Vertical Velocity Statistics in Fair-Weather Cumuli at the ARM TWP Nauru Climate Research Facility

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ABSTRACT

Fair-weather cumuli are fundamental in regulating the vertical structure of water vapor and entropy in the lowest 2–3 km of the earth’s atmosphere over vast areas of the oceans. In this study, a long record of profiling cloud radar observations at the Atmospheric Radiation Measurement Program (ARM) Climate Research Facility (ACRF) at Nauru Island is used to investigate cloud vertical air motion statistics over an 8-yr observing period. Appropriate processing of the observed low radar reflectivities provides radar volume samples that contain only small cloud droplets; thus, the Doppler velocities are used as air motion tracers. The technique is applied to shallow boundary layer clouds (less than 1000 m thick) during the 1999–2007 period when radar data are available. Using the boundary layer winds from the soundings obtained at the Nauru ACRF, the fair-weather cumuli fields are classified in easterly and westerly boundary layer wind regimes. This distinction is necessary to separate marine-forced (westerlies) from land-forced (easterlies) shallow clouds because of a well-studied island effect at the Nauru ACRF. The two regimes exhibit large diurnal differences in cloud fraction and cloud dynamics as manifested by the analysis of the hourly averaged vertical air motion statistics. The fair-weather cumuli fields associated with easterlies exhibit a strong diurnal cycle in cloud fraction and updraft strength and fraction, indicating a strong influence of land-forced clouds. In contrast over the fair-weather cumuli with oceanic origin, land-forced clouds are characterized by uniform diurnal cloudiness and persistent updrafts at the cloud-base level. This study provides a unique observational dataset appropriate for testing fair-weather cumulus mass flux and turbulence parameterizations in numerical models.

1. Introduction

Trade wind cumuli and fair-weather cumuli in general play an important role in the climate system. These clouds are observed over extensive areas of the earth and outrank in number any other type of cloud (Norris 1998). Although their fractional cloud coverage is small, fair-weather cumuli play key roles in maintaining the energy and moisture budgets of the tropics (Neggers et al. 2009) and the lower-tropospheric thermodynamic structure over extensive areas of the earth’s surface. In addition, since fields of fair-weather cumuli often have fractional cloudiness of 25% or more (Albrecht 1991), they can have an important effect on the radiation budget of the surface and the boundary layer (BL; Chertock et al. 1993). For climate sensitivity studies, marine boundary layer clouds remain at the center of tropical cloud feedback issues (Bony and Dufresne 2005). However, observations needed to define the percent of the fair-weather cumuli that are active in the moisture and energy transport processes relative to the total cloud amount and basic statistics on the vertical velocities in these clouds are lacking.

The representation of fair-weather cumulus clouds in numerical models is challenging because of the scale of these clouds and their intimate involvement with small-scale processes in the boundary layer, which are also difficult to parameterize. The vertical transport due to cumulus clouds in large-scale models is often parameterized using a mass-flux approach (e.g., Tiedtke et al. 1988). The
mass-flux approach for representing shallow cumulus clouds has also been used in simple bulk models of the boundary layer (e.g., Albrecht et al. 1979; Bretherton and Park 2008). Modifications and advancements of the mass-flux treatment of shallow cloud processes have been based on large-eddy simulation (LES) studies (e.g., Neggers et al. 2006).

The early studies of the Barbados Oceanographic and Meteorological Experiment (BOMEX; in 1969) and the Atlantic Trade Wind Experiment (ATEX; in 1969) provided key analyses (Augstein et al. 1973; Holland and Rasmusson 1973; Betts 1975), leading to a better understanding of the effects of cumulus transports on maintaining the large-scale heat and moisture budgets of the trades. In these early budget studies, there were no direct observations of the clouds responsible for the transports obtained from the large-scale budgets. Nicholls and LeMone (1980) used aircraft observations made in the subcloud layer of fields of shallow cumulus clouds that included observations at cloud base to examine mass transports into the cloud layer; however, they did not have detailed in-cloud measurements within the entire cloud layer. In general, cloud characterizations based on in situ aircraft penetrations have limitations, since the volume sampled during aircraft penetrations is relatively small and it is difficult to study the time evolution of the vertical structure of clouds from penetrations at a single level. Both the early budget and aircraft in situ studies were made before the development of sophisticated remote sensing systems such as cloud radars and lidars that can be used to define cloud properties (e.g., Kollias et al. 2001, 2007).

