Observations of Shallow Trade-Wind Cumulus Cloudiness and Mass Flux Variability and their Relationship to Boundary Layer Structure

Katia Lamer

Master of Science

Supervised by Professor Pavlos Kollias
Department of Atmospheric and Oceanic Sciences

McGill University
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Ultimately, this experience shaped me as a proud and strong student, scientist and person and despite the challenges, I do not regret undertaking this journey.
Contribution of authors

This text is to be submitted to a journal promptly. The research in this thesis was entirely carried out by me under the supervision of Prof. Pavlos Kollias. Through multiple discussions and revisions, Prof. Pavlos Kollias helped improve the quality of my analysis and manuscript. The Max Plank institute of meteorology provided the raw radar data and specifically Ilya Serikov provided Raman Lidar data products while Louise Nuijens contributed sporadic scientific comments.
Abstract

Oceanic cumuli are ubiquitous in the subtropics and help regulate the thermodynamic structure of the lower troposphere and the large-scale tropical circulation. Yet, observations of the vertical structure of marine cumulus cloudiness and mass flux remain sparse. Fortunately, modern island-based remote sensors more specifically, the Barbados zenith cloud radar posses the sensitivity to study hydrometers which, when small enough, can also be used as tracers of air motion.

Over the 2-years analyzed, the daily cloud fraction profile of cumulus non-precipitating (peak 4.5 % at 710 m), precipitating without (peak 2.3 % at 809 m) and with attached stratiform layers (peak 1 % at 1679 m) oscillate slightly (~ 3 %). Precipitating cumuli with stratiform outflows are deeper and contain more abrupt vertical variations in reflectivity and Doppler velocity than other precipitating cumuli. Non-precipitating cumuli exhibit an elevated reflectivity core and an ascendant center surrounded by a subsiding shell. Bulk (3 h) statistics reveal that these cloud are active and organized. They contain 79 % updrafts with 86 % of them being organized in large coherent structures contributing to a maximum updraft mass flux of 0.008-0.036 kgm$^{-2}$s$^{-1}$ just above cloud base. Alternatively, downdrafts contribute insignificant mass flux and show little vertical and temporal variability (0.000-0.007 kgm$^{-2}$s$^{-1}$). Together with complementary Raman lidar information, normalized updraft mass flux profiles suggest that updraft mass flux profile slope is inversely related to environmental relative humidity. Similarly, the presence of stratiform detrained layers coincides with moistening under the inversion layer. Altogether, the analysis presented here is useful for evaluating and constraining the representation of cloudiness and mass flux schemes in numerical models.
Les cumulus océaniques sont omniprésents dans les régions subtropicales et aident à réguler la structure thermodynamique de la basse troposphère ainsi que la circulation tropicale à grande. Pourtant, une pénurie d’observations de la distribution verticale des cumulus océaniques et de leur flux de persiste. Par chance, les capteurs à distance modernes basés sur des îles, plus précisément le radar de nuage vertical de la Barbade possède la sensibilité pour étudier les hydrométéores qui, quand ils sont assez petits, peuvent aussi être utilisés pour retracer le mouvement de l’air.

Au cours des 2 ans analysés, le profile de fraction nuageuse des cumulus sans précipitation (maximum 4.5 % à 710 m), avec précipitation sans (maximum 2.3 % à 809 m) et avec segments stratiformes attachés (maximum 1 % à 1679 m) oscille légèrement (~ 3 %). Les cumulus avec précipitation et couche stratiforme détachée sont plus épais et contiennent des variations de réfléctivité et de vitesse Doppler plus abruptes que les autres cumulus pluvieux. Les cumulus sans précipitation possèdent un noyau de réfléctivité surélevé et un centre ascendant entouré par une coquille descendante. Des statistiques globale (3 h) montrent que ces nuages sont actifs et organisés. Ils contiennent 79 % de courants ascendants et 86 % d’entre eux sont organisés en de larges structures cohérentes qui contribuent au flux de masse maximum de 0.008-0.036 kgm$^{-2}$s$^{-1}$ juste au dessus de la base des nuages. D’un autre côté, les courants descendants contribuent un flux de masse négligeable and montrent peu de variations verticales et temporelles (0.000-0.007 kgm$^{-2}$s$^{-1}$). Complémente d’information du lidar Raman, les profiles de flux de masse ascendant normalisés suggèrent que l’inclinaison du profile de flux de masse ascendant est inversement proportionnel à l’humidité environnementale. De façon similaire, la présence de couches stratiformes détachées coïncide avec l’humidification de l’environnement sous la couche d’inversion. Considérant ceci, l’analyse présentée ici est utile pour évaluer et restreindre la représentation des nuages et les paramétrisations de flux de masse dans les modèles numériques.
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Chapter 1
Introduction

