On Trade Wind Cumulus Cold Pools

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ABSTRACT

Shallow precipitating cumuli within the easterly trades were investigated using shipboard measurements, scanning radar data, and visible satellite imagery from 2 weeks in January 2005 of the Rain in Cumulus over the Ocean (RICO) experiment. Shipboard rainfall rates of up to 2 mm h$^{-1}$ were recorded almost daily, if only for 10–30 min typically, almost always from clouds within mesoscale arcs. The precipitating cumuli, capable of reaching above 4 km, cooled surface air by 1–2 K, in all cases lowered surface specific humidities by up to 1.5 g kg$^{-1}$, reduced surface equivalent potential temperatures by up to 6 K, and were often associated with short-lived increases in wind speed. Upper-level downdrafts were inferred to explain double-lobed moisture and temperature sounding profiles, as well as multiple inversions in wind profiler data. In two cases investigated further, the precipitating convection propagated faster westward than the mean surface wind by about 2–3 m s$^{-1}$, consistent with a density current of depth ~200 m. In their cold pool recovery zones, the surface air temperatures equilibrated with time to the sea surface temperatures, but the surface air specific humidities stayed relatively constant after initial quick recoveries. This suggested that entrainment of drier air from above fully compensated the moistening from surface latent heat fluxes. Recovery zone surface wind speeds and latent heat fluxes were not higher than environmental values. Nonprecipitating clouds developed after the surface buoyancy had recovered (barring encroachment of other convection). The mesoscale arcs favored atmospheres with higher water vapor paths. These observations differed from those of stratocumulus and deep tropical cumulus cold pools.

1. Introduction

The trade wind cumulus region is important for the planetary radiation budget because it encompasses a large area, capable of compensating for a typically low cloud fraction (Medeiros et al. 2008). The steady surface winds encourage surface evaporation, with the moisture ultimately feeding into the tropical convergence zones (e.g., Tiedtke et al. 1988). Recent papers attest to the need to improve large-scale model representations of shallow clouds if the planetary radiation budget and hydrological cycle are to be simulated correctly (e.g., Bony and Dufresne 2005; Neggers et al. 2007). The challenge remains, given weak external forcing (i.e., slowly evolving sea surface temperatures and sporadic midlatitude synoptic disturbances), how to best represent local processes that may not yet be well observationally characterized. This includes precipitating trade wind cumulus. Shallow clouds can precipitate in amounts deemed climatically significant (Short and Nakamura 2000) but have been less studied than nonprecipitating shallow cumuli [see, e.g., the review by Stevens (2005)]. Basic questions as to the subcloud energetics and the convective triggering mechanisms of precipitating shallow cumulus remain.

The Rain in Cumulus over the Ocean (RICO; Rauber et al. 2007) experiment, held in the eastern Caribbean from December 2004 to January 2005, allowed for a more comprehensive view of shallow precipitation and its organization through combining scanning surface-based precipitation radar with satellite, aircraft, and ship measurements. An important and possibly new documentation was that nearly all clouds producing significant precipitation (>1 mm h$^{-1}$) were associated with arc-shaped formations reminiscent of cold pool outflow
boundaries formed by previous precipitation (Rauber et al. 2007; Snodgrass et al. 2009).

Early studies on shallow clouds built a paradigm that emphasizes subcloud moistening and cooling from the evaporation of light precipitation, doing little to change the subcloud moist static energy (Malkus 1958; Nitta and Esbensen 1974; Augstein et al. 1973; Albrecht 1993; Siebesma et al. 2003). Concomitant mesoscale organizations are typically either nonprecipitating cloud streets parallel to the wind or more randomly distributed clouds capable of some precipitation (e.g., LeMone and Pennell 1976; Nair et al. 1998). More recent work on stratocumulus cold pools has highlighted slight increases in the near-surface moist static energy of open cellular convection within stratocumulus (e.g., Stevens et al. 2005; vanZanten et al. 2005; Savic-Jovic and Stevens 2008; Wang and Feingold 2009b; Wood et al. 2011). In this nuance, evaporation from precipitation decouples the subcloud layer, suppressing vertical mixing and allowing an accumulation of surface fluxes (e.g., Savic-Jovic and Stevens 2008). The modeling work of Xue et al. (2008) show such moist outflows converging to form new cells, which they hypothesize may extend to the RICO mesoscale arcs. Another process that may have bearing on the RICO precipitation takes wind shear into account, documenting squall-line behavior at the advance of moist cold pools for stratocumulus off the coast of Australia (Jensen et al. 2000).

Snodgrass et al. (2009) and Nuijens et al. (2009), however, found RICO precipitation rates for the mesoscale arcs that were an order of magnitude higher than the roughly 1 mm day$^{-1}$ rate typical of stratocumulus. While the higher rain rates are not unheard of for trade wind cumulus (e.g., Szumowski et al. 1997), they are more typical of oceanic convection capable of exploiting ice microphysical processes. Tropical deep convective clouds can draft down air from above the cloud base capable of drying the surface air and reducing its moist static energy (e.g., Zipser 1969, 1977; Betts 1976; Gaynor and Ropelewski 1979; Houze and Betts 1981; Barnes and Garstang 1982; Addis et al. 1984; Young et al. 1995; Saxen and Rutledge 1998; Kingsmill and Houze 1999). Remarkably, Warner et al. (1979) focused on what appear to be mesoscale arcs of shallow convection from one day of the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE). They anticipated some of the work presented here, if lacking the precision and resolution available with today’s technologies. This suggests that the mesoscale arcs observed during RICO may have more in common with features observed during a tropical convection field experiment than with previous shallow convection observations, despite a mesoscale organization containing no ice microphysics. It was not clear to us a priori what to expect for the thermodynamic properties of the satellite-inferred cloud-free pools noticed by Snodgrass et al. (2009) and Rauber et al. (2007). Deep convective cold air pools can be either moister or drier than the unmodified surface air surrounding them (e.g., Addis et al. 1984; Young et al. 1995), and precipitation downdrafts were not observed near the surface in Hawaiian trade cumulus (Szumowski et al. 1997). Barnes and Garstang (1982) found that a precipitation rate of greater than about 2 mm h$^{-1}$ differentiated surface air that was dried rather than moistened by convection, but they did not distinguish convection by the presence (or lack) of ice particles. Precipitation rates during RICO straddled the Barnes and Garstang (1982) 2 mm h$^{-1}$ threshold (Snodgrass et al. 2009; Nuijens et al. 2009).

Previous studies from GATE and Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) highlight that the boundary layer recovery process can be highly sensitive to the convective organization (e.g., Saxen and Rutledge 1998), which provides little guidance on the boundary layer thermodynamics of the RICO mesoscale arcs. For example, mesoscale subsidence on cold pools induced by cirrus anvil precipitation is common for tropical deep convection (e.g., Zipser 1977; Houze and Betts 1981) but could be safely discounted for RICO, as cirrus anvils were not common.