The LES provides the modeling framework for the study and evaluation of proposed theories for boundary layer cloud formation, evolution, dissipation, and cloud-scale processes such as entrainment and turbulence (e.g., Randall et al. 1992; Cuijpers and Duynkerke 1993; Siebesma and Cuijpers 1999; Stevens et al. 2001). The large-eddy simulations (Siebesma and Cuijpers 1995) based on BOMEX have supported the idea that the mass-flux concept can reproduce most of the cumulus vertical turbulent flux. Early field studies, however, were not designed with LESs in mind, and the finescale observations needed to constrain and evaluate LESs are limited. Furthermore, LESs applied to fair-weather cumulus conditions can produce copious statistics on the cloud dynamics simulated in the models. However, there are no equivalent large-eddy observations (LEOs), for evaluating these statistics. Thus, there is an increasing need for observations that can be used to evaluate LESs directly. In the Rain in Cumulus over the Ocean (RICO; Rauber et al. 2007) experiment, cloud statistics on updraft cores were obtained from about 850 cloud penetrations made at different levels in a variety of trade wind cumulus clouds using three aircraft (Abel and Shipway 2007). These statistics were compared with equivalent statistical characterizations from a cloud-resolving model.

In this study, we attempt to partially address the recognized need for cloud-scale observations in support of LESs and explicit cloud models using multiyear, high-resolution cloud radar observations of fair-weather cumuli observed (during an 8-yr period) at the U.S. Department of Energy (DOE) Atmospheric Radiation Measurement Program (ARM) Climate Research Facility (ACRF) at the tropical western Pacific (TWP) Nauru Island (Stokes and Schwartz 1994). The profiling cloud radar measurements are used to measure the in-cloud vertical air motion in nonprecipitating fair-weather cumuli that can be used to derive the diurnal cycle of updrafts’ and downdrafts’ magnitude and fraction. The observational dataset is classified according to the prevailing wind conditions into two different regimes—easterly and westerly winds—to isolate the island effect (McFarlane et al. 2005). The ARM observations for the Nauru site provide a unique opportunity to develop extensive statistics on the structure of fair-weather cumuli for both process studies and LES evaluations.

2. Methodology

The Nauru ACRF was established on the island republic of Nauru because of its location on the eastern edge of the tropical warm pool (0.52°S, 166.92°E), and the corresponding variability associated with the El Niño–Southern Oscillation (ENSO) cycle. The facility is located at the northwest shore of the island and continuous observations started in December 1998. The primary cloud measurements used in this study are the Doppler moments recorded by the ARM Millimeter Wavelength Cloud Radar (MMCR; Moran et al. 1998; Clothiaux et al. 2000; Kollias et al. 2007), and the observations cover an 8-yr study period (January 1999–July 2007). Millimeter wavelength radars have been established in the last two decades as the primary research tool for the investigation of the vertical structure of clouds and particularly for studying the properties of boundary layer clouds and the processes that affect those properties (Lhermitte 1987; Clothiaux et al. 1995; Kropfli et al. 1995; Kollias and Albrecht 2000; Shupe et al. 2008). Doppler lidars also provide turbulence information on the subcloud layer (e.g., Hogan et al. 2009). The MMCR temporal resolution is 10 s, the range resolution is 45 m, and its sensitivity in the overlying atmospheric column is better than −50 dBZ at all heights in the troposphere. This permits the detection of weak echoes from nonprecipitating broken clouds. The
Active Remote Sensing of Cloud Locations’ (ARSCL; Clothiaux et al. 2000) daily files are the starting point for our analysis. The ARSCL product discriminates clear-air radar volumes from hydrometeor-containing volumes. This is based on a radar cloud mask scheme (Clothiaux et al. 1995) along with lidar detections of the first cloud-layer base. The objective of the ARSCL hydrometeor mask is to detect all radar volumes that contain hydrometeors. However, in this study we are interested in high-quality Doppler measurements that often require high signal-to-noise ratio (SNR) conditions. Thus, additional quality control is added by removing all ARSCL-reported hydrometeor volumes with Doppler spectrum width values below 0.1 m s\(^{-1}\) and very low SNR conditions (below \(-10\) dB) to ensure high-quality Doppler measurements. Furthermore, the ability to detect the cloud edges depends on the radar reflectivity values near the cloud edges and the degree of the partial radar beam filling.