Trade-wind cumuli are shallow clouds (few hundred meters thick) that form in the atmospheric boundary layer. As such, they are intimately tied to the surface their source of moisture and heat. Despite their low fractional coverage (cloud cover < 60 %), these clouds play a fundamental role in maintaining the thermodynamic budget of the lower troposphere. Specifically, at any level in the atmosphere, their formation through water vapor condensation release heat while their disappearance through evaporation absorbs heat. Trade-wind cumuli are ubiquitous across the subtropical oceans and in number out-rank any other cloud type [Warren et al., 1988; Norris, 1998]. Through the trades, they participate in providing the moisture needed to fuel deep convection in the intertropical convergence zones [Riehl et al., 1951] and subsequently the Hadley cell circulation [Siebesma and Cuijpers, 1995].

The lack of exhaustive marine cumuli observations complicate their adequate representation in numerical weather prediction and climate models [Kollias and Albrecht, 2010; Rauber et al., 2007]. According to the Fourth Intergovernmental Panel on Climate Change (IPCC) reports, cloud feedbacks are one of the largest sources of uncertainty in predicting future climate change [Solomon et al., 2007]. More specifically, current disagreements of quantities as simple as cloud fraction may account for warming of the same order of magnitude as that of the increase of greenhouse gases [Stephens, 2005]. Tiedtke [1989] also demonstrated that including shallow convection in simulations is essential to represent realistically surface evaporation, precipitation and even the Hadley circulation.

Models do not posses fine enough grid resolution to resolve all the scales of boundary layer turbulence and thus cannot explicitly represent all processes and must rely on parameterizations. A parameterization is a numerical relationship that describes the mean behavior of a process happening over a “large” domain. For parameterizing shallow cloudiness and subsequently estimate thermodynamic fluxes, the bulk mass-flux (\(M, \text{kgm}^2\text{s}^{-1}\)) based approach is very appealing since it embeds a cloud ensemble into a one-dimensional plume and
represents turbulent transport using air density \( (\rho, \text{kgm}^{-3}) \), draft magnitude \( (w, \text{ms}^{-1}) \) and its fractional areal coverage \( (a, \text{m}^2) \) \cite{Tiedtke, 1989}.

\[ M = \rho wa \quad (1) \]

It is used in combination with the difference of fractional entrainment rate \( (\varepsilon; \text{inflow of environmental air in the cloud}) \) and detrainment rate \( (\delta; \text{outflow of cloudy air into the environment}) \) to describe the vertical gradient of mass transport \( (\partial M / \partial z) \) in a cloud ensemble \cite{de Rooy and Siebesma, 2008}.

\[ \frac{\partial M}{\partial z} = (\varepsilon - \delta)M \quad (2) \]

Expressions for these rates undergo active research. Physically, mass flux needs to reach zero at cloud top. Considering a situation where buoyancy and vertical velocity do not change, this calls for smaller (higher) detrainment (entrainment) rates for deeper cloud layers. Moreover, these rates should take into account the environmental conditions for example: In a humid environment, the entrained air and cloud mixture reach equilibrium quickly leading to less evaporative cooling and subsequently slower reduction of the mass flux with height \cite{de Rooy and Siebesma, 2008}. Nevertheless, fractional entrainment and detrainment rates are often prescribed as constant \cite{Siebesma and Holtslag, 1996} and only few studies consider the effects of cloud-layer depth and environmental humidity \cite{Kain and Fritsch, 1990; Derbyshire et al., 2004; Bechtold et al., 2008}]; All of which have yet to be validated by mass flux profiles derived by observations.