The goal of this study was to relate the mesoscale arcs observed during RICO to the subcloud layer. The convective triggering mechanisms were of particular interest. We used data from the R/V Seward Johnson (hereafter referred to as RVSJ) along with radar and satellite data. The RVSJ plied the waters northeast of Barbuda ($17^\circ37'N, 61^\circ48'W$) during 9–24 January 2005, with 15–16 January spent in port. Its measurements included high-density surface thermodynamic data gathered from a flux tower, temperature and moisture profiles from radiosondes, continuously retrieved water vapor and liquid water paths from a microwave radiometer, a vertically pointing X-band (9.4 GHz or 3.2-cm wavelength) precipitation radar, a scanning K-band (35 GHz or 8.66-mm wavelength) cloud radar, and a motion-compensated 10-µm wavelength Doppler lidar. The ship was typically within 50 km of a scanning S-band dual-polarization Doppler (S-Pol; 10-cm wavelength) radar located on Barbuda. The S-Pol radar along with Geostationary Operational Environmental Satellite-12 (GOES-12) visible satellite data provided an overview of the cloud field for the point observations gathered on the ship. Despite a time series of only 14 days, a range of shallow clouds was observed within truly marine boundary layers of 2–4-km depth, both nonprecipitating and precipitating. Cold air pools were sampled almost daily.
This study is organized as follows: first a brief overview of all the data is provided, followed by a summary of the rain events recorded on the RVSJ and their impact on the thermodynamic measurements made on the ship. Three case studies of cold pools are presented next, followed by an interpretation of their data. This includes a discussion of the cloud vertical structure and of cloud triggering mechanisms. A schematic of the mechanisms associated with cold pool convection is presented and discussed, followed by conclusions.

2. Large-scale context and representativeness of data sample

A time series of selected observations provides a flavor for the measurements (Fig. 1); individual datasets are detailed further in the appendix. Figure 1a indicates radiosonde-inferred inversions [up to two per sounding, identified through dT/dp > 0.4 K (5 hPa)−1], along with cloud top inferred from hourly cloud radar reflectivity–height histograms. These show a range in cloud-top and inversion heights but also a preference for the 2- and 4-km height scale, particularly after 17 January. The lifting condensation level (LCL) was almost always between 500 and 700 m, typical for trade wind cumulus (Stevens 2005). Figure 1b indicates that ascending motion [as assessed by National Centers for Environmental Prediction (NCEP) reanalysis] prevailed at 500 hPa during 9–18 January, with descending motion thereafter. Hourly-mean ceilometer-derived cloud fractions are also shown. The mean cloud fraction for cloud bases up to 1500 m was approximately 18%, a typical value for tradewind cumulus.

Figure 1c shows the sea temperature measured at 5-cm depth (hereafter simply called SST; mean = 26.2°C) as well as the 15-m air temperature (Tair; mean = 25.3°C). The air temperature was often depressed by approximately 1 K from that of the ocean, also noted in the RICO aircraft data (Jensen 2006). Over an oceanic region with weak SST gradients, this typically indicates air cooled through the evaporation of precipitation. Figure 1d shows the rain rates recorded by gauges on the ship deck, and S-Pol-derived rain rates within a 1- and 6-km radius centered on the ship calculated using the Snodgrass et al. (2009) Z–R relationship. The radar-derived precipitation rates corresponded remarkably well to those measured by the ship’s optical rain gauges (see Fig. A2).

Figure 1e shows the total accumulation-mode aerosol concentrations from 18 January onward, measured at ambient relative humidity. Values ranged between 25 and 200 cm−3 (high values may have contained ship exhaust). Two water vapor path (WVP) values are shown in the following panel, one a high-time-resolution value derived from a microwave radiometer, and another derived from 6-hourly radiosonde measurements. The microwave-derived WVPs slightly exceeded the radiosonde-derived WVPs, but the two time series tracked each other well. The microwave-derived liquid water path (LWP) values are at times best interpreted as a precipitation proxy.

Figure 1g shows the latent and sensible heat fluxes, calculated using the bulk COARE 3.0 algorithm (Fairall et al. 2003) from the surface air specific humidity qair, Tair, and surface wind speed measurements, described further in the appendix. The latent and sensible heat flux time-mean values were 138 and 5.3 W m−2 respectively, with a considerable range. The bulk calculation of the latent heat fluxes may overestimate at higher wind speeds compared to the more precise inertial dissipation and eddy covariance calculations (see Fig. A1), while the bulk-calculated sensible heat fluxes can be lower by 2–4 W m−2 than the more accurate calculations independent of wind speed. This means the Bowen ratio of sensible to latent heat fluxes may be underestimated, particularly at high wind speeds. Figure 1h shows the winds. These were primarily easterly, with a mean wind speed of 6.1 m s−1.

While the area was not highly disturbed during the 9–24 January time period (Nuijens et al. 2009), on 9, 10, and the latter half of 13 January the ship sampled organized synoptic disturbances, and convection extended to temperatures below the 0°C level, evident in X-band radar data as a melting-level brightband. These days were also treated separately by Nuijens et al. (2009) because they possessed S-Pol radar echo fractions that exceeded three standard deviations beyond the mean value. Excluding these three disturbed days, the S-Pol radar documented ubiquitous rain throughout its scanning range, but at echo fractions not exceeding 0.1 (Nuijens et al. 2009). No cloud tops detected by the ship X-band radar extended to below 0°C levels on undisturbed days. We primarily focused on the undisturbed days because they contained the trade wind cumulus clouds that are the focus of this study.

3. Cold pools

Thirty-seven separate surface rain events were recorded on the RVSJ, of which 17 occurred on the undisturbed days. Their larger-scale spatial context was identified by overlaying half-hourly GOES-12 visible satellite data with the S-Pol radar reflectivity values and ship tracks. Of the 17 surface rain events recorded during the undisturbed days, 15 were clearly associated with mesoscale arcs of clouds encompassing cloud-free areas. The shipboard vertically pointing X-band radar showed that, barring one event, all surface rain events were associated with clouds deeper than 2 km. For 10 of the 17 events, the
ship appeared to pass through the center of a convective cell (as opposed to the periphery). For these 10 cases, the maximum surface rainfall rates exceeded 2 mm h\(^{-1}\), with surface rainfall accumulated over the period of the ship sampling typically still less than 1.5 mm. The precipitation amounts correlated with the temperature decreases (Fig. 2), a feature also noted by Barnes and Garstang (1982). Visual inspection suggested a correlation between the depth of convection and the size of cold pools.