Cloud droplets have negligible terminal velocities (0.3 and 7 cm s\(^{-1}\) for a 10- and a 50-\(\mu\)m droplet, respectively). Observations of vertical air motions in shallow clouds are typically an order of magnitude higher (e.g., Warner 1977; Blyth and Latham 1985; French et al. 1999; Kollias et al. 2001; Damiani et al. 2008). As a result, we can use the cloud droplets as a tracer of the vertical air motion in fair-weather cumuli. This is a valid assumption if larger size droplets are not included in the radar resolution volume (Kollias et al. 2001). The exclusion of MMCR echoes that contain drizzle or raindrop size hydrometeors is often accomplished with the use of a radar reflectivity threshold (e.g., Sauvageot and Omar 1987; Frisch et al. 1995; Mace and Sassen 2000). In a recent study, Liu et al. 2008 proposed a parameterization of the radar reflectivity threshold for the transition from cloud to cloud and drizzle as a function of the total cloud droplet concentration.

Here, MMCR echoes that include contributions from large hydrometeors are rejected in our analysis using a two-step process. The sampling technique is applied to each MMCR range gate in the boundary layer (0.1–3 km) and to hourly blocks of observations that contain at least 5% of the time hydrometeor detections. Starting with a very low radar reflectivity threshold (\(-30\) dBZ), the correlation coefficient of the subset of MMCR radar reflectivities that are below this threshold and their corresponding Doppler velocities is estimated. The process is repeated for small increments (2 dB) of the radar reflectivity threshold to a maximum value (\(-5\) dBZ). Using the correlation coefficients estimated for the set of radar reflectivity thresholds, we select a threshold as the maximum value of radar reflectivity for which the correlation remains near zero and does not significantly deviate from the estimated value for lower threshold values. This dynamic allocation of the radar reflectivity threshold for the separation of cloud only and cloud and larger particles MMCR returns enables the use of the maximum available cloud returns and allows for fluctuations in the total cloud droplet concentration. On average, the selected radar reflectivity threshold was between \(-15\) and \(-22\) dBZ, with lower values near the cloud base and higher values toward the cloud top.

3. Results

Using the hourly estimated radar reflectivity threshold, the MMCR hydrometeor detections are classified as those containing returns only from small cloud droplets (cloud detections) and those containing larger drops with or without the presence of cloud droplets in the MMCR volume. At the Nauru TWP ACRF, more than 90% of the fair-weather cumuli detections were classified as cloud detections (Fig. 1). This illustrates the high fraction of very low reflectivities observed in the shallow clouds at the Nauru ACRF. Thus, the fair-weather cumuli studied here rarely produced precipitation size particles, and the vertical velocity statistics are representative of the entire shallow cumulus cloud field. Using only the MMCR echoes classified as cloud detections, we estimated the profile of averaged radar reflectivity in the fair-weather cumuli layer (Fig. 1). The averaged cloud base radar reflectivity is \(-32\) dBZ and increases to \(-25\) dBZ near the cloud top.

a. Long-term cloud properties trends at the Nauru ACRF

The monthly averaged hydrometeor fraction at Nauru during the 1999–2007 period is shown in Fig. 2a. The harsh environmental conditions and remoteness of the Nauru TWP ACRF resulted in long periods with no MMCR observations. From the beginning of the MMCR observations (January 1999) to the middle of 2000 when the first major break in the data occurs, the profile of hydrometeor fraction is bimodal with a shallow (suppressed) mode in the boundary layer, cirrus clouds, and no middle or deep convective clouds. During this period the average wind direction (from the soundings) in the boundary layer was almost always from the east (Fig. 2b). During the same period, the multivariate ENSO index (MEI) that is often used to monitor ENSO had negative values that are associated with La Niña conditions, as shown in Fig. 2b. The MEI is based on six variables: sea level pressure, zonal and meridional components of the surface wind, sea surface temperature, surface air temperature, and total cloudiness fraction of the sky (Wolter and Timlin 1993, 1998). In mid-2001, the MMCR malfunctioning hardware was replaced.
and cloud and precipitation observations resumed. During the same period, El Niño conditions began developing in the tropical Pacific. The onset of El Niño conditions resulted in weaker easterlies and the frequent observation of westerlies. Cloudiness increased at Nauru, and deep convection was observed frequently filling the entire atmospheric column with clouds. In general, these two distinct monthly averaged hydrometeor profiles (suppressed and deep) are highly correlated with the variable wind direction in the boundary layer.