Marine cloud observations are sparse due to the difficulty to capture their array of undisturbed properties from steady land-based platforms. Land-base platforms biases may include 1) the presence of orographic effects which can be seen as systematic upwards or downwards mean velocities 2) a fictitious diurnal cycle caused by the low-heat capacity of land versus ocean \cite{Kollias and Albrecht, 2010} 3) abnormally small cloud droplets generated by urban pollution which increases the condensation nuclei population and tends to reduce the capacity of radars to observe clouds \cite{Chandra et al., 2013}. Past ship-based campaigns collected
basic information at a coarse temporal resolution (~ 3 h) over short periods (~ 1 month) about the atmospheric column and the sea surface state (e.g. pressure, temperature, relative humidity and horizontal wind). This was infrequently supplemented by point measurement from aircraft of in-cloud properties during cumulus-topped boundary layer (e.g. BOMEX, 1969 [Holland and Rasmusson, 1973] and ATEX, 1969 [Stevens et al., 2001]). These early field campaigns were conducted well before the development of remote sensing instruments such as cloud radars and lidars that have come to be critical in the definition of cloud properties and thus lacked the vertical information required to characterize the depth and dynamics of the shallow cumulus. Nevertheless, they provided a detailed view of the tropospheric thermodynamic structure and surface fluxes and thus enabled some intercomparison studies [Siebesma et al., 2003]. This data also allowed the modeling of ideal cloud fields with Large Eddy Simulations (LES), which contributed to increase our understanding of cumulus cloud processes [Benistion and Sommeria, 1981 Nicholls et al., 1982].

The aforementioned modeling studies would benefit from robust, long-term observations of cloudiness and mass flux. Fortunately, the recently established island-based observing facilities at Azores (http://www.arm.gov/sites/ena) and Barbados (http://barbados.zmaw.de/) are equipped with state-of-the-art remote sensors and thus, such observational datasets are within reach. Cloud radars (active remote sensors) are suited to study shallow cloudiness since they can provide discrete vertical information at high temporal (10 s) and spatial (30 m) resolution. Cloud radars do not directly observe air motion and so nor can they be used to directly estimate mass flux; rather they record the Doppler velocity of targets similar in size to their wavelength. In the case of mm-wavelength radars the atmospheric targets are cloud and drizzle droplets. Doppler velocity is the component of motion of the targets relative to the instrument (a.k.a. towards or away from the instrument in the direction of the transmitted electromagnetic wave). When observing vertically, this motion can be decomposed as vertical air motion and droplet terminal velocity (the effect of gravity against drag). Since terminal velocity is negligible for small droplets, radar derived mass flux can be estimated using the mean Doppler velocity (VD) of small hydrometeors and assuming a constant density of 1 kg m\(^{-3}\) following from the Boussinesq approximation [Kollias and Albrecht, 2010; Protat et al., 2010; Kalesse and Kollias, 2013].
\[ M = VD_{\text{small droplets}}a \]  

This information can be complemented by Raman lidars (active remote sensors as well), which observe profiles of environmental thermodynamic properties (e.g. temperature and humidity).

Some important statistical features of these marine cumulus clouds have emerged recently in the literature. Their cloud bases persistently coincide with the lifting condensation level (Azores 710 ± 140 m [Ghate et al., 2011]; Barbados 700 ± 150 m [Nuijens et al., 2014]; Nauru 500-600 m [Kollias and Albrecht, 2010]). Kollias and Albrecht [2010], using 649 h of cloud radar data indicated that marine clouds do not exhibit a diurnal cycle. Nuijens et al. [2014] using 2 years of ceilometer data in Barbados also commented on the seasonal stability of the non-precipitating (n-p) cloud cover. In fact, the variability of the Barbados cloud field in particular can be attributed mainly to the presence of overlaying stratiform layers. Nuijens et al. [2014] discussed the possible presence of the cumulus valve mechanism, which describes a self-regulated shallow cumulus field. In short, increased cumulus cloudiness leads to increased entrainment of dry air that subsequently elevates the Lifting Condensation Level (LCL) and suppresses cloudiness. Alternatively, lower cloudiness causes inefficient moisture removal from the mixed layer, where the build up of relative humidity lowers the LCL and promotes cloud formation.

In this study, we seek to further document the vertical and temporal variability of the Barbados trade-wind cumulus cloudiness and dynamics (e.g. up/down draft organization, velocity, areal coverage and mass flux). In addition, we aim to investigate how cloudiness and mass flux relate to boundary layer thermodynamic structure. This is an observational study based on a three-month record of Raman lidar humidity and temperature retrievals and a two-year record of profiling cloud radar data that contains the largest record of radar-documented marine shallow cumuli to date. The presented work is organized as follows. Chapter 2 details the data processing steps, chapter 3 presents the statistical methods and results and finally, a discussion and comparison with literature is undertaken in chapter 4.
Chapter 2
Data processing

2.1. Remote sensing data collection and quality control

The Max Planck Institute for Meteorology at Hamburg in collaboration with the Caribbean Institute for Meteorology and Hydrology and several other institutions established in April 2010 the Barbados Clouds Observatory (BCO, Stevens et al., 2014). The BCO is located on Deebles Point on the east coast of the island and is exposed to the undisturbed trade wind cumulus field.