Cold air pools are defined as “a region or ‘pool’ of relatively cold air surrounded by warmer air” or “any large-scale mass of cold air” (Glickman 2000, p. 157). This definition does not assume an origin for the colder air. Our study focused on precipitation-induced cold pools and we sought an additional explicit connection between the precipitation and a cold pool (e.g., Gaynor and Ropelewski 1979; Barnes and Garstang 1982; Young et al. 1995; Tompkins 2001). We defined the beginning of a cold pool time period through the onset of surface rain (0.0 mm h\(^{-1}\)) and the end of a cold pool time period by the end of the subsequent temperature recovery. These criteria were similar to those of Young et al. (1995), but we did not discriminate by a particular rain rate because we were interested in characterizing the thermodynamic response of the boundary layer to all shallow cumuli rain events. The surface air temperature

![Diagram](image-url)
recovery could either continue until the air temperature was in equilibrium with the sea surface temperature (i.e., fully recovered) or it could be terminated by the intrusion of another nearby cold pool decreasing $T_{\text{air}}$ again, a common occurrence in cold pools (e.g., Young et al. 1995; Tompkins 2001). With this definition, the ship spent between 15 and 70 min in individual cold pools, with half of the events occurring in the morning (between sunrise and noon). Such time intervals, deduced from a ship making nonstandardized transects through different cold pools, do not correspond directly to the size of the cold pools.

The pre-rain- to rain-event extremes in surface air properties are summarized in Fig. 3, with the 17 shallow-convection-only cases highlighted. The values represented in Fig. 3 were from the convective outflow region directly underneath the precipitation and not representative of the postconvection cold-pool recovery process. Notably, all of the rain events produced drier surface air (Fig. 3a) and a drop in the equivalent potential temperature (Fig. 3b). This indicated cohesive downdrafts from higher altitudes dominated the change in the surface air specific humidity, rather than evaporation of precipitation. Wind speeds generally increased during the rain events, by 2–4 m s$^{-1}$ (Fig. 3c), but not always. In response to the stronger winds, and drier, cooler surface air, latent and sensible heat fluxes increased (mean increases of 54 and 17 W m$^{-2}$, respectively; Fig. 3d). The sensible heat fluxes increased proportionally more (for a mean Bowen ratio increase of about 0.08). The modification of the subcloud boundary layer was similar on disturbed and undisturbed days, although the variability was greater within the “disturbed” day dataset (Figs. 3a,b).

**a. Case study selection**

We examined the convective triggering process and the postconvection boundary layer recovery using three case studies selected from the “undisturbed” days. Cases from 11 and 14 January each contained completely recovered cold pools, easing interpretation. In addition, the ship transects were upwind and bisected the convection and cold pool. This also eased interpretation; as discussed in Tompkins (2001), the spatial relationship of a sampled point to the convective core can influence the appearance of a cold pool. For the 11 January case, the surface air temperature within the postconvection cold pool fully equilibrated with the sea surface temperature, with the wake recovery process fully documented by the ship measurements. For the 14 January case, the ship documented convection coming into a completely recovered cold pool from the east, providing further insight into the convective initiation. A case from 19 January case was selected because it had been previously analyzed (Abel and Shipway 2007; Snodgrass et al. 2009), although its cold pools did not completely fit the pattern of the 11 and 14 January cases.

The synoptic conditions were described in Caesar (2005). A low-level easterly wind maximum on 11 January preceded a surge and weak trough that slowly moved west over the domain in the next couple of days. This was capped by dry air above 500 hPa advected in by northwest winds, with further drying after 13 January. On 18 January a cold front moved south through the region, increasing low-level moisture and destabilizing the lower troposphere. Its cross-frontal temperature gradient became lost but a wind shear line remained (Caesar 2005). Dry air above 500 hPa continued to cap the convection, and the clouds remained liquid in phase.

**b. 11 January**

On this day the ship traversed a cloud-free cold pool of diameter $\sim$40 km, visually estimated from the 1215 UTC satellite image in Fig. 4. A radiosonde was coincidentally launched at 1200 UTC [0800 local time (LT)] simultaneous with an almost 5-km-deep precipitating cloud, on the downwind, western edge of the sampled cold pool. This was arguably the most developed cell within the mesoscale arc encircling the cloud-free region. The ship was moving toward the east-northeast, while the cold pool it sampled was moving to the west-northwest (Fig. 4), so that the ship was sampling surface air that with time
increasingly recovered from its earlier modification by precipitation.

The surface air temperature within the postconvection recovery zone recovered completely to match the sea surface temperature (Fig. 5a). The initial rain-induced surface drops in temperature and specific humidity tracked each other well. The surface air specific humidity increased by 1 g kg\(^{-1}\) between 1200 and 1220 UTC and decreased slowly thereafter. The temperature increase during recovery was monotonic and more gradual, with \(T_{\text{air}}\) by 1320 UTC, approximately 80 min after the rain event. By 1320 UTC the warmed surface air was buoyant enough to support thermal plumes capable of reaching the lifting condensation level, at a surface relative humidity value of about 75%. Cloud development was apparent in the microwave-derived LWP values and in visible satellite imagery (Fig. 4), but below the detection thresholds of the S-Pol radar and shipboard X-band radar, indicating a lack of precipitation. While the wind speeds did not change significantly as the ship traversed the cold pool, the sensible heat fluxes decreased as \(T_{\text{air}}\) increased, so that the Bowen ratio (of sensible to latent heat fluxes) also decreased. The latent heat fluxes did not show a clear trend. This contrasts with results from tropical deep convection, wherein increased wind speeds during the recovery periods also enhanced the turbulent heat transfers (Addis et al. 1984; Young et al. 1995; Saxen and Rutledge 1998).

The 1200 UTC radiosonde through the convective cell at the western end is shown alongside the 0600 and 1800 UTC soundings in Fig. 6. The 0600 and 1800 UTC soundings were similar, allowing more interpretation of the sounding going through the cloud core. One feature was a dropoff in the 1200 UTC winds at 2.5 km, becoming westerly above 3 km, while the specific humidity did not drop off until above 4 km. The X-band radar returned signals up to 5 km, which indicated that precipitation may have reached into and above the wind shear region. While common for trade wind cumuli (e.g., Stevens et al. 2005), this highlights how wind shear helps in moistening the trade wind regions.

Details of the near-surface cold pool are provided in Figs. 6e and 6f. The 1200 UTC surface cooling extended...
Fig. 4. (top) GOES-12 visible image at 1-km resolution at 1145 UTC (0745 LT) 11 January. Islands are lightly outlined in yellow, and the cold pool observed by the ship is highlighted lightly in fuchsia. (middle) GOES-12 visible radiances overlaid with the S-Pol radar reflectivities, shown at 1145, 1215, 1245, and 1315 UTC, in region highlighted by the yellow box at top. Red line indicates the ship’s track, with red filled circle indicating the ship, and white dashed lines indicate 18°N and 62°W. The cold pool discussed in the text is highlighted with a pink dashed line. (bottom) Vertically pointing shipboard X-band radar data from 0800 to 1500 UTC. Black dashed line indicates the 1200 UTC sounding. Color bar indicates uncalibrated radar signal-to-noise ratio, best interpreted relative to itself.
up to 500 m. A decrease in the specific humidity profile explained \( \theta_e \) minima at 200 and 600 m. Comparison of the 1200 UTC \( \theta_e \) profile to the pre- and post-rain-event \( \theta_e \) profiles suggested the drying in the 0–200-m levels of the 1200 UTC sounding reflected air originating from at least 600 m. A mixing line analysis (shown later in Fig. 14) suggested that downdraft air reaching 200 m thereafter mixed with environmental air down to the surface. The conditionally unstable \( \theta_e \) profile in the lowest 200 m, forced from below by the surface fluxes, would encourage a new surface-based mixed layer to develop quickly to a depth of at least 200 m (e.g., Zipser 1969).