In addition to cloudiness, the strength of easterlies and El Niño at the Nauru ACRF are examined with respect to the fair-weather cumuli clouds’ cloud-base velocity, cloud-base updraft fraction, and cloud boundaries (Fig. 3). During suppressed conditions (almost always easterlies and large negative MEI values), the monthly averaged vertical air motion at the cloud base has a mean around zero and the updraft fraction is close to 50%. In suppressed conditions, all observed clouds contribute to the monthly averaged statistics shown in Fig. 3. During deep convective conditions, the monthly averaged vertical air motion is positive (0.1–0.3 m s\(^{-1}\)) and updrafts at the cloud base of shallow cumuli clouds occupy 60%–70% of the area covered with clouds. In deep convective conditions, only the shallow clouds (thickness <1 km) contribute to the monthly averaged statistics, since deeper clouds have higher radar reflectivities and larger drops and thus are rejected by our radar threshold. Although we sample only the shallow mode of convection, the majority of the hourly averaged cloud scenes have strong updraft motions and high updraft fraction. Lastly, the cloud thickness remains constant (around 500 m); however, the cloud base elevates 100 m during westerly wind periods.

b. The island effect

The strength and direction of the boundary layer winds not only correlate with the strength of El Niño in the eastern Pacific but affect the low-level cloudiness at Nauru. The ACRF location is often at the leeward side relative to the prevalent easterly winds. Despite its small size (21.2 km\(^2\)) with little topography (highest elevation is 60 m), an island effect has been documented (McFarlane et al. 2005; Nordeen et al. 2001). During the day the island surface air temperatures are several degrees higher than the ocean surface temperature, and the surface air temperature differences can be as large as 6°C (Savijärvi and Matthews 2004) compared with about a 1°–2°C difference over the ocean. At night the island surface temperatures under easterly flow can be about 2°C cooler than the ocean temperature, although the surface air temperature differences are small and closer to the over water values. During the day a cloud plume over the island can be generated because of a contrast in
surface temperature and roughness length and a thermal heat island circulation, as shown by Savijärvi and Matthews (2004). They modeled the flow over Nauru, and the results show a rising motion downwind of the island and descending motion upwind of the island, as in previous studies (Garstang et al. 1975). When the wind direction is easterly, the generated cloud plume is often sampled by the ACRF. In their study, McFarlane et al. (16 June–15 July 1999) found differences in the low-level cloudiness and surface radiation during periods that the ACRF sampled cloud plumes and periods with non-plumes. Although the effects on cloud can be substantial during the day, a simple heat budget calculation of the change in the boundary layer temperature due to an enhanced surface heat flux from the island indicates that the subcloud (boundary) layer heating as air flows a distance of 5 km over the island would be about 0.05°C for a 5°C surface air temperature enhancement and a subcloud depth of 500 m. During night the effects on the boundary layer thermodynamic structure will be negligible.

The fair-weather cumuli observations are classified into periods with easterly flow (contribution from land-forced clouds) and westerly flow (oceanic fair-weather cumuli). The classification is based on the sounding record (twice a day) at the Nauru ACRF. For each sounding, the mean wind direction in the 0.5–1.5-km layer where fair-weather cumuli clouds form, evolve, and dissipate is calculated. The low-level horizontal wind flow is characterized as easterly if the mean wind direction is between 45° and 135° and westerly if the mean wind direction is between 225° and 315°. From a total of 6873 soundings available during the 8½ year period, 62.7% are classified as low-level easterly flow and 11.7% are classified as low-level westerly flow. The mean wind direction is interpolated hourly to match the temporal resolution of the hourly estimated cloud statistics. Since fair-weather cumuli are the primary focus of this study, the soundings are further classified into those that correspond to fair-weather cumuli conditions only. This reduces the number of soundings suitable for the description of the thermodynamic structure during easterly and westerly flow to 613 and 47, respectively.

Figure 4 shows the averaged thermodynamic structure of the lower atmosphere at the Nauru ACRF during fair-weather cumuli conditions for easterly and westerly low-level flow. In the cloud layer (0.5–1.5 km) the observed wind direction of the easterly and westerly flow is constant with height—90° and 280°, respectively—and exhibits little variability in direction. The averaged wind speed profiles for easterly and westerly flow are similar in pattern and intensity with a small reversal around averaged cloud-base height. Below the mean cloud base, the westerly flow is stronger (higher shear) and above the mean cloud base, the easterly flow is stronger. During BL westerly flow, the atmosphere contains higher
amounts of water vapor, as shown by the averaged profiles of mixing ratio. The largest differences in the mixing ratio are observed above the fair-weather cumuli layer (above 1.5 km). The layer from the surface to 0.6 km is well mixed during easterly BL wind flow (constant potential temperature $\theta_v$). During westerlies, the layer from the surface to 0.5 km is less well mixed (small gradient in $\theta_v$) because of a small gradient in potential temperature ($\theta$). The lifting condensation level (LCL) is about 0.5–0.6 km high for both wind regimes. Above the LCL, an undiluted parcel originating from near the surface would experience negative buoyancy to about 3.5 km during easterly BL winds. The presence of a deep stable layer is consistent with the lack of deep clouds during easterlies. In contrast, during westerlies, the level of free convection is around 1.5 km, which is consistent with the observation of deeper precipitating clouds (Fig. 4). The soundings obtained under suppressed convective conditions observed with easterlies show little evidence of well-defined inversion layer at cloud top that is prevalent under trade wind conditions in areas where subsidence may be stronger. The lack of a sharp trade inversion in the soundings is similar to that found during the RICO (Rauber et al. 2007) experiment, which was conducted in a somewhat similar meteorological environment.