The main dataset used in this study is from the 35-GHz profiling cloud radar (KATRIN) during 2012-2013 (402 days; Fig. 4e), which is a significantly longer marine cumulus data record than was previously found in literature [Ghate et al., 2011; Kollias and Albrecht, 2010; Wang and Geerts, 2013]. KATRIN provides time-height observations of the atmospheric column with a temporal resolution of 10 sec and a vertical resolution of 30 m. Shallow cumuli typically contain low amounts of liquid water content and small droplet sizes making them challenging to study using radars. At a sensitivity of -45 dBZ at 2 km range (typical sensitivity for the US Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) profiling radars [Kollias et al., 2007; Moran et al., 1998]), clouds with liquid water path below 50 gm$^{-2}$ are missed especially over land [Chandra et al., 2013]. Thus, the exceptionally high sensitivity (-65 dBZ at 1 km range) and Doppler velocity resolution (0.002 ms$^{-1}$) of the KATRIN radar combined with the larger amounts of liquid water content and larger droplet sizes typically found in oceanic cumuli enables us to observe most, if not all, of the shallow cumulus cloud field.

The KATRIN radar data are subjected to the standard real-time data processing routine developed for the suite of 35-GHz radars produced by Meteorologische Messtechnik (METEK; Bauer-Pfundstein, ). Briefly, at each range gate, a slight variation of the Hildebrand and Sekhon [1974] technique determines the noise floor of the full Doppler spectrum information. Then, the radar Doppler spectrum is separated into its various peaks, if any, which are the contributions
from the different targets present in the observation volume (e.g. cloud, precipitation and clutter). The algorithm identifies targets using the linear depolarization ratio and Doppler velocity of each peak along with complementary temperature information from the nearest METAR station. For uncertain peaks, the information from the neighboring observations (in time-height) is used to consolidate the target classification.

Due to its proximity to the ocean and poor antenna cross-channel, the KATRIN radar side lobes capture signal from oceanic waves creating clutter in its lowest range gates. The location of this clutter band varies daily from 0.3 to 0.6 km and is not well filtered during the routine processing (E.g. Fig. 1b at ~ 0.5 km). A complementary filtering routine is applied to the radar moments to remove the sea-clutter. First, the lowest 510 m of KATRIN radar data are discarded. In addition, all radar echoes that span two range gates or less (60 m) or in clusters smaller than nine pixels are removed (Fig. 1c, green pixels). Given the position and the stability of cumuli cloud bases at Barbados (700 ± 150 m, [Nuijens et al., 2014]) this procedure is not expected to eliminate real cloud echoes since there is little overlap of the sea clutter echoes and cloud echoes. However, it does remove rain shafts below the cloud base, which are anyways not of interest in this study.

In a recent study, Ghate et al. [2011] commented on the difficulty to characterize the marine boundary layer (MBL) thermodynamic structure at the coarse resolution provided by soundings (every 6 hours). Here, a Raman lidar instrument, with a pulsed laser beam at a temporal and spatial resolution of 2 min and 6 m respectively is preferred. The ratio of the measured backscatter signals in the vibrational Raman branch of water vapor and nitrogen is used to retrieve water vapor mixing ratio at night (~ 8 pm to 6 am LT) while the pure rotational Raman lidar technique is employed to estimate the air temperature in the ultraviolet and visible portion of the solar spectrum. Error estimation is derived using Poisson statistics for lidar signals on a 2 h-averaging period. Data with error beyond 5 % are not considered in our analysis.
2.2. Cloud classification