Four independent estimates of the propagation speed of the 1200 UTC convection suggested that the precipitating cloud moved faster than the mean surface wind speed of about 10 m s\(^{-1}\) shown in Fig. 5. The ship measurements for both days (Figs. 3 and 5) showed stronger surface wind speed at 1200 UTC by about 3 m s\(^{-1}\) than 6 h before or after. From the satellite visible imagery (Fig. 4), the line of convection was estimated to move about 70 km in 90 min, or about 12.7 m s\(^{-1}\). A more careful calculation tracked a particular 10-dBZ region from 1200 to 1206 UTC within the S-Pol data and produced an estimate of 14 m s\(^{-1}\) for the average propagation speed. The sonde-derived 1200 UTC wind speeds were about 12 m s\(^{-1}\) up to 2 km, significantly stronger than the 0600 and 1800 UTC winds. Warner et al. (1979) suggested that the mesoscale arcs they observed were moving at the
speed of the (faster) cloud-base winds, but the Fig. 6 wind profile is more consistent with the entire cloud cell was moving faster than the ambient air, for this case.

c. 14 January

On 14 January, a larger cloud-free area (estimated diameter of about 60 km from the 1545 UTC satellite image in Fig. 7) was examined. Its boundaries appeared to be formed of several separate mesoscale arcs, including on its eastern end. Some of the individual arcs can be seen developing and expanding within the 1545–1645 UTC satellite images. The visible imagery for the entire RICO domain was dramatic (Fig. 7). Similar to 11 January, the ship traveled eastward. Average surface winds measured on the ship were stronger than on 11 January, by about 1.5 m s\(^{-1}\).

The eastern end of the large, cloud-free zone apparent in the 1545 UTC satellite image (Fig. 7) was sampled by the ship from 1500–1600 UTC. The ship measurements (Fig. 8) also included vertically pointing lidar data from 1546 to 1612 UTC. The 1500–1600 UTC ship measurements showed temperature-recovered air (\(T_{\text{air}} \sim T_{\text{SST}}\)), and \(q_a \sim 15.2 \text{ g kg}^{-1}\), decreasing slightly with time for a mean relative humidity \(\sim 73\%\) (Fig. 8), lower than within the 11 January recovery zone. The zenith-pointing lidar prior to the 1600 UTC convection highlighted the deepening boundary layer and rising cloud base (Fig. 9).

Multiple cloud bases became apparent in the lidar signal-to-noise ratio within the 5 min preceding the convection (1555–1600 UTC). Closer inspection associated the higher bases with weak updrafts, and the lower bases with weak downdrafts. Downdrafts typically dry boundary layer air through entrainment, but in this case they may have constituted moist outflow from the approximately 3-km-high cloud encountered at 1600 UTC. Although its 10-min surface rain rate was only 0.5 mm h\(^{-1}\), the S-Pol radar showed that this cloud was bordered by stronger convective cells to its north and south. Lidar vertical velocities confirmed a strong, narrow, updraft at 1600 UTC, followed thereafter by a downdraft and a rain shaft with \(w \sim -5 \text{ m s}^{-1}\). The surface rain was accompanied by a temperature drop of over 2 K, drier surface air (from 15 to 13.6 g kg\(^{-1}\)), and lower \(\theta_e\) (from 344 to 337 K). Rain and clear-air velocities could not be decisively discriminated from each, but the narrow band of \(w < -4 \text{ m s}^{-1}\) extended above 1 km, consistent with descended dry air from above the cloud base. The lidar-perceived weaker updraft thereafter, despite surface rain, reinforced this interpretation.

Similar to the 11 January convective cell, independent estimates of the cell propagation speed also suggested the 1600 UTC 14 January convection moved more quickly than the mean winds. Surface wind speeds increased at the ship at 1600 UTC by about 2 m s\(^{-1}\) (Fig. 8). The visible imagery suggested the 1600 UTC cloud line propagated about 47 km from 1545 to 1645 UTC, or about 13 m s\(^{-1}\), almost 2 m s\(^{-1}\) faster than the average surface wind speed of 11.3 m s\(^{-1}\) (Fig. 8). The cloud line propagation speed estimated from tracking S-Pol radar reflectivity was 15.5 m s\(^{-1}\) from 1557:24 to 1607:12 UTC, also faster than the mean surface wind speed.

After 1600 UTC the ship measurements became chaotic in the presence of more convection encroaching from the east (at 1650 UTC a more developed convective cell reaching 5 km passed over the ship). A sounding released at 1625 UTC profiled a well-mixed surface cold pool up to about 200 m, with drier air above (Figs. 10c,e,f). The wind profile was similar to that of the 1200 and 2400 UTC soundings. The strong westerly wind shear at 4.5 km (Figs. 10a,b) explained the increased cloud fraction apparent in Fig. 7.

FIG. 6. Soundings at 0625, 1200, and 1753 UTC 11 January (black, blue, and green lines, respectively): (a) 0–5-km wind speed, (b) 0–5-km wind direction (90° indicates easterly), (c) 0–5-km specific humidity, (d) 0–5-km equivalent potential temperature, (e) 0–1-km potential temperature, and (f) 0–1-km equivalent potential temperature. Gray areas indicate 1200 UTC cloud. Filled circles near surface indicate flux tower values.
Fig. 7. (top) GOES-12 visible image at 1-km resolution at 1615 UTC (1215 LT) 14 January. (middle) GOES-12 visible radiances overlaid with the S-Pol radar reflectivities, shown at 1545, 1604, 1625, and 1645 UTC, in region highlighted by the yellow box at top. Red line indicates the ship’s track, with red circle indicating the ship, and white dashed lines indicate 18°N and 62°W. A dashed fuchsia ellipse in the 1545 UTC image indicates the cold pool recovery zone discussed in the text. (bottom) Vertically pointing shipboard X-band radar signal-to-noise ratios from 1400 to 2000 UTC. Black line indicates the 1625 UTC sounding.
d. 19 January

Although circular cold pools were apparent in the 19 January satellite imagery (Fig. 11), the linear cloud features oriented from southwest to northeast were more striking. A cold front had dissipated into a wind shear convergence line by midday (Caesar 2005). At 1335 UTC deep clouds reached 5 km, puncturing a strong inversion at 4 km aloft (Fig. 11), and generating surface rain rates of 45 mm h\(^{-1}\), more than an order of magnitude greater than for the previous two cases. A thorough synoptic analysis was not done for this case, but the narrow, intense rainband, followed by weaker surface rain rates at 1600 UTC from a cloud system with greater cloud coverage, would appear consistent with the surface and elevated cold-frontal rainbands described in Markowski and Richardson (2010, p. 130).