c. Vertical velocity statistics in fair-weather cumuli

The diurnal cycle of the vertical velocity (air motion) statistics for fair-weather cumuli is presented in this section for easterly BL winds (EBLWs) and westerly BL winds (WBLWs). The EBLW fair-weather cumuli periods are defined as periods with averaged BL wind from between 45° and 135°, hourly averaged cloud base below 2 km, and cloud thickness less than 1 km (shallow-only clouds). The WBLW fair-weather cumuli periods are defined as periods with averaged BL wind between 225° and 315°, hourly averaged cloud base below 2 km, and cloud thickness less than 1 km (shallow-only clouds). Using the aforementioned criteria, 8948 h of EBLW conditions and 649 h of WBLW conditions were identified. For each hour, several parameters derived from the vertical velocity time–height fields (e.g., updraft fraction, mean updraft velocity) were estimated and composite diurnal cycles for the two regimes are presented. As a

![Fig. 3](image-url)
word of caution, it is important to note that the bulk of the statistics presented here are estimated only in areas where we have radar observations; thus, they are “in cloud” estimates. This technique is a departure from the standard method applied in LES model output for the estimation of similar parameters and uses both clear and cloudy areas of the model domain. The updraft mass flux is the only presented variable that has been scaled with the cloud fraction to provide a “cloud + environment” estimate, assuming negligible contribution from the environment.

The diurnal cycle of the cloud fraction during EBLW and WBLW conditions is shown in Fig. 5. During EBLW
conditions, a strong diurnal cycle is observed. The nighttime cloud fraction is about 15%–16% and gradually reaches a maximum of 25% at 1400 LST. This is consistent with the island effect that is amplified during the daytime. In contrast, the cloud fraction shows little variability during WBLW conditions. Local minima in cloud fraction are observed around 2400 LST (midnight) and 1000 LST. Overall, the cloud fraction during WBLW conditions matches the nighttime cloud fraction during EBLW conditions. This suggests that we can use the nighttime cloud observations to study cloud fields unaffected by the presence of the island. The profiles of the cloud properties are shown in absolute height, which could affect the vertical variation of some of the properties shown in Figs. 6–8.

The diurnal cycle of the profile of cloud fraction during the two wind regimes is shown in Fig. 6a. During EBLW conditions, the cloud base elevates from 0.5 to 0.6 km and the cloud top from 1.1 to 1.2 km during the daytime as the cloud-layer thickness remains stable at 0.6 km. The vertical profile of the cloud fraction exhibits a parabolic profile during the daytime with a maximum cloud fraction of 25% observed in the middle of the cloud layer (0.8–0.9 km). During WBLW conditions, the cloud boundaries fluctuate with no clear diurnal cycle. The lack of a diurnal cycle seems to contradict earlier studies (typically in deeper, precipitating clouds) showing a maximum in cloud coverage and precipitation over the ocean occurring near dawn. The limited sample size (compared to the one available during EBLW conditions) and the exclusion of shallow precipitating clouds that are experiencing a diurnal cycle in this study can explain the lack of a clear diurnal cycle during WBLW conditions.

