were used. A 100-m resolution mode produced measurements from 100 m to about 2000 m, and a 300-m resolution mode provided observations up to about 20 km during precipitation. The radar cycled through the two height modes and three
beam directions about every 6 min. The main causes of gaps in the data record were equipment failures due to nearby lightning strikes. Only RASS data at Maxwell’s Creek were used, as a lightning strike at Pickertaramoor led to the failure of an acoustic amplifier channel and a subsequent reduction in maximum observable height to less than 300 m.

The profilers used standard National Oceanic and Atmospheric Administration Aeronomy Laboratory signal processing for radial velocity estimation (as described in Strauch et al. 1984). Outliers in the radial wind measurements were removed using consensus averaging (Fischler and Bolles 1981). Analysis of the wind profiler data showed no signal interference resulting from birds or aircraft. However, radial velocity estimates at Maxwell’s Creek were severely affected by ground clutter up to a height of about 1 km necessitating the reanalysis of the spectral data. Comparisons with wind measurements from radiosondes launched nearby showed biases of 2–3 m s\(^{-1}\), comparable to the low-level wind speeds.

A clutter removal algorithm that consisted of interpolation over the spectral spike at zero frequency and half-plane subtraction (Passarelli et al. 1981) was applied to the data. This process consists of removing the part of the spectrum that is symmetric about zero velocity and moments are calculated on the residual. The clutter removal was only applied if the maximum in the velocity spectrum was at zero. Using this technique, the bias between measurements made from approximately 100 radiosonde ascents launched about 500 m from the profiler site and the wind profiler derived velocities was reduced from 2 to 3 m s\(^{-1}\) below 900 m to less than 0.2 m s\(^{-1}\) with an rms difference in wind speed of 1.4 m s\(^{-1}\). This is comparable to the results of other comparisons (e.g., May 1993; Riddle et al. 1996).

The profilers were also used for reflectivity measurements using the detected signal to noise ratio multiplied by the range squared. However, these systems are not calibrated and the clear air reflectivity shown in this paper is in relative units only. It has been documented that there often exists a clear enhancement in reflectivity at the top of the mixed layer because of the large vertical gradients in humidity and potential temperature (Angevine et al. 1994). This along with radiosonde data is used to track the growth of the mixed layer during the morning.

During MCTEX the C-Pol Doppler (C-band polarimetric) radar was located at Nguiu on Bathurst Island. Radial velocity and reflectivity (along with polarimetric variables) were measured at 300-m-range intervals. The C-Pol Doppler radar is described in detail by Keenan et al. (1998a). A velocity–azimuth display (VAD) analysis (e.g., Browning and Wexler 1968) was carried out on the radial velocity measurements using data at ranges between 3 and 9 km giving profiles of horizontal wind over the radar. The range used for the VAD was limited to ensure the wind profiles were representative of the local wind near the radar. During MCTEX, the Doppler radar was primarily used in a weather-scanning mode, with scan elevation angles generally limited to less than 5°. As result, VAD wind profiles typically extended to a height of 700 m.

Frequent radiosonde ascents, AWS data, two all-sky cameras, and a small remote controlled aircraft, the aero-
sonde (Holland et al. 1992), were used in conjunction with the radar network to document the boundary layer processes. Automatic weather stations were positioned at 15 sites across the island recording horizontal winds, temperature, pressure, and humidity at 1-min intervals (Keenan et al. 1998b). A Vaisala Marwin MW 12 sounding system was operated at Maxwell’s Creek, providing measurements from near the surface to about 20 km. Launches were carried out every 2 h on 11 intensive observation periods (20, 22, 23, 25, 26, 28, 29, and 30 November and 1, 2, and 4 December 1995). On the other days there were some soundings at requested times. In
total 98 soundings were carried out during the experiment. All sky cameras were located at Pirlangimpi and Maxwell’s Creek. These cameras photographed the sky using a fish-eye lens at 15-min intervals from sunrise to sunset. The cameras operated during the period 19 November–4 December 1995. The fish-eye lens showed a 360° view of the sky down to the horizon. However, identification of individual clouds is limited to a much smaller horizontal radius (about 2 km). Aerosondes (small remote controlled aircraft with a wing span of 2.9 m) flew a series of missions investigating sea breeze and gust front circulations as well as general atmospheric profiles during the day and night. Nine flights generating suitable data records for analysis of boundary layer development. These were on 22, 23, 26, 27, 28, 30 November and 1, 2, 4 December 1995. The aerosonde measured the wind speed and direction at 200–1200-m intervals, as well as air pressure, temperature, humidity, and water vapor mixing ratio at 200-m intervals along the flight path.