The ship was primarily moving northeastward as the clouds moved southward over the ship and approximately aligned with the linear cloud features, exaggerating their sampling by the ship (Fig. 11). The convective feature at 1340 UTC was highly visible to the X-band radar, while the detrained cloud from the convection sampled by the ship at 1600 UTC dominated the visual imagery. The ship measured relatively moist air compared to the earlier cases (mean RH; 80%; Fig. 12). The wind speed either decreased or did not change near the precipitation, in contrast to the two other cases. The surface air was nevertheless modified similarly (see also Fig. 3): the surface temperature only decreased by 1–2 K and the specific humidity by
A radiosonde, released at 1605 UTC during the second sampled rain event, showed the surface cold pool extended up to about 600 m, while for this case the specific humidity reached its minimum at the surface rather than at about 200 m as for the 11 January case. The $e$ profile was consistent with surface air originating from just above the cloud base (Fig. 13), but the depth of the surface air depression implied a more cohesive downdraft than for the other two cases. This, along with the order-of-magnitude heavier rain rate, was the most pronounced difference from the more self-propelled 11 and 14 January cases.

4. Cloud vertical structure

The cloud lines that propagated faster than the mean wind in the 11 and 14 January cases within an environment with some westerly shear has implications for the cloud vertical structure. Weaker winds aloft allowed upper-level precipitation to fall outside of the main precipitation shaft (see, e.g., the X-band radar data in Fig. 4 and the rain shaft in Fig. 9). If the air outside the rain shaft was drier than in the rain shaft, the precipitation would be more likely to evaporate without reaching the surface (than if it fell through the rain shaft).

A mixing line $q_a-\theta_e$ plot for the 0600 and 1200 UTC 11 January soundings, shown in Fig. 14, highlighted a region with elevated $q_a$ and $\theta_e$ values of 9–11 g kg$^{-1}$ and 335–340 K in the cloudy 0600 UTC sounding, consistent with updraft air. This bulge helped define a lower-altitude, lower-$\theta_e$ slot at 335 K with $q_a = 12$ g kg$^{-1}$. Both soundings also contained a sharp dropoff in $\theta_e$ and $q_a$ above 4 km, where the atmosphere transitioned into a dry westerly wind (Fig. 6). One interpretation for the $\theta_e$-$q_a$ minimum at 1.5 km was mixing of upper-level, drier, lower-$\theta_e$ air from a precipitation-driven downdraft with the environmental air. This has also been documented in Hawaiian trade wind cumuli with aircraft Doppler radar (Szumowski et al. 1997).

The 11 January broad $\theta_e$ minimum between 1 and 2 km (Fig. 6d) was also a minimum in relative humidity. In wind profiler data, the change in the backscattered reflectivities at about 1.5 km was indicative of an inversion (Fig. 15, top panel). The inversion appeared after the convection and, given its coincidence with the cold pool recovery zone, was consistent with the remains of a previous upper-level downdraft. Similar changes in wind-profiler reflectivities (and the presence of multiple inversions) for 19 January also suggested upper-level downdrafts (Fig. 15, bottom panel).

The inference of precipitation-driven downdrafts bringing down dry lower-$\theta_e$ air to midcloud levels helps explain another feature of the RICO data: two extrema in moisture and temperature values, near 2- and 4-km altitudes, in the soundings when composited by water vapor path into dry and moist conditions (Fig. 16). Sounding composites based on S-Pol radar echo fractions in Nuijens et al. (2009) highlighted the 2-km layer of enhanced temperature stability and moisture dropoff evident here for the driest tercile but did not emphasize such a double-lobed structure. Such a double structure was previously noted by Betts and
Albrecht (1987) for deeper convective boundary layers in the First GARP Global Experiment (FGGE) soundings over the equatorial Pacific and by Zuidema (1998) for the western tropical Pacific. They suggested that precipitation-driven downdrafts could be responsible, an inference that seems supported by this study. We note that the new, lower inversions discourage the vertical development of new shallow cumuli in the cold pool recovery.

Separate cloud populations, detached from the main subcloud layer with bases at the lifting condensation level, were evident in ceilometer-derived cloud heights near 2 and 4 km (Fig. 17), consistent with mass detrainment near layers of enhanced stability in the soundings, and also evident in satellite data (Genkova et al. 2007). More comprehensive surveys of trade wind cumuli have reported a continuum of inversion heights (Nuijens et al. 2009; Medeiros et al. 2010) and cloud tops (Zhao and Girolamo 2007). A double-inversion structure was also not a characteristic of the Barbados Oceanographic and Meteorological Experiment (BOMEX) and Atlantic Tropospheric Experiment (ATEX) soundings (Betts and Albrecht 1987). This may imply that a bimodal cloud distribution and double-inversion sounding structure is more likely to be associated with more heavily precipitating shallow cumulus, which is also consistent with upper-level precipitation-driven downdrafts. The height of the upper-level inversion would still reflect the larger-scale circulation.

5. Convective triggering

A goal for this study was to identify the mechanisms promoting cloud and convection development within the mesoscale arcs and around their recovery zones. For both the 11 and 14 January cases, the precipitating clouds sampled within mesoscale arcs moved faster than the mean winds, at the surface and at cloud-base level. We interpreted the surface wind speed increases under the convection as divergent air from precipitation-driven downdrafts that strengthened and moved faster than the prevailing winds to the west, and weaken (or converge) with the prevailing winds to the east. The downdrafts drove density currents of colder, drier air that, at the western end, could mechanically lift air that may form new convection, allowing the whole convective line to propagate faster than the prevailing wind, expanding the cold pool. The X-band radar data provided some evidence of new cloud development on the western side of the 1200 UTC 11 January convection (Fig. 4), and of the 1600 UTC 14 January convection (Fig. 7). A distinct line of enhanced reflectivities could also be seen to the west of the convection in the S-Pol radar data for 14 January (e.g., south-southwest of the ship track within the 1604 UTC image in Fig. 7). The reflectivity line was likely caused by frigate birds, who tend to prefer updrafts. The presence of low cloud bases within an otherwise deepening subcloud boundary layer (Fig. 9) suggests that it may be the moist outflow air that was lifted.

A density current propagation speed could be calculated, if we assume the cooler, drier air from the precipitation-driven downdraft became a density current. A density current speed estimate can be estimated from $\sqrt{gH(\Delta \theta / \theta)}$, where $H$ is the density current depth and $g$ is the gravitational acceleration (Simpson 1987). We estimated a depth $H$ of about 200 m from the depth of the surface-based conditionally unstable ($d\theta_e/dz < 0$) layer in the 1200 UTC 11 January sounding (Fig. 6), and surface-based cold pool depths within the 1600 UTC 12 and 19 January soundings (Figs. 10 and 13). This was combined with a buoyancy depletion estimate $\Delta \theta / \theta$ of 1 K/299 K to provide a current speed estimate of about
3 m s \(^{-1}\). New convection, when it was triggered by the mechanical lifting from the density current, would then propagate at a similar speed relative to the mean winds as the density current.