The time–height variability of the hourly averaged mean vertical velocity during EBLW and WBLW conditions is shown in Fig. 6b. Large differences are observed that extend beyond the daytime period and suggest that the differences in the observed fair-weather cumuli during EBLW and WBLW are not due to island effects only. During nighttime EBLW conditions, the lower two-thirds of the cloud layer has a nearly 0 hourly mean vertical air motion and the upper one-third cloud layer has hourly mean downward velocities (0.10–0.15 m s\(^{-1}\)). After 0800 LST, the hourly mean vertical velocity field gradually changes. Positive (upward) vertical velocities are observed in the entire cloud layer during the 0800–1800 LST period. The hourly mean vertical velocity profiles exhibit a parabolic shape with the maximum values (0.30–0.35 m s\(^{-1}\)) observed in the middle of the cloud layer at 1400 LST. During WBLW conditions a very different time–height hourly averaged vertical velocity field is observed. There is no distinct diurnal cycle, and the cloud layer is characterized by upward vertical velocities in the lower 80% of the cloud layer. The magnitude of the hourly averaged vertical velocities fluctuates between 0.10 and 0.35 m s\(^{-1}\) with the high values observed in the lower half of the cloud layer. In the top 20% of the cloud layer, we observed downward vertical velocities throughout the 24-h period. The vertical velocities are in-cloud only and do not reflect directly—but may infer indirectly—mass compensating velocities in the clear area around the clouds.

The hourly estimated standard deviation of the vertical velocity is shown in Fig. 6c. In both regimes (EBLWs and WBLWs), the maximum of the vertical velocity variance is observed near the cloud top throughout the 24-h period. The minimum vertical velocity variance is observed at the cloud base along with a gradual increase from cloud base to cloud top. The magnitudes of the standard deviations are comparable for the two periods. During daytime EBLW conditions, we observe higher values.

The vertical velocity measurements are further decomposed using direct sampling to identify updrafts and downdrafts at each MMCR range gate within the cloud layer. Figure 7a shows the time–height variability of the hourly estimated in-cloud updraft fractional coverage. It is important to note that the in-cloud updraft fraction is over the total cloudy area only. Thus, a 0.5 in-cloud updraft fractional coverage in a 0.2-cloud-field fraction is equivalent to a 0.1 updraft fractional coverage for the entire 1-h period. As shown previously, the daytime and nighttime EBLW periods have distinctly different updraft fractional coverage. During nighttime, the updraft fractional coverage is close to 0.5. In the lower two-thirds of the cloud layer, the updraft fractional coverage is 0.5–0.55 with a local maximum near the middle of the
Fig. 6. Diurnal cycle of the hourly averaged (a) cloud fraction and (b) vertical air motion (m s⁻¹) with positive velocities indicating updrafts and (c) its standard deviation during (left) EBLWs and (right) WBLWs.
FIG. 7. (a),(b) As in Figs. 6a,b, but for updraft fraction. (c) As in Fig. 6c, but for updraft mass flux (kg m$^{-2}$ s$^{-1}$).
cloud layer (0.55). The upper one-third of the cloud layer has an updraft fractional coverage of less than 0.5 with a minimum (0.45) at the cloud top. During the daytime, updrafts cover a higher fractional area. Between 1000 and 1600 LST, updrafts cover more than 0.6 of the cloudy areas with a maximum near-mid-cloud level (0.65). During WBLW conditions, no diurnal cycle is observed. Updrafts cover on average 0.7 of the cloudy areas in the lower half of the cloud layer and only near the cloud top does the updraft fraction drop to 0.5. This implies a large percentage of active fair-weather cumuli for the WBLW conditions.

The time–height variability of the hourly averaged updraft magnitude is shown in Fig. 7b. During nighttime EBLW conditions, updraft magnitudes increase linearly from the cloud base to the cloud top. Thus, the highest updraft values are observed near the cloud top (0.7–0.8 m s\(^{-1}\)) and the lowest near the cloud base (0.4–0.5 m s\(^{-1}\)). The same pattern is observed during the daytime EBLW conditions; however, the cloud-base updraft magnitude is 0.75 m s\(^{-1}\) and the cloud-top updraft magnitude is >1 m s\(^{-1}\). During WBLW conditions, no diurnal variability is observed and the updraft magnitude increases within the cloud layer with minimum updrafts.
values at the cloud base (0.55–0.65 m s\(^{-1}\)) and maximum at the cloud top (0.8–1.0 m s\(^{-1}\)). The contribution of the observed in-cloud updrafts to the mass flux (kg m\(^{-2}\) s\(^{-1}\)) was estimated as a product of the updraft fractional coverage, the updraft magnitude, and a reference air density of 1.1 kg m\(^{-3}\) (Fig. 7c). During nighttime EBLW conditions, the maximum updraft velocities near the cloud top and the maximum cloud fraction and updraft fractional area near the middle of the cloud layer results in a relatively uniform updraft mass-flux profile (0.04–0.06 kg m\(^{-2}\) s\(^{-1}\)). The highest values of updraft mass flux (0.16–0.18 kg m\(^{-2}\) s\(^{-1}\)) are observed during daytime at the middle of the cloud layer, where both the updraft fractional area and cloud fraction are at a maximum. During WBLW conditions, the presence of the strongest updrafts near the base of the cloud layer and the maximum of cloud fraction near the middle of the cloud layer results in maximum updraft mass fluxes near the lower and midcloud levels (0.08–0.10 kg m\(^{-2}\) s\(^{-1}\)).