In this section, the MBL clouds are isolated and categorized to subsequently investigate their relative contribution to cloud fraction as well as their microphysical (as depicted by radar reflectivity) and dynamical (as depicted by radar Doppler velocity) structure. After the application of the sea clutter rejection filter (section 2.1.), the first hydrometeor base (cloud, drizzle or remaining rain shaft) and top (cloud only) height is estimated (Fig. 1c, black circles indicate bases). Given the relative persistence of the LCL at Barbados the cloud base height serves as a basis to determine the cloud type. Also, we rely on the duration of the hydrometeor clusters rather than an estimate of their horizontal dimension since wind speed profiles are not available at all times. After surveying the radar observations and literature [Nuijens et al., 2014; Siebesma et al., 2003] four categories of MBL clouds are defined (Table 1). Figure 1c illustrates an example of the cloud classification. Hydrometeor clusters with bases between 1 and 3 km (clearly elevated relative to the LCL) lasting at least 2½ minutes are classified as stratiform layer (Str.). Cumulus with attached stratiform layers (Cu.Str.), analogous to the clouds observed during the ATEX field experiment, are hydrometeor clusters that have bases below 1 km for at least 5 minutes (the cumulus segment; clear box) and bases above 1 km for at least 2½ minutes (the stratiform segment; shaded box). Finally, hydrometeor clusters with mean cloud base height below 1 km form the cumulus cloud type (Cu). Clouds extending through the boundary layer (i.e. mean cloud base height below 3 km, mean cloud top height above 3 km and depth beyond 3 km) are classified as deep.

Table 1. Cloud type characteristics and occurrence.

<table>
<thead>
<tr>
<th>Cloud types</th>
<th>Str.</th>
<th>Cu.Str.</th>
<th>Cu</th>
<th>Deep</th>
</tr>
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<tbody>
<tr>
<td>Hydrometeor top Height (km)</td>
<td>&lt;3</td>
<td>&lt;3</td>
<td>&lt;3</td>
<td>&gt;3</td>
</tr>
<tr>
<td>Hydrometeor base Height (km)</td>
<td>[1 3]</td>
<td>&gt;1 &amp; &lt;1</td>
<td>&lt;1</td>
<td>&gt;2.5 &amp; &gt;5</td>
</tr>
<tr>
<td>Duration (min)</td>
<td>&gt;2.5</td>
<td>&gt;2.5</td>
<td>&gt;5</td>
<td>&gt;5</td>
</tr>
<tr>
<td>Number of clouds</td>
<td>617</td>
<td>2,620</td>
<td>35</td>
<td>40,564</td>
</tr>
<tr>
<td>Precipitating N-p</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>All available observations</td>
<td>7,171 h</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Continuous X h periods with n-p cumulus only</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3 h</td>
<td>1,003 periods = 3,009 h</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2 h</td>
<td>1,804 periods = 3,608 h</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1 h</td>
<td>4,142 periods = 4,142 h</td>
<td></td>
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</table>
Fig. 1. Three-hour period on 27-01-2012 of a) filtered Reflectivity b) unfiltered Doppler velocity (Positive indicates upward motion and negative indicates downward motion) c) cloud classification and filtered pixels (i.e. sea clutter, rain shafts and cloud segments of negligible size) [Green: filtered pixels, red: stratiform, orange: n-p cumulus, light blue: precipitating cumulus, dark blue: precipitating cumulus with stratiform outflow]. Also identified are cloud bases [back dots], an horizontal line at 1 km marking the divide between the stratiform and cumulus bases, and the cumulus [clear box] and stratiform [shaded box] portion of the cumulus with stratiform outflow.

To demonstrate that our rather subjective criteria generates two distinct classes of cumulus (with/without stratiform outflow), two joint probability distribution functions of hydrometeor base height versus thickness are produced (Fig. 2). Be aware that these joint probability distribution functions only consider precipitating clouds given that only them produce stratiform outflows (Table 1) and that all cumulus thicknesses may be slightly overestimated due to rain shaft residues post filtering (see section 2.1. for rain shaft filtering and section 2.4. for precipitation criteria). The joint probability distribution function of cumulus with stratiform outflow shows two clusters; one comprising bases below 1 km and one comprising bases beyond 1.1 km (Fig. 2b). The cluster of elevated bases is consistent with the stratiform
definition (thin, < 1 km). Furthermore, the thickness of the stratiform layers decreases with increasing cloud base height suggesting a stable top constrained by an inversion. The second cluster, the cumulus segment, is deeper with hydrometeor column thickness interquartile range (IQR) 0.8 to 1.8 km. The cumulus without stratiform outflow have multimodal thicknesses but are overall thinner IQR 0.5 to 1.5 km moreover they do not exhibit an elevated cluster which confirms that our criteria produces two distinct cumulus classes. For completeness, note that the low fraction of hydrometeor bases detected above 700 m can be attributed to thinner cloud boundaries.