3. Mean circulation

During the buildup to the Australian summer monsoon, extensive convective activity tied to islands occurs throughout the region referred to by Ramage (1968) as the “Maritime Continent.” The term “buildup” is used to represent the time between the end of the dry season in the Northern Territory, Australia, and the first burst of the monsoon (Bureau of Meteorology 1995). Typically this buildup period occurs during November and early December.

Composite analyses of the wind profiler and AWS horizontal wind and virtual temperature profiles have been prepared using the method described by May and Wilezak (1993). These composites are constructed by averaging wind estimates (themselves consisting of 30-min averages) and RASS data for a particular height and local time collected during the whole experiment. These composites were constructed with a time resolution of 30 min. Radiosonde profiles during MCTEX were obtained at a 2-h interval during intensive observation periods and longer intervals on other days. Composites of radiosonde data were therefore created by first placing each profile during the experiment into a 2-h bin beginning at 0000 LST. The corresponding bins for each day were then averaged to produce a composite data set for the entire experiment.

These mean daily time–height composites have been constructed for the wind profiler site at Pirlangimpi (Fig. 2a) and for the Nguiu Doppler radar (Fig. 2b). AWS winds have been used for the surface observation. These composites verify the persistence of the easterly trade wind flow above about 1 km. A transition to westerly flow occurs above 4 km as is typically observed by the wind profiler over this region at this time of year (not shown in these plots). Below about 1 km, the mean wind rotates from an easterly flow to a westerly flow below about 500 m. The westerlies were associated with a heat low over northern Australia; however, the low-level flow was quite variable. Two dominant low-level (below 500 m) wind regimes occurred during MCTEX, an east to northeasterly flow (25 November–2 December 1995) and a west to southwesterly flow (19–24 November and 3, 4, 6 December 1995). These two regimes impacted the location of the first convection. Favorable sites were
where the sea breeze opposed the low-level flow (e.g., Carbone et al. 2000).

The mean pressure weighted wind shear during MCTEX was calculated from a composite of the 30-min consensus averaged horizontal winds. A 2-h averaging interval was employed (i.e., the day has been broken into 12 2-h blocks for averaging). The unweighted shear is the difference in mean wind speed between each 100-m wind profiler measurement level. The pressure weighted shear is the unweighted shear at each level weighted by the standard pressure at that level. The standard atmosphere (COESA 1962) is used for the pressure weighting. The mean 0–2-km shear at 2-h intervals was calculated by averaging over height. During the day [1130 LST (local standard time)] the mean pressure weighted wind shear over the lowest 2 km was $2.6 \times 10^{-3} \text{ s}^{-1}$, $1.4 \times 10^{-3} \text{ s}^{-1}$, and $1.9 \times 10^{-3} \text{ s}^{-1}$ for Pirlangimpi, Maxwell’s Creek, and Pickertaramoor, respectively. These values are similar to that obtained by Keenan et al. (1994a) during ITEX at Pirlangimpi of $2.5 \times 10^{-3} \text{ s}^{-1}$ during the day. Thus both ITEX and MCTEX were conducted in similar wind and shear environments. Since the shear is also calculated over the layer influenced by the sea breeze, the largest values of wind shear occur at the coast (Pirlangimpi and Pickertaramoor). The shear estimates suggest a stronger sea breeze influence at Pirlangimpi, which is closest to the coast and also closer to sea level than Pickertaramoor. The mean winds above 1 km are easterly, while the winds near the surface at Pirlangimpi show a west to northerly sea breeze flow. At Pickertaramoor the sea breeze was usually weaker and from the south. In contrast, the mean winds at Maxwell’s Creek (not shown here) have a greater variability during the day with no clear preference in direction.

4. The diurnal cycle

During the experiment, thunderstorms occurred over the islands every day, although the location and intensity of the thunderstorms varied. The locations of thunderstorms and low to midlevel cloud during MCTEX were influenced by the environmental winds. Once thunderstorm activity ceased, high-level cloud persisted over much of the island until sunset. Figure 3 shows the cloud cover in hourly intervals for the 27 November 1995 over Maxwell’s Creek. The cloud development cycle shown in this figure was typical of that which occurred during MCTEX. The morning generally began cloud free. By 0930 LST, small cumulus appeared near the top of the boundary layer. The cloud amount increased and there was evidence of some penetration through the top of the boundary layer by about 1200 LST. This penetration was shown as enhanced radar reflectivity above the top of the boundary layer.