The time–height variability of the hourly averaged downdraft velocities is shown in Fig. 8a. During nighttime EBLW conditions, the strongest downdrafts are observed near cloud top (0.9–1.0 m s\(^{-1}\)). The downdraft magnitude decreases with increasing height and thus the weakest downdrafts are observed at the cloud-base level (0.5–0.6 m s\(^{-1}\)). A similar pattern of downdraft magnitudes is also observed during the daytime EBLW conditions; however, the downdrafts at cloud base are stronger (0.7 m s\(^{-1}\)). During the WBLW conditions, a similar trend is observed; however, the observed downdrafts at the cloud top are stronger (1.1 m s\(^{-1}\)). Using the hourly observed vertical velocity measurements, we estimated the skewness of the vertical velocity (Fig. 8b). Positive skewness indicates the presence of few narrow strong updrafts and negative skewness indicates the presence of few narrow strong downdrafts (e.g., Kollias and Albrecht 2000). A Gaussian distribution of vertical air motion measurements has zero skewness and 0.5 updraft fractional coverage. Positive values of skewness indicate that the distribution of vertical air motion measurements has a tail of few but strong updraft values that are compensated by a larger number (thus higher fraction) of weaker downdrafts. During nighttime EBLW conditions, the vertical velocity skewness is near zero, with positive values near cloud base and negative values near cloud top. This is consistent with updraft fractional coverage more than 0.5 near the cloud base and less than 0.5 at the cloud top. Nevertheless, the skewness values during nighttime are close to zero and the asymmetry of the vertical velocity field is small. However, during daytime EBLW conditions, the skewness is higher and negative throughout the cloud layer. This implies the presence of a tail of few (narrow) strong downdrafts in the observed distribution of vertical air motion and such distribution is consistent with the observations of high updraft fraction and updraft magnitude. During WBLW conditions, the vertical velocity skewness has large negative values near the cloud base. This is consistent with the very large updraft fraction at the lower half of the cloud layer during WBLW conditions. In contrast, smaller skewness values are observed at the upper half of the cloud layer, consistent with near-0.5 updraft fractional coverage.

4. Summary

Fair-weather cumuli have been extensively studied using LES models. However, detailed, cloud-scale observations of in-cloud vertical velocities suitable for direct evaluation of the in-cloud vertical velocity statistics in LESs are not available. A comprehensive analysis of a multiyear fair-weather cumuli dataset from the Nauru ACRF is presented here in an effort to provide the necessary observations for LESs of fair-weather cumuli. The partitioning of the dataset into easterly (EBLW) and westerly (WBLW) conditions allows us not only to isolate island-generated effects but also to study the vertical velocity statistics in shallow cumuli during normal (easterly) and anomalous (westerly) conditions in the TWP. The significantly smaller sample of WBLW conditions resulted in more noisy diurnal composites of cloud properties. However, the sample is large enough to highlight the main features of fair-weather cumuli during WBLW conditions.

The presence of island effects is documented during the daytime EBLW conditions. The island influences the observed cloud fraction and the cloud vertical velocity statistics. For the daytime EBLW conditions, cloud fraction gradually increases from about 15% in the morning to a maximum of 25% in the afternoon. The hourly mean in-cloud vertical velocity is upward with an afternoon maximum in the middle of the cloud layer, and the variance of the vertical velocity is greatest near the cloud top. The island-forced clouds’ updraft fractional coverage, updraft magnitude, and updraft mass flux exhibit a strong daytime cycle. The updraft fractional area reaches a maximum at midcloud level in early afternoon, and the profile of the updraft magnitude is increasing with height. The maximum of the updraft mass flux coincides with the location of the maximum updraft fractional coverage. During the same period, downdrafts have their maximum value at the cloud top and decreasing magnitude with decreasing height. The large number of hours included in the diurnal statistics (8945) provides a very clear picture of vertical velocity statistics of shallow cumuli forced by localized surface heat sources.
The findings are suitable for the evaluation of models that treat surface in homogeneities.