Fig. 2. Joint probability distribution of thickness (bin size 100 m) and hydrometeor base height (bin size 100 m) for a) precipitating cumulus bases and b) precipitating cumulus with attached stratiform layers. The median thickness for each hydrometeor base height bin [circle] and its interquartile range [IQR, whiskers] are included.

2.3. The influence of precipitation on radar observables

In addition to cloud fraction, the KATRIN mean Doppler velocity is used to retrieve the vertical air motion and estimate the cumuli up/down draft magnitude, fractional coverage and
mass flux. Since cloud droplets contribute negligibly to the mean Doppler velocity (0.3 and 7 cm s\(^{-1}\) for 10 and 50 \(\mu\)m droplet respectively; effectively an order of magnitude smaller than vertical air motion in shallow clouds) they can be used as air tracers. However, drizzle droplets and raindrops have significant sedimentation velocity (0.2 to 9 ms\(^{-1}\)), and thus induce a downward bias to the observed mean Doppler velocity and need to be removed from any vertical air motion analysis [Kollias and Albrecht, 2010].

Here, the detection of precipitation size particles in the radar sampling volume is accomplished in a statistical way using the observed radar reflectivity factor versus mean Doppler velocity (Z-V) relationship. The Z-V analysis relies on the assumption that the vertical air motion, over a long enough time period, averages to 0 ms\(^{-1}\) any systematic negative (downward) trend is attributed to the contribution of large hydrometeors fall velocity. This approach has been implemented in the past in broken shallow cumuli [Kollias and Albrecht, 2010] and stratiform layered clouds [Protat et al., 2010; Kalesse and Kollias, 2013]. Using the KATRIN data, we explore the Z-V relationship of our entire shallow cumuli dataset. The analysis reveals that aside from the expected downward trend at high radar reflectivity (beyond -20 dBZ) an upward trend emerges around -20 dBZ. In order to further investigate this trend, the Z-V relationship is reproduced for three distinct heights (660, 1090 and 1530 m) that correspond to the average height of the base, middle and top of the cumulus clouds analyzed in this study (Fig. 3, left panels). At 660 m, the majority of the observed mean Doppler velocities are distributed around 0 ms\(^{-1}\) for reflectivity lower than -30 dBZ. At 1090 m, the distribution of the mean Doppler velocities slowly shifts upwards with increasing radar reflectivity. This is mostly visible in the -30 to -20 dBZ range. Section 3.2 contains a discussion about the location of each reflectivity range within individual clouds; In short, the upper centermost part of the n-p cumulus clouds contains this reflectivity range collocated with the most vigorous updrafts. At radar reflectivity higher than -20 dBZ, the particle sedimentation controls the distribution of observed mean Doppler velocities. Finally, at 1530 m, the upward trend is less pronounced since the n-p cloud population sampled decreases at the expense of the precipitating clouds.

Based on the previous analysis, a threshold of -20 dBZ is selected to distinguish radar echoes affected by precipitation size particles. Subsequently, a cloud containing at least four
radar echoes with reflectivity above -20 dBZ and at the same time Doppler velocity below -0.5 ms\(^{-1}\) is considered precipitating. Similar thresholds that have been proposed in the literature (e.g. [Frisch et al., 1995, Liu et al., 2008, Ghate et al., 2011]). Kollias and Albrecht [2010] relied on a reflectivity threshold between -22 and -15 dBZ, but removed only the precipitating segments of clouds rather than excluding entire precipitating clouds. Thus, the “precipitating cloud filter” proposed here is aggressive and may remove precipitation that has yet to leave the cloud and cannot be detected by usual ceilometer, microrain radar or disdrometer techniques. This, however, ensures unbiased vertical velocity statistics and prevents mass flux underestimations. Fortunately, the shear number of available clouds (tens of thousands) permits the use of a conservative n-p cloud definition without compromising the robustness of the derived statistics (Table 1). The subset of Z-V observations that pass through the precipitation filter is shown in figure 3 (right column). The total frequency of occurrence of each of the panels is calculated as the number of occurrences of the n-p clouds divided by the occurrences of all clouds observations at the corresponding height.

Fig. 3. Joint probability distribution of Reflectivity-Velocity across all cumulus observations at three levels. Left panels: All cumulus clouds observed, right panels: non-precipitating (n-p) cumulus clouds. Bottom panels: 660 m, middle panels: 1,090 m, and top panels: 1,530 m. The black line indicates the mean value. The total frequency of occurrence of each right panel is calculated as the number of occurrences of the n-p clouds divided by the total number of occurrence of all cumuli at the corresponding height.