As clouds start to penetrate the boundary layer, the wind profiler reflectivity shows an enhancement in vertically elongated bands. This for example is shown in the mean time–height cross section for Maxwell’s Creek on the 27 November 1995, with the shaded region highlighting relatively large reflectivity (Fig. 4). The cloud photography (Fig. 3) shows small cumulus by 0930 LST. The cloud amount and thickness increased until the first heavy rainfall at 1537 LST. From about 0900 LST to about 1030 LST the radar reflectivity and the boundary layer height algorithm (solid line) show the developing boundary layer. The algorithm (which simply determines the height of maximum reflectivity) fails when the ground clutter signal (horizontal band of enhanced reflectivity between 700 and 800 m) is stronger than the reflectivity near the top of the boundary layer. The long vertically elongated enhancements represent clouds passing over the radar. A particularly large enhancement is seen just after 1000 LST. The cloud imagery shows more opaque (hence thicker) clouds at this time. After about 1030 LST, the boundary layer top is poorly defined by the reflectivity and algorithm, with vertically elongated bands of enhanced reflectivity that are discontinuous over time. The vertically elongated reflectivity enhancement prior to precipitation is a result of turbulence within clouds. The first rain at the site occurred after 1430 LST and is verified by raindrops on the all-sky camera dome in the 1537 LST image. The long bands of enhanced reflectivity at this time are the result of precipitation. During the period where clouds penetrate the top of the boundary layer, the actual top of the boundary layer is difficult to define from the profiler data alone (e.g., Angevine et al. 1994). Since the automated height detection often failed the boundary layer top was primarily determined by visual examination of the reflectivity data.

Cumulonimbus affected the area with associated rain and thunderstorm outflows primarily between about 1400 and 1600 LST. There was extensive cirrus overcast even on days where there was no rain at the site. Generally the sky remained mostly cloud covered until sunset. This cycle was extremely regular during the experiment. The second camera located at the coast showed a similar cycle, except that the sky cleared behind the sea breeze front as it advanced. This clearing is also seen in satellite imagery.

The thermal structure of the island boundary layer during MCTEX over the diurnal cycle is described in Figs. 5 and 6 at Maxwell’s Creek. Figures 5a and 5b show mean virtual potential temperature and water vapor mixing ratio profiles, respectively. The profiles consist of data averaged over all of the radiosonde profiles collected during MCTEX and composited into a mean day with 2-h resolution. If more than four profiles were used in an average a standard deviation was calculated and shown as a horizontal line about the mean. The time displayed in Fig. 5 is the centre of the bin so that 0800 LST represents the mean sounding from 0700 LST to 0900 LST. Figure 6 shows the RASS derived virtual temperature time–height composite for Maxwell’s Creek.
The diurnal cycle can be broken into three distinct phases based on cloud imagery and thermal and moisture data. The first phase from just after sunrise (0613 LST) until the passage of storms is characterized by boundary layer growth; the second is marked by the storm passage and associated cold outflows; and the third is the nighttime phase with the ground-based inversion and a fossil mixed layer above it. These three phases will now be discussed in turn.
Fig. 4. Wind profiler derived range corrected signal-to-noise ratio (SNR) for 27 Nov 1995. The shaded region shows the highest reflectivity levels. The solid line indicates the maximum range corrected SNR.

Fig. 5. Mean island boundary layer radiosonde profiles at Maxwell’s Creek of (a) virtual potential temperature and (b) water vapor mixing ratio. Each vertical dashed line represents a virtual potential temperature of 32°C and a mixing ratio of 15 g kg⁻¹. The distance between vertical dashed lines is 10°C for the virtual potential temperature profiles and 10 g kg⁻¹ for the water vapor mixing ratio profiles. The std dev about the mean is shown if more than four soundings were available for the mean profile.
**a. Boundary layer growth phase**

The growth of the mixed layer over the island centre is a key factor controlling the potential for deep convection. Surface temperatures started to increase at each of the island AWS sites at about 0630 LST every day. This was a common characteristic at each site. The only exception during the period of wind profiler observation, occurred on 1 December when a cold outflow from thunderstorms over the mainland of northern Australia passed over the islands. The cold air mass persisted on the island for about 3 h from 0430 to 0730 LST. The surface temperatures at each of the wind profiler sites only began to rise at about 0730 LST. During the experiment, the AWS derived water vapour mixing ratio at each of the sites also increased after 0630 LST in response to the surface heating, but reached a peak value by about 0800 LST.