Other suggested convective triggering mechanisms include the convergence of moist outflows from previous open cellular convection (Xue et al. 2008), squall-line behavior at the advance of moist cold pools (Jensen et al. 2000), and a thermodynamic-only generation postulated by Tompkins (2001) for otherwise unorganized tropical deep convection. In the Tompkins (2001) mechanism, air moistened by evaporating precipitation was pushed outward by a drier precipitation-driven downdraft and triggered new convection when the surface air had recovered enough in temperature to become sufficiently buoyant. The convection occurred after the outflow velocity had dissipated away, and little if any dynamical forcing was involved.

These three mechanisms were difficult to detect in the observations presented here. Convergence of surface moisture to the east of the convective events was most evident as a slight increase in relative humidity in Figs. 5 and 8. This could serve to prolong the convective lifetime but would not serve as a convective trigger or expand the size of the cold pools. For the 1200 UTC 11 January case, previous convection and another nearby cold pool to the west may have increased the moisture supply available for the 1200 UTC convection, either as convergence of moist outflows (e.g., Wang and Feingold 2009a; Xue et al. 2008), or evaporated precipitation Tompkins (2001), but we lack the observational detail to confirm this. The wind shear convergence line on 19 January would also aid convergence of moist surface air. In general, however, we had trouble identifying positive water vapor perturbations in the surface data, and if so, their source. The dry, unsaturated downdraft of 1200 UTC 11 January did not reach the surface unmixed, creating a conditionally unstable atmosphere in the lowest 200 m that will further dry near the surface. Much of the evaporation may have remained at cloud level, consistent with an observed lack of small drops in RICO rain shafts (Baker et al. 2009) and the deduction of Snodgrass et al. (2009) that only about one-third of the

FIG. 11. (top left) GOES-12 visible image at 1-km resolution at 1555 UTC (1155 LT) 19 January. (top right) GOES-12 visible radiances overlaid with the S-Pol radar reflectivities, shown at 1245, 1315, 1345, 1445, 1515, 1545, 1615, 1645, and 1715 UTC, in region highlighted by the yellow box at top right. Red line indicates the ship’s track, with red circle indicating the ship, and white dashed lines indicate 18°N and 62°W. (bottom) Vertically pointing shipboard X-band radar signal-to-noise ratio from 1200 to 1800 UTC. The data are uncalibrated, with color indicating relatively stronger signal. Dashed line indicates the 1605 UTC sounding.
total precipitation was returned to the ocean surface, with drier air descending thereafter to lower levels.

Perhaps not surprisingly, the observations collected here are interpreted more readily using findings from other observational studies rather than from modeling studies. The Jensen et al. (2000) observations, of shallow convection organized in cloud lines propagating more quickly than the ambient wind, find common ground with the RICO convection propagation, even though their cold pools were moist rather than dry. A dry cold pool would be less buoyant than a moist cold pool, all else equal, and form a more pronounced density current. The temperature recovery processes for moist and dry cold pools should also differ.

Remarkably, the behavior documented here seemed most similar to the behavior of a tropical squall line (e.g., Barnes and Garstang 1982), in that new convection was triggered by a drier cold pool emanating from a propagating cloud line. Barnes and Garstang (1982) suggested that a (catchment-averaged) rain rate exceeding 2 mm h$^{-1}$ produced downdrafts of dry air driven by evaporative cooling penetrating into the subcloud air, a threshold that was met (by 5-min measurement averages) within our data sample. Observed rain rates may then help explain the difference between the RICO cold pools and the moist cold pools from stratocumulus and shallower cu-muli, and both types of cold pools could conceivably co-exist in the same region. Low-level wind shear may also

![Figure 12](image-url)
help explain the RICO observations of mesoscale arcs of warm rain (e.g., Rotunno et al. 1988), although modeling simulations more specific to RICO conditions would need to be done to assess the importance of wind shear (e.g., Bryan et al. 2006). Wind shear was not considered within either Tompkins (2001) or Xue et al. (2008).

The 1600 UTC 14 January convection was not strongly developed and encroached on air of lesser $\theta_e$ to its east (Figs. 8 and 9). In contrast, the 1200 UTC 11 January convection was well vertically developed with higher-$\theta_e$ air to its east (348 vs 344 K; Figs. 4 and 5). The origin of the higher-$\theta_e$ air on 11 January was unclear, but what does seem clear is that, if the cloud lines were propagating more quickly than the mean wind, more energy will be transferred per unit mass by mesoscale arcs moving into boundary layers with thermodynamically favorable properties (Barnes and Garstang 1982). If propagating convection is dependent on the properties of the air it is moving into, then it makes sense that moister conditions will support more precipitating convection. This provides an interpretation of another finding: when all of the scanning cloud radar reflectivities were composited by water vapor path, the deepest clouds and highest radar reflectivities occurred within the moistest tercile (Fig. 18), which also had a mean LCL 150 m lower than that for the driest tercile. Sixteen of the 17 rain events recorded at the RVSJ occurred during the upper WVP tercile. Savic-Jovcic and Stevens (2008) also note that their simulated new cumuli-like convection preferred the boundary regions with high-$\theta_e$ air; the observations here support the idea that new cumulus updrafts, occurring at the edge of current convection, are more likely to thrive if encroaching upon regions with high $\theta_e$.

A schematic for our interpretation of the mesoscale organization of precipitating trade wind cumuli is shown in Fig. 19. The top panel is intended to emphasize the depth of convection of the mesoscale arcs in relation to the direction of the prevailing winds: at the downwind, western end (left side of figure), new convection propagates faster than the surface winds by a value consistent with a density current calculation (Simpson 1987). Although not explicit in this schematic, the new convection thrives because it is entering an environment with high moist static energy. At the upwind, eastern end of the recovery zone (right side of figure), cloud develops when the surface buoyancy has recovered sufficiently to lift surface air to the lifting condensation level, marking the eastern edge of the cold pool. The bottom panel outlines the thermodynamic recovery process: at the downwind, western end, unsaturated convective downdrafts cool and dry the subcloud layer down to the surface. Immediately thereafter a new surface-based mixed layer
develops driven both by surface sensible and latent heat fluxes and entrainment warming and drying from above. The new mixed layer warms while the mixed layer mean specific humidity undergoes little change (after an initial rebound) and may even decrease, represented here by a lifting condensation level rising slightly with time. New cloud occurs when the mixed layer depth is sufficient to support thermal plumes capable of reaching the lifting condensation level. The size of the cloud-free zone is set at the eastern end by the recovery time (barring encroaching convection from the east) and, at the western end, by the propagation velocity normal to the convection.