Chapter 3
Statistical methods and results

3.1. Daily hydrometeor fraction time-series analysis

Hydrometeor fraction is one of the most fundamental descriptors of the cloud field. Here, we estimate the hydrometeor fraction at each KATRIN range gate as the amount of hydrometeors (cloud and/or precipitation) detected relative to the entire sky (clear and cloudy/precipitating) within a day. It is not to be confused with cloud cover, which is its projection on the surface. The daily hydrometeor fraction profiles along with the KATRIN data availability for the 2 years of BCO observations analyzed here are shown in figure 4 along with the interquartile ranges derived from the 2-years analyzed. Note that radar data availability takes a value of 0 % if no observations were collected in a day and 100 % otherwise (independent of the presence or absence of clouds). Furthermore, a 3-day long running mean filter is applied to the daily hydrometeor fraction and data availability to smooth out extremes. Be aware that the radar data record contains extensive periods (from 15 days to 4 months) with no radar data (Fig. 4e).

The daily profiles of hydrometeor (cloud and/or rain) fraction reveal the omnipresence of MBL clouds in the lowest 3 km (Fig. 4a). Deeper hydrometeor layers that extend to the top of the troposphere and associated cirrus clouds have a noticeable seasonal cycle with a maximum in June-November (reaching 50 % occurrence) and a minimum in January-March. The highest cloud tops reach 14-15 km and their extent exhibit small seasonal variability.

The contribution of the three main MBL cloud types to the total observed MBL hydrometeor fraction (Fig. 4a) is shown in Fig. 4b,d. The interquartile range (IQR) of the observed daily hydrometeor fraction from each cloud type is shown in Fig. 4I,II. Figure 4I shows that the most frequent precipitating cloud type at BCO is cumulus with no stratiform outflow with a peak fraction of 2.3 % at 809 m. They are more frequently found between 540-1542 m. In contrast, precipitating cumulus with stratiform outflow have an elevated maximum fraction of 1
% at 1679 m that typically marks the height of their stratiform layer. Finally, unattached stratiform clouds contribute 0.3 % of the cloud fraction. Over the 2-years sampled, both precipitating cumulus cloud categories show similar variance (IQR 3 %). Looking at the daily time series, the two precipitating cumulus classes seem to occur concurrently throughout the year (Fig. 4b,c). There is a noticeable absence of precipitating boundary layer clouds in September-November 2013 (no data in available in 2012) and higher occurrence of stratiform layers in the January-March of both years. Furthermore, the presence of precipitating clouds is intermittent with periods of 3-5 days of precipitating conditions followed by dry conditions.

The n-p clouds present at Barbados are predominantly cumulus (Fig. 4II). Their 2-year average cloud fraction peaks at 4.5 % at 660 m and is “maintained” between 660-839 m. Over the 2-years sampled, this peak oscillates little with an IQR of ~3 % (Fig. 4II). Overall, the n-p cumuli persist through out the year and exhibit a weak seasonal cycle with minimum cloud fraction and vertical extent during the summer period and maximum during the winter period (Fig. 4d). Their weak seasonal cycle is in opposite phase with that of the deeper and cirrus cloud systems (Fig. 4a).
Fig. 4. Two-year time series of a) daily hydrometeor fraction in the troposphere for all hydrometeors (clouds and precipitation) and daily hydrometeor fraction in the boundary layer per cloud type b) precipitating cumulus with stratiform outflow, c) precipitating cumulus d) n-p cumulus; also included e) binary radar data availability; All following the 3-day smoothing. To the right are the interquartile range (IQR) of hydrometeor fraction (from the 2-year time series) per boundary layer cloud category I) Precipitating clouds and II) Non-precipitating clouds; Both using the unsmoothed dataset. In the n-p clouds panel (II) cloud fraction instead of hydrometeor fraction is indicated, since in the absence of precipitation the only hydrometeors present are clouds.

3.2. The internal structure of individual cumulus clouds

Radar reflectivity and Doppler velocity can also be used as descriptors of the cloud field. In the absence of precipitation, reflectivity is related to liquid water content and Doppler velocity to vertical velocity. In the presence of precipitation, these relationships are more complex nevertheless, reflectivity and velocity be used to make educated guesses on precipitation strength. Thus high-resolution composite of the internal structure of cumulus clouds were generated using the KATRIN radar observations. The following analysis is presented with the caveat that the use of a profiling remote sensing suite may result to clouds being sampled along different chord length that do not represent their diameter. The availability of a large number of
sampled clouds (hundreds and more depending on the cloud type; Table 1) should help to minimize possible biases.