The morning mean radiosonde derived virtual potential temperature sounding (0800 LST) at Maxwell’s Creek shows a surface-based inversion up to about 300 m (Fig. 5a). The largest water vapor mixing ratios are also in this layer (Fig. 5b). Above this is a mixed layer from 300 m to about 650 m capped by a temperature inversion. The water vapor mixing ratio profile through this layer is close to constant at about 15 g kg⁻¹. Although the virtual potential temperature sounding shows a weakly stable layer, the constant mixing ratio suggests a mixed layer. Hence the term “mixed layer” is used to describe this layer. The evolution of the morning virtual potential temperature profile with a higher temporal resolution is shown in Fig. 6. After 0600 LST, the temperature in the layer from the surface to 200 m increases. However, from 0600 to 0800 LST, the temperature in this layer is cooler near the surface than above, indicating the persistence of a layer of cooler air close to the surface. The mean early morning AWS measured temperatures near the surface at Maxwell’s Creek were typically lower by about 0.5°C–2°C than surrounding areas. Maxwell’s Creek is located in a basin 20 m above sea level. The topography rises to 80 m to the west, south, and east. Therefore, cold air generated by nocturnal cooling and drainage flows persisted in this area for a longer period than at other sites, especially those sites that are on the ridge to the south. The mean surface winds for the experiment at 0800 LST showed a south-easterly flow off the topography to the south showing that drainage flows are still in effect (Fig. 7). At levels above the surface, westerly winds advect island boundary layer air that developed on the higher topography to the west over the cold pool at Maxwell’s Creek forming the observed temperature profile.

Figure 8 shows time–height cross sections of reflectivity that have been composited in the same way as the wind data in Fig. 2. An enhancement in reflectivity is usually seen near the top of the boundary layer because of the large gradient in temperature and humidity (e.g., Angevine et al. 1994). The dashed line in Fig. 8 shows the time and height of this maximum reflectivity indicating the depth of the boundary layer as a function of time. The wind profiler estimates of boundary layer depth at Maxwell’s Creek show good agreement with the radiosonde estimates of the boundary layer top indicated by the crosses at 0800, 1000, and 1200 LST (Fig. 8a). The top of the boundary layer, marked by the crosses, was determined as the height in the mean virtual potential temperature and water vapor mixing ratio radiosonde profiles (Fig. 5) with the maximum temperature and moisture gradient. A similar boundary layer development was observed at Pirlangimpi (Fig. 8b) and Pickertaramoor (Fig. 8c). In the latter two diagrams, the top of the residual mixed layer can be seen from 0000 to 0600 LST but periods before 0800 LST were edited out of Fig. 8a because of ground clutter contamination. Surface heating led to the erosion of the surface inversion by 0800 LST. The boundary layer continues to grow and surface temperatures increased until about 1200 LST. At this stage the boundary layer had reached its maximum depth of about 1.5 km at Maxwell’s Creek, 1.4 km at Pirlangimpi, and 1.2 km at Pickertaramoor. The depth of the boundary layer was recorded for each day that showed a clear peak in the 1200–1400 LST reflectivity profile. This showed a variation in maximum boundary layer depth from 1 to 1.6 km at Maxwell’s Creek, 0.9 km to 1.7 km at Pirlangimpi, and 0.7 km to 1.7 km at Pickertaramoor.

The depth of the boundary layer was primarily influenced by rain and thunderstorm outflows near Maxwell’s Creek. Since surface temperatures began to increase at
Fig. 7. Mean AWS surface winds, virtual temperature (°C), and mixing ratio in brackets (g kg⁻¹) during MCTEX. A full feather represents a wind speed of 2.5 m s⁻¹. A featherless barb represents a wind speed greater than 0 m s⁻¹ but less than 2.5 m s⁻¹.

0630 LST, the duration over which the boundary layer developed was limited by the timing of thunderstorms. At Pickertaramoor and Pirlangimpi, the wind direction was also a determining factor. The timing of thunderstorms was recorded by the observed cooling in the radar derived virtual temperature profiles associated with gust fronts, and by enhancements in reflectivity and downward velocity associated with rain.

At Maxwell’s Creek, when thunderstorms occurred after 1330 LST, the final boundary layer depth was close to 2 km. When thunderstorms occurred prior to this time, the depth was closer to 1.5 km. There is only a small variation, as the initial boundary layer deepening phase beginning at 0800 LST is very rapid, with a depth of close to 1.5 km being reached by 1100 LST before thunderstorms affected the area. The additional growth occurs more slowly after this time. On 1 December when surface heating was suppressed until 0730 LST, the boundary layer reached 2 km as thunderstorms did not affect the area until after 1500 LST. At Pirlangimpi and Pickertaramoor, the maximum boundary layer depth tended to be closer to 1 km on days when the primary wind direction was from the closest shoreline. On days where the air passed over a longer stretch of land the boundary layer depth was closer to 1.5–2 km.