Nighttime observations of vertical velocity do not exhibit the striking differences that we observed during daytime between easterly and westerly flows. Although this suggests that nighttime EBLW conditions are not affected by the presence of the island to the extent that daytime clouds are, the observations are insufficient to totally eliminate the possibility of nighttime island effects during easterly flow clouds. But a comparison of the LCLs from night time monthly averaged 3-m temperature and humidity measurements from the 1999–2000 period when easterlies prevailed at Nauru indicates good coupling between conditions at the surface and the observed cloud-base height from ceilometer measurements. The 24-month average of the LCL is 600 ± 83 m and the cloud-base height (from the lowest 25th percentile of cloud-base height measurements during each hour) is 676 ± 71 m. The averaged cloud fraction is 15%–16%, the mean vertical velocity is near zero, and the vertical velocity variance has its maxima near cloud top. The mean updraft and downdraft velocities have their respective maximum at the cloud top and their minima at the cloud base and their magnitudes are similar. Overall, the vertical velocity field can be considered as “symmetrical” with similar fractional coverage for updrafts and downdrafts and near-zero vertical velocity skewness values.

During WBLW conditions we sampled fair-weather cumuli that formed over the ocean during disturbed conditions at the Nauru ACRF that result in deep precipitating clouds. Overall, the cloud properties exhibit no significant diurnal cycle. However, the exclusion of shallow precipitating cloud fields and the relatively limited available dataset (649 h) may have contributed to the noisy character of the WBLW cloud-layer statistics. The cloud fraction is about 15%–16%, which is very similar to the cloud fraction observed during the nighttime EBLW conditions. However, the equal cloud fraction amount is the only similarity between the oceanic fair-weather cumuli during nighttime EBLW and WBLW conditions. The clouds observed during WBLW conditions are positively buoyant (active clouds) within their envelope, which is supported by their very high updraft fractional coverage (0.7) in the lower half of the cloud layer and mean upward velocity throughout the 24-h period (Stull 1985). In contrast, the shallow cloud fields observed during nighttime EBLW conditions are characterized by near zero in-cloud updraft mass flux and could be classified as passive “decaying” clouds (Albrecht 1981; Stull 1985). According to Albrecht (1981) these decaying clouds may account for a significant portion of the cloud field, depending on the evaporation rate.

The observed cloud characteristics from this study can be compared with results from LESs made for shallow cumulus clouds as part of the Global Energy and Water Cycle Experiment Cloud System Study Working Group. In one case, simulations of fair-weather cumuli under a strong inversion as observed during ATEX were studied (Stevens et al. 2001). In a second case, LESs were used to study the characteristics of shallow cumulus clouds associated with a weaker inversion as observed during BOMEX (Siebesma et al. 2003). In both cases, the cloud fraction near cloud base is about 6% and decreases with height; although for the ATEX case, the cloudiness increases to about 30% at cloud top. For the Nauru case, the cloudiness (excluding the daytime easterly conditions) is about 10% at cloud base and increases to maximum in the middle of the cloud layer of about 15%. For both the ATEX and BOMEX cases, the updraft core fraction is about 0.03. The Nauru observations, however, indicate updraft fractions that are larger than those simulated by LESs with 0.05 at cloud base for the nighttime easterly case and closer to 0.07 for the westerly cases. In the ATEX LES, the core velocities increase with height to a maximum of about 1.2 m s⁻¹ at about 500 m above cloud base. In the BOMEX case, the updrafts increase upward to about 1.6 m s⁻¹ at 500 m above cloud base and further increase to about 3 m s⁻¹ at 1500 m above cloud base. The BOMEX simulation of core vertical velocities is similar to those observed during RICO from aircraft penetrations (Abel and Shipway 2007). The Nauru cases with a shallower cloud layer indicate a similar rate of increase of cloud updraft velocity with height with maximum values of 0.8–1.0 m s⁻¹ reached more than 400–500 m above the cloud base. The mass fluxes for the Nauru cases are maxima in the lower and middle parts of the cloud and are higher than the values reported in the LES cases, which are about 0.02 kg m⁻² s⁻¹ and are either relatively constant with height or decreasing with height.

This study has provided for a unique set of cloud radar observations of the dynamical and macroscopic characteristics of shallow cumulus clouds at a tropical ocean site. A more detailed analysis of the cloud statistics generated will allow for detailed comparisons with LESs driven by the Nauru environmental conditions. Furthermore, these observations will be useful in evaluating parameterizations of cumulus mass fluxes and fractional cloudiness and in improving our understanding of the factors and processes that control these parameters.

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