6. Conclusions

Snodgrass et al. (2009) concluded that the most heavily precipitating trade wind cumulus occurred in mesoscale arcs around cloud-free areas, reminiscent of outflow boundaries of cold pools created by earlier convection. We focused on the relationship of such precipitating clouds to the subcloud layer using observations from the R/V Seward Johnson along with satellite and radar data. Of the 37 recorded shipboard rain events, 17 occurred on days typical of the trade wind cumulus regime. Nine of these 17 events passed through convection centers, as determined from satellite and S-Pol radar data.

The main findings were as follows:

- All of the rain events drafted down air that dried the surface air by up to $1.5 \text{ g kg}^{-1}$ compared to the environment (Fig. 3). The cooler, drier surface air resulted in decreases in the surface equivalent potential temperatures of up to 6 K. The decreases contrast with previous observational studies of precipitating shallow boundary layer clouds that documented cold pools moistened by the evaporating precipitation with higher values (e.g., Jensen et al. 2000; Stevens et al. 2005; vanZanten et al. 2005; Wood et al. 2011). The decreases documented in this study point to cohesive precipitation driven downdrafts capable of bringing down drier air from a higher altitude. This is more consistent with observations of convective outflow from tropical deep convection capable of invoking ice microphysical processes (e.g., Zipser 1969; Gaynor and Ropelewski 1979; Houze and Betts 1981; Barnes and Garstang 1982; Addis et al. 1984; Young et al. 1995) but has received little if any previous documentation for shallow cumuli.

- Ship traverses from west to east into easterly winds of two large (40–60-km diameter) cloud-free areas sampled surface air that monotonically recovered in temperature while its specific humidity remained relatively constant (1215–1305 UTC 11 January, shown in Fig. 5, sampled more of the recovery process than the 1500–1600 UTC 14 January example shown
in Figs. 8 and 9, which sampled air that was already mostly recovered). The eastern end of the recovery zone was either marked by buoyancy-driven cloud development (1320 UTC 11 January), or encroachment of new, faster-moving, convection from the east (1600 UTC 14 January). In both cases, the recovery zone surface air did not moisten despite high latent heat fluxes, implying compensation by entrainment of drier air from above as the recovering surface-based mixed layer deepened. Stratocumulus cold pools, in contrast, are thought to allow moisture to accumulate near the surface through suppressed vertical mixing (e.g., Savic-Jovcic and Stevens 2008). As the cold pools recovered, the wind speeds and latent heat fluxes remained approximately constant, while the sensible heat fluxes and Bowen ratio decreased back to pre-rain-event values. This differs from the stronger winds and enhanced turbulent fluxes documented within the cold pool recovery zones of tropical deep convection (e.g., Addis et al. 1984; Young et al. 1995; Saxen and Rutledge 1998).

The precipitating convection at 1200 UTC 11 January and 1600 UTC 14 January propagated faster than the mean surface wind, by 2–4 m s$^{-1}$, as inferred independently from radar data, satellite data, surface wind speeds, and the 11 January radiosondes. The increased wind speed was consistent with density current calculations. The finding of mesoscale arcs of warm rain moving faster than the mean surface wind is new, although Jensen et al. (2000) also found faster-moving boundary layer squall lines within a mostly stratocumulus region. One implication is that the cloud lines’ ability to thrive should depend on the moist static energy of the environment they are moving into, consistent with all the

![Figure 17](image-url)  
**FIG. 17.** (right) Ceilometer-derived cloud-base heights (both the lowest and second-lowest cloud bases are shown if available) as a function of the LCL. Gray shading indicates the cloud base fraction contained within a $100 \times 100$ m box, (left) Cloud fraction at each level at 100-m vertical resolution (black line; first cloud-base heights only), and the cumulative cloud fraction (blue line). Data from 9, 10, and 13 January were excluded. The ceilometer-derived cloud fractions are higher than those reported using aircraft aerosol lidar data (Nuijens et al. 2009; Medeiros et al. 2010), maybe because of different instrument thresholds or because near-surface aircraft lidar returns were excluded.

![Figure 18](image-url)  
**FIG. 18.** Scanning K-band reflectivities as a function of altitude, shown separately for terciles of microwave-derived WVP. The tercile-mean LCLs are indicated as black horizontal lines, with the tercile-mean value indicated within each panel: (left) 687, (middle) 617, and (right) 514 m. Left-hand plots of each panel show the integral over the reflectivities at each altitude. The gray bars indicate percentage of occurrence; note the change in scale for each tercile. The three WVP terciles represent <3.75, 3.75–4.1, and >4.1 cm.
mesoscale arcs occurring within the most moist water vapor path tercile.

- Higher-altitude evaporation, cooling air sufficiently to drive downdrafts to lower midcloud levels before mixing with the surrounding air, can explain a double maxima in the moisture and temperature sounding composites (Fig. 16) and in an individual sounding containing a new $\theta_e$ minimum (Fig. 6) that coincides with a wind profiler-inferred inversion (Fig. 15), as well as provide evidence of multiple inversions (Fig. 15). Cohesive precipitation-driven downdrafts have been postulated by Betts and Albrecht (1987) to explain such double-lobed structures in equatorial Pacific soundings, and Doppler-cloud-radar-resolved upper-cloud-level downdrafts previously in Hawaiian trade wind cumuli (Szumowski et al. 1997). The downdrafts, by bringing down upper-level lower-$\theta_e$ air to a midcloud level, discourage subsequent surface-based convection from reaching above 2 km despite the presence of moisture above the new lower inversion. This is an internal mechanism capable of altering boundary layer depth that is distinct from large-scale circulation influences.

For the three cases examined here the temperature and moisture inversion at 500–400 hPa was sharply defined, with westerly winds aloft that likely had a continental origin and were very dry (RH $\sim$ 20%). The very dry upper air may have strengthened the upper-level downdrafts by aiding evaporation. Cold fronts with a midlatitude continental origin, encouraging synoptic postfrontal subsidence and midtropospheric drying, may be overrepresented in the 2-week time period, aided by the proximity to the American landmasses and the January time period. Nevertheless, dry midtropospheric air intruding into the tropics, as well as the episodic occurrence of a trade wind regime, through whatever causes, is a well-known feature of the tropics [see, e.g., Johnson and Lin (1997) and references therein].

Further work will be needed to assess how general the triggering of new shallow convection by drier cold pools is, and if indeed the heaviest shallow convection can always be thought of as warm-rain squall lines. The ability of new shallow convection to thrive depending on the moist static energy (moisture in particular) of the environment the convection is moving into also needs to be evaluated further, with more comprehensive statistics than are provided here. New tools able to document the spatial distribution of moisture may help (e.g., Ellis and Vivekanandan 2010). The assessment of how shallow precipitation in the trade wind regime varies with water vapor path in other field experiments, satellite data, and model output can increase confidence in each dataset.