During the evening the layer above 100 m is decoupled from the surface by the nocturnal temperature inversion. During this time, the mean winds for Pirlangimpi show an easterly drainage flow in the surface layer, with stronger westerlies above. In the morning as the temperature inversion is eroded and the daytime boundary layer grows, the winds in the lowest 1 km weaken as a result of increased turbulence and surface friction (Fig. 2a). The weaker winds within the growing boundary layer often lead to a more easterly flow (or weaker westerly flow) within the boundary layer as the relatively stronger easterly flow above 1 km is mixed downward. The effect of this process can be seen at about 0900 LST between about 400 m and 1 km for Pirlangimpi (Fig. 2a). The shaded region represents winds with an easterly component. Thunderstorm down-drafts also transport momentum but this process is only important in the afternoon.

The major perturbation of the flow about the mean during the boundary layer growth phase is associated with the development of the sea breeze circulation. Figure 7 shows maps of the observed surface temperatures, water vapor mixing ratios, and winds across the island for various times. The surface temperatures and water vapor mixing ratios across the island increased by about 4° to 5°C between 0600 and 0800 LST. By 1000 LST the first sea breeze was evident on the east coast of Melville Island. The surface temperatures continued to increase through the morning, but the water vapor mixing ratios at three of the northern coastal sites, Millikapiti (MILL), Andranangoo Creek (ANDR), and Yunanti Bay
(YUNA), decreased. By 1200 LST most of the sites on the northern coastline of Melville and Bathurst Islands have shown a decrease in water vapor mixing ratio. Between 0800 and 1200 LST each of the sites across the island had shown a decrease of between 1.2 and 2 g kg$^{-1}$ from peak values of around 21 g kg$^{-1}$. The decrease in mixing ratio is typical at each of the sites during the morning and is probably a result of increased boundary layer mixing depth. This can be verified in the mean radiosonde derived water vapor mixing ratio profiles (Fig. 5b). At 1000 LST there is an initial increase in the water vapor mixing ratio from the surface...
to about 500 m, associated with increased evapotranspiration. After 10 LST the boundary layer depth increases rapidly and the mixing ratio in the surface to 500-m layer decreases. The decrease in mixing ratio is a result of the entrainment of drier air from above the boundary layer and the supply of moisture from evapotranspiration or advection being insufficient to maintain constant values (e.g., Andre et al. 1978).

By 1200–1400 LST the maximum temperatures and sea breeze strengths have been reached on all coastlines. The mean 1200 LST radiosonde profile at Maxwell’s Creek showed a relatively constant virtual potential temperature of about 32°C up to 1.2 km (Fig. 5a). The water vapor mixing ratio at this time is about 15 g kg⁻¹. The height coverage is less in the wind profiler derived virtual temperature composite; however, the composite shows a decrease in virtual potential temperature of about 3°C from the surface to 200 m and a relatively constant profile above.

The sea breeze inflow by early afternoon was about 1 km deep at the coastal sites, close to the depth of the island boundary layer. A northwesterly flow is seen in the composite for Pirlangimpi between 0900 and 1500 LST (Fig. 2a). The onset of the sea breeze at Pirlangimpi was difficult to identify when the mean environmental flow was an onshore flow. During these events, water vapor mixing ratio and temperature changes could often not be associated with a distinct sea breeze onset. The clearest sea breeze modification of the boundary layer occurred when the wind changed from an across-island flow (e.g., westerly) to a northwesterly flow. A southerly sea breeze is shown at Nguiu between 0900 and 1500 LST in the wind composite (Fig. 2b). There is a clear sea breeze signal, as the VAD analysis represents an average over a disk of 9-km radius that is centered on the east to west oriented shoreline. The mean inflow depth is about 1 km.

The sea breeze was often not detected in the wind profiler winds at Pickertaramoor most likely due to the site being on top of the ridge on the southern part of the island. Figure 9 shows the temperature variation in a north–south direction across the island for an aerosonde flight on the 2 December. This flight was chosen as it covered the full north to south extent of Melville Island. Most other flights focused on height profiles at Pickertaramoor. Radar imagery showed that most of this flight occurred through air that was not affected by thunderstorms. The only exception to this is between 1300 and 1400 LST, when the aerosonde is close to a storm at about 11.7° south. The temperature gradient decreases near and to the north of Pickertaramoor. This, in conjunction with Pickertaramoor being near the top of an east–west oriented ridge, will lead to a weaker sea breeze near Pickertaramoor. The mean meridional wind vectors measured by the aerosonde and averaged in 0.05° latitude bands are shown near the bottom of Fig. 9. These vectors show the sea breeze from the north and south coasts stalling north of Pickertaramoor. The north coast seabreeze has a much greater penetration due to the relatively flat topography north of the ridge and as a result of the northerly environmental wind. Note that these data also indicate significant modification of the sea breeze air as it propagates over the land leading to the rather gradual gradients in temperature across the island.