The KATRIN radar observations are initially provided in time-height coordinates (profiling). The maximum vertical and temporal extent of each cumulus are used to create a normalized coordinate system [-1:1, -1:1] were -1 and 1 correspond to the first and last cloud detection in time and height respectively and as such (0,0) corresponds to the center of the cloud. Then using a spacing of 0.05 normalized units a 21 x 21 gridded clouds is created by interpolating the KATRIN radar data. In the case of cumulus with stratiform outflow, only the cumulus segment is considered (E.g. Fig. 1b, clear box). Using this normalization framework, all members of a particular class of MBL cloud can be overlapped together to produce, for each grid point individually, percentiles of radar reflectivity and mean Doppler velocity (Fig. 5). The plotted percentiles do not correspond to any one cloud but rather represent the median behavior of all clouds in each grid location. In addition, the reflectivity and velocity percentiles do not correspond but are overlayed to save space. 40,576 n-p cumuli, 2,620 precipitating cumuli and 617 precipitating cumuli with stratiform layers were normalized, which represents the largest oceanic cumulus cloud sample to date (Table 1).

The n-p cumulus clouds contain an elevated reflectivity core (IQR [-40 -28] dBZ) centered near the cloud top (Fig. 5 top row, colormap). As we progress towards higher quartiles (Fig. 5 top row from left to right), the elevated reflectivity core expands vertically and horizontally. At each quartile an average vertical increase of 15 dBZ is observed from the cloud base to the center of the reflectivity core. In parallel, a 10 dBZ horizontal increase in the radar reflectivity is observed from the lateral boundaries to the core. In terms of vertical air motion, the n-p cumulus clouds contain a less defined elevated core (IQR [0.65 0.00] ms^{-1}) (Fig. 5 top row, contours). This feature maintains its shape and a 1 ms^{-1} vertical gradient across the quartiles. Contrastingly, the velocity is constant on the vertical boundaries. The 25th and 50th percentiles show strictly upward vertical motion while the 75th percentile shows a band of downward motion concentrated on the vertical boundaries.

The two categories of precipitating clouds (without and with stratiform outflow) have
similar reflectivity structures (Fig. 5 middle and bottom row, colormap). The differences between them manifest in the magnitude of the overall reflectivity. Roughly, from the 25th to 75th percentiles cloud with an outflow are 5, 7 and 11 dBZ stronger. The precipitating clouds initially (at the 25th percentile) appear similar to the 75th percentile of the n-p clouds since they contain an elevated reflectivity core, (25 and 21 dBZ for the cloud without and with an outflow respectively) yet their vertical gradient is weaker (+10 dBZ). Interestingly, as we progress towards higher percentiles this isolated core and associated gradient eventually vanish (the vertical gradient from the base to the core changes from +6 dBZ at the 50th percentile to 0 dBZ at the 75th percentile). In terms of velocity (combination of air motion and hydrometeor fall velocity), both precipitating cloud categories contain only downward motion. These are strengthening vertically from cloud top to base and horizontally towards the center part of the clouds. The velocity differences between the cloud types appear in the magnitude but also in the vertical gradients. Clouds with outflows show downdraft velocities about 0.20 ms\(^{-1}\) stronger at cloud top and contain vertical gradients about 30\% stronger.

Fig. 5. Normalized cloud Reflectivity [colormap] and Doppler velocity [contours] composites. Left panels) 25th percentile, middle panels) 50th percentile, right panels) 75th percentile. Top row) non-precipitating cumulus, middle row) precipitating cumulus, bottom row) cumulus segment of the precipitating cumulus with stratiform outflow.
3.3. Non-precipitating cumulus bulk mass flux

In this section, we isolate the updraft and downdraft structures and quantify their respective organization and contribution to mass transport. Mass flux is a common and convenient way to describe cumulus clouds in bulk combining information about microphysics (density) and dynamics (vertical velocity).

It is important to note that the mass flux analysis is limited to n-p cumuli since it requires that the observed radar Doppler velocities can be considered as a proxy for air motion. Vertical velocity and mass flux statistics will be reported in bulk (3 h averaging periods). Only 3-h periods that contain n-p cumulus only are considered since the presence of precipitating clouds would reduce the space available for n-p cumulus drafts and create a mass flux reduction bias. The use