The ability for models to reproduce these observations of mesoscale arcs still needs to be explored. The RICO large-eddy-scale simulations discussed in vanZanten et al. (2011) emphasized roughly 2-km-deep clouds lacking organization on larger scales. In their study the cloud-layer evaporation of precipitation served to deepen the boundary layer only. Nevertheless, the vanZanten et al. (2011) study highlighted the importance of the vertical structure of
evaporation and thereby microphysical schemes—double-moment schemes are generally better at representing the smaller bulk fall velocities of the smaller drops contributing most to the overall evaporation [see Fig. 6 of vanZanten et al. (2011) and Table 2 of Shipway and Hill (2011, manuscript submitted to Quart. J. Roy. Meteor. Soc.), although aircraft observations during RICO highlighted a lack of small drops in rain shafts (Baker et al. 2009). In addition, the role of wind shear in initiating convection needs to be explored (e.g., Rotunno et al. 1988; Bryan et al. 2006). The fraction of significant shallow precipitation is only about 2% of the nominal 20% global-mean trade wind cloud fraction (Snodgrass et al. 2009), but the precipitation is climatically significant and worthy of further examination through both observations and models.

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APPENDIX

Descriptions of Individual Datasets

a. Flux data

The flux and meteorology measurement system is described in Fairall et al. (1997) and turbulent fluxes were calculated using the bulk COARE 3.0 algorithm (Fairall et al. 2003). Measurements were taken 14.3 m above the ship’s water line. The input $T_{air}$ was measured by a Vaisala HMP-300 series probe, housed in a radiation shield and aspirated by an electric fan, and is deemed accurate to $\pm 0.1^\circ C$. The surface specific humidity $q_a$ was computed from $T_{air}$ and the relative humidity. The relative humidity was also measured by the Vaisala HMP-300 probe, to an accuracy of $\pm 1.7\%$ RH. The wind speed was measured by a sonic anemometer and considered accurate to $\pm 2\%$ over a 10-min time period.

The RICO sensible heat fluxes calculated from the bulk formula were about 2 W m$^{-2}$ lower than the more accurate but more computationally expensive eddy covariance and inertial dissipation techniques at all wind speeds (not shown). The RICO bulk latent heat fluxes only compared well to the eddy covariance and inertial dissipation techniques up to wind speeds of approximately 6 m s$^{-1}$, however, and led to overestimation at the higher wind speeds (Fig. A1). The reason for this was unclear. The Bowen ratio may therefore be underestimated, particularly at wind speeds greater than 6 m s$^{-1}$.

b. S-Pol radar data

The National Center for Atmospheric Research (NCAR) S-band (10.62-cm wavelength) radar was located on Barbuda, typically 30–50 km away from the ship. The impact of sea clutter was diminished at this distance, though frigate birds could still be a source of high reflectivity (apparent within the 14 January case as a streak in S-Pol reflectivities to the south of the ship at 1600 UTC). A 7-dBZ threshold was used to exclude Bragg scattering returns (Snodgrass et al. 2009), effectively excluding cloud sensing. Only the scans at 0.58 elevation were used in this study. These included the 360° surveillance scans occurring every 20 min and the plan position indicator scans. At 30–50-km distance, the radar beam approximately scans a near-surface layer of 500–700 m deep, corresponding to the subcloud layer. The RICO S-Pol data were rewritten into network Common Data Form (netCDF) for this study and are available from NCAR’s Earth Observing Laboratory.

Several rain rate–reflectivity relationships have been applied to the S-Pol data. Snodgrass et al. (2009) developed $Z = 88R^{1.5}$ from NCAR C-130 data acquired on 19 January, while Nuijens et al. (2009) applied the TRMM $Z = 148R^{1.55}$ relationship. The exponents, which are sensitive to drop fall speed and thereby drop size, are approximately the same, but the Snodgrass et al. (2009) intercept parameter will estimate rain rates approximately 60% higher than those estimated using the
TRMM $Z-R$ relationship. For this study, we applied the $Z-R$ relationship developed by Snodgrass et al. (2009) because it was specifically developed for RICO. This S-Pol-derived rain rate within a 1-km radius centered on the ship compared well to the ship-based rain rates (Fig. A2).

c. Scanning 35-GHz cloud radar data

A scanning K-band (8.66-mm wavelength) cloud radar was located on the starboard side of the ship (i.e., the right side when facing the front or bow of the ship). The radar detection ranged from −20 to 30 dBZ and thus was able to detect cloud, distinguish cloud from light precipitation, and overlapped with the dynamic range of the S-Pol radar. The nonautomatic radar was only run during the daytime hours, performing range–height indicator scans up to 60° elevation, with a range of approximately 15 km. The radar was not stabilized, but the radar data altitudes were reported at earth-referenced values that indicated altitudes above the sea surface correctly.

d. Lidar data processing

A 10-μm wavelength Master Oscillator Power Amplifier (MOPA) Doppler lidar was run using four types of scans, of which we only showed an example of the signal-to-noise ratio and line-of-sight vertical velocity estimate from a zenith stare. A total of 942 min of lidar data were available, occurring in 20-min time blocks at 1-s time resolution, of which almost 97 min (10%) contained lidar-perceived cloud below 1.5 km. The nonautomatic lidar was only run during the day, at approximately the same time as the cloud radar. The data were initially corrected for both the horizontal and vertical ship motion through a spectral analysis, with the column-mean horizontal and vertical velocities removed. The 20-min-mean vertical velocities were thereafter removed from each 20-min sample. In addition, the mean velocity over the observation time period at each level was removed from the reported $w$, and outliers identified as $|w| > 5$ m s$^{-1}$ were also removed.

Thus, the vertical velocities reported here are perturbation velocities from an approximately 20-min mean. These are summarized in Fig. A3, separated into cloud and clear based on the lidar values themselves. The histograms of all available clear-sky vertical velocities follow a normal distribution. Most clouds are not associated with near-surface updrafts, consistent with a more buoyancy-driven origin for the clouds.

e. X-band radar data

The vertically pointing 9.4-GHz (3.2-cm wavelength) University of Miami radar was not motion stabilized, and only uncalibrated signal-to-noise ratio values were shown here. Its data were at a 30-m vertical resolution and 1–2-s time resolution. The radar 3.2-cm wavelength is more sensitive to precipitation than cloud particles.

f. Radiosondes

Sixty-one Vaisala RS-92 radiosondes were launched at 4-h intervals, at 0000, 0400, 0800, 1200, 1600, and 2000 UTC.

g. Aerosol data

A Particle Measuring Systems Lasair II particle counter used during RICO was contaminated during leg
1 and replaced during the port stay. The particle counter draws in air at ambient relative humidity through an optical forward scattering probe and counts aerosol particle concentrations in 6-μm diameter bins: 0.1–0.2, 0.2–0.3, 0.3–0.5, 0.5–1.0, 1.0–5.0, and >5.0 μm. Only the total aerosol concentrations are shown, with spikes attributed to ship exhaust removed.

**h. 915-MHz wind profiler**

The stabilized National Oceanic and Atmospheric Administration (NOAA) wind profiler inferred temperature and humidity gradients from the backscattered reflectivities, as described further in Ecklund et al. (1988).

**REFERENCES**