Maxwell’s Creek is near the center of the island and the sea breeze was rarely observed to penetrate this far. Surface temperatures near the center of the islands were cooler than surface temperatures closer to the coast throughout most of the day (Fig. 7). The cooler inland surface temperatures should reduce the inland penetration of the sea breeze and this may be the cause of the relatively low wind speeds at sites such as Maxwell’s Creek (MAXW), Three Ways (TWAY), and GOOS.

The large-scale boundary layer convergence over the western part of Melville Island was calculated over a triangle encompassing Pirlangimpi, Nguiu, and Maxwell’s Creek, using a technique described by Doswell and Caracena (1988). The convergence is associated with both the local (gust front and sea breeze) and environmental circulations. The technique uses three non-collinear measurements of the orthogonal wind vector and assumes a linear variation between the observations. While this assumption is clearly broken with sea breeze fronts between the radar sites, these measurements do give an indication of the mean convergence into the center of the island. A mean time–height composite of the convergence during the day over Melville Island is shown in Fig. 10. This showed a daytime island boundary layer (over the western part of Melville Island) dominated by convergent flow associated with local and
large-scale environmental circulations. The maximum convergence occurred in the early afternoon (about 1400 LST) between heights of 200 and 300 m above ground level. It persisted for about 2 h prior to the onset of the first significant thunderstorms in this area.

The standard error of the mean (standard deviation divided by the square root of the number of observations) for these convergence estimates is shown in the middle panel of Fig. 10. The units for the convergence and uncertainty estimates are $10^{-4} \text{ m s}^{-1}$. The uncertainty primarily represents the daily variability in the circulations. The minimum error in the estimates occurred at about 1200 LST when the sea breeze on all coasts had reached its maximum strength. Relatively large uncertainties occurred in the estimates before 1000 LST and after about 1400 LST. After 1400 LST, the uncertainty increases as thunderstorms begin to affect the circulations in the area and there were significant day-to-day variations of the preferred storm locations. Larger uncertainties before 1000 LST and after 1400
LST indicate a larger variability in the wind strength and direction prior to the sea breeze and once thunderstorm outflows affect the island circulation.

By integrating the large-scale convergence over the lowest 700 m a mean vertical velocity field can be deduced. This is shown in the bottom panel of Fig. 10. The vertical velocity increases from zero at the surface to an upward velocity of about 0.12 m s$^{-1}$. The peak upward velocities occurred between 1200 and 1400 LST. It is likely that these maxima in convergence and vertical velocity are related to the sea breeze strength at Pirlangimpi and Nguiu. Once thunderstorms formed, cool air from thunderstorm outflows was introduced at the surface, stabilizing the boundary layer and generating a divergent flow across the island. This is seen with the divergence values after 1500 LST and the downward velocities.

b. Afternoon boundary layer stabilization

The detailed storm structure is beyond the scope of this paper, but the storms have a profound effect on the boundary layer structure. An important process on the island scale is the spreading of a layer of cooler, dryer air about 500 m deep associated with thunderstorm downdrafts and outflows. The maximum cooling occurs below about 300 m associated with gust front passage and cold pool formation (Figs. 5 and 6). There is a relatively larger variability in the radiosonde virtual potential temperature profile near the surface in the 1400, 1600, and 1800 LST profiles than other times as a result of different onset times of thunderstorm activity and the arrival of gust fronts at Maxwell’s Creek. As a result, the mean 1600 and 1800 LST virtual potential temperature profiles near the surface are not representative of an instantaneous profile, but of a mean profile influenced by the timing of thunderstorm outflows. By 1600 LST, temperatures in the center dropped by an average of $3^\circ$–$4^\circ$C (Fig. 7). May (1999) showed a typical temperature decrease of $6^\circ$C associated with gust front passage and an outflow depth of 500 m or more on a limited sample high-resolution observations. These were cases where gust fronts passed directly over the wind profiler. The smaller temperature drops of $3^\circ$–$4^\circ$C shown here are due to this value being an average across the island and over a number of days. The gust fronts are modified as they pass over the warm island surface, the largest drops in temperature being near the source of the gust front. The outflows extend over the whole island domain, well beyond the areas where the thunderstorm outflows were generated. The stable outflow spreads underneath the original boundary layer air leading to stabilization of the boundary layer. By 1800 LST, offshore winds were observed on the southeast coast of Melville Island. Despite extensive cirrus overcast (Fig. 3) there is often a partial recovery of the boundary layer prior to sunset.

c. Nocturnal boundary layer development

After sunset (1856 LST for this latitude, longitude, and time of year) a surface-based temperature inversion forms (Figs. 5 and 6). The mean evening radiosonde profiles over the experiment (Fig. 5) show a weakly mixed layer above the surface. The mean profiles show a 500 m decrease in the depth of this weakly mixed layer at Maxwell’s Creek over a period of about 4 h from 1800 to 2200 LST. This shallower weakly mixed layer between about 200 and 700 m above ground level then persists throughout the evening. Evidence for this mixed layer persisting throughout the evening is also found in the mean RASS estimates of the diurnal variation of virtual temperature at Maxwell’s Creek (Fig. 6). The shaded region shows the layer in which the virtual potential gradient is less than 0.25°C. The small virtual potential temperature gradients occur in the daytime boundary, layer but also in a shallower layer during the evening and morning. During the nocturnal period, the virtual potential temperature variation over time in this layer is about 1°C. This layer develops above the nocturnal surface inversion, which reaches a maximum depth of about 200–300 m by midnight.

Measurements of the oceanic boundary layer depth were not made during MCTEX, however the depth of the typical oceanic boundary layer in the Tropics is about 500 m (e.g., Betts and Ridgway 1989). The variation of SST measured by the R/V Franklin south of Bathurst Island over the diurnal cycle was about 0.5°C and the diurnal variation of the air temperature was about 2°C.

It is hypothesised that two processes are involved in the appearance of the mixed layer above the nocturnal surface inversion. The nocturnal mixed layer is initially a residual mixed layer consisting of the remnants of the daytime island boundary layer. Later, a mixed layer of oceanic origin is advected over the island and across the ground-based inversion thus replacing the island modified residual mixed layer.

The mean winds for Maxwell’s Creek were northerly winds at about 2 m s$^{-1}$ after about 1700 LST. Northerly and northwesterly winds persist through most of the evening. Figures 2a and 2b for Pirlangimpi and Nguiu, respectively, show that the northerly and northwesterly winds during the evening occur over most of the island. The northern coastline is about 20 km from Maxwell’s Creek. Air of oceanic origin takes about 3 h to reach Maxwell’s Creek given this typical wind speed profile. The northerly component above about 900 m is weaker, implying a longer period for oceanic air at this level to reach Maxwell’s Creek. This suggests that if an oceanic mixed layer is advected across the island, it should be detected at Maxwell’s creek by about 2000 LST, although the air of oceanic origin above 900 m should arrive somewhat later.

There is evidence for the elevated mixed layer of oceanic origin on both the north and south of the island.
The reflectivity time–height cross sections for Pirlangimpi and Pickertaramoor (Figs. 8b and 8c) show an enhancement at about 300–700 m above sea level from 0000 LST to about 0600 LST. The Pickertaramoor site was 107 m above sea level so this should be taken into account when comparing the sites as the composite takes ground level as its zero height.

The residual layer comprising the remnants of the daytime island boundary layer air was first replaced by oceanic boundary layer air in a layer from about 300–700 m. Later, the stable air from above the oceanic mixed layer arrives at Maxwell’s Creek completely replacing the island residual mixed layer. The presence of the residual oceanic mixed layer is important as it essentially “resets” the development cycle every night.

5. A comparison of boundary layer development on two individual days

Examination of boundary layer development on two different days can further illustrate the processes important in modulating the boundary layer development cycle. Development on 23 November 1995 occurred in a predominantly southwesterly wind regime. Initial convection developed on the northeastern part of the island, and was primarily confined to the northern part of the island as the system moved toward the west. In contrast, on the 26 November 1995, development occurred in a northeasterly wind regime. Initial convection was on the southwestern part of Melville Island, where the south coast sea breeze converged with the northeasterly environmental flow. The system remained to the south of Maxwell’s creek and moved off the west coast of Bathurst Island by about 1800 LST. Convective activity was limited to only a small part of the island.

The morning boundary layer development for both cases was typical of that observed during the experiment (as described in the previous sections). The surface temperature and water vapor mixing ratio began to rise at about 0630 LST (Figs. 11a,b). By 0800 LST the water vapor mixing ratio had reached a maximum. At 0800 LST the boundary layer rapidly deepened reaching a maximum depth of about 1.5 km before the onset of thunderstorm activity. During the same time, the mixing ratio decreased in response to the increased mixing depth.

On 23 November 1995 there were two rain events, one at 1330 LST and the other at 1700 LST. At 1330 LST, the water vapor mixing ratio at the surface increased in response to precipitation (Fig. 11a). The surface temperature during the same time decreased by 3°C. Above 200 m the virtual potential temperature decreased by about 2°C (Fig. 12a). However, between the rain events at 1330 and 1700 LST, the temperature profile showed a period of partial recovery. The surface temperature increased in response to precipitation (Fig. 11a). The surface temperature during the same time decreased by 3°C. Above 200 m the virtual potential temperature decreased by about 2°C (Fig. 12a). However, between the rain events at 1330 and 1700 LST, the temperature profile showed a period of partial recovery. The surface temperature continued to decrease after sunset with the formation of a temperature inversion near the surface. A layer from about 300 to 800 m remains weakly stratified with a temperature gradient of less than 0.25 (the shaded region in Fig. 12a). For most of the afternoon there was a southerly wind component suggesting that the weakly stratified layer was initially residual island boundary layer air.

In contrast, the 26 November 1995 case had a single period of rain. After 1430 LST rain and thunderstorm outflows led to a surface cooling of 7°C (Fig. 11b). The diverging cold pool was evident in the surface winds with the meridional wind becoming southerly from the surface to about 400 m. Despite the limited area of the thunderstorm, its influence was long in duration. Southerly outflows persisted until after sunset when typically a southerly drainage flow forms. The surface layer did
not show a recovery as the cooling associated with thunderstorm outflows persisted until sunset. Above 400 m, however, the mean flow was northerly leading to the formation of a weakly stratified layer of residual oceanic boundary layer air (Fig. 12b).

These two cases illustrate the importance of thunderstorm timing and duration on the boundary layer development. When a thunderstorm passed directly over a site, cool, stable air was introduced through the full boundary layer profile at that site. The outflow from the thunderstorm spread out across the island with residual boundary layer air above. The mean winds advected this residual boundary layer air over the sites where thunderstorms had led to a cooling of the full boundary layer profile. Subsequently an increase in temperature was observed above the cool surface. When a reduction in the influence of thunderstorm outflows in an area occurred prior to sunset, an increase in surface temperature was also observed as shown in the development for 23 November 1995. In the case where thunderstorm activity is long in duration, no postthunderstorm recovery of the surface temperature was observed (as shown in the development for 26 November 1995). Instead there is a stabilization of the boundary layer, followed by the appearance of a mixed layer of residual island boundary layer air, later replaced by boundary layer air of oceanic origin.

6. Summary

The mean boundary layer development over Bathurst and Melville Islands was studied using a network of radar, AWS, all-sky cameras, radiosonde, and aerosonde. The data were collected during the Maritime Continent Thunderstorm Experiment (10 Nov–10 Dec 1995). The diurnal cycle is summarized in Fig. 13. The principle processes are the following.

1) Soon after sunrise, boundary layer air from higher
topography surrounding Maxwell’s Creek is advected over the cool air (near the surface) that has pooled in the shallow basin during the evening and early morning. A similar situation is expected at other sites that allow the pooling of cold air. Surface temperatures in areas where the pooling of cold air does not occur, increase more rapidly during the morning, and are expected to be warmer than the temperatures above the surface.

After about 0800 LST, the boundary layer grows rapidly as a result of surface heating, reaching a depth of about 1.5 km by early afternoon. During this time the surface temperature increased, but the water vapor mixing ratio decreased. Small cumulus develop near the top of the boundary layer late in the morning, with cloud cover increasing until thunderstorm activity initiates at about 1400 LST. The mean surface flow was generally westerly to about 500 m. However, the westerly winds weakened or were replaced with more easterly winds as the boundary layer mixing increased. The local circulations across the island show sea breeze circulations increasing in strength with a maximum in convergence at 1200 LST. The inflow generally has a depth of about 700–1000 m, with stronger easterly environmental flow above. Thunderstorms initially develop on the east coast and in areas where the sea breeze opposes the local environmental flow.

2) Thunderstorm outflows stabilize the boundary layer with the maximum cooling occurring near the surface. The convergence across the island is reduced and later replaced by a divergent flow because of the thunderstorm outflows replacing the sea breeze circulation. After sunset, a nocturnal temperature inversion forms and divergent drainage flows develop.

3) By about 2100 LST air of oceanic origin is advected over the islands introducing a weakly stratified layer and stable air above. This process essentially resets the clock for boundary layer development the next day.

This study was limited in the extent to which it could describe the horizontal variation in the boundary layer across the island. Three sites that formed a north to south chain were compared. Only one of these profilers (Maxwell’s Creek) had good quality RASS data. A future experiment would benefit by having wind profilers equipped with RASS in a north to south chain and an east to west chain. This would enable a more complete investigation of the effect of distance from the coast, thunderstorm activity, and thunderstorm outflows on the boundary layer depth.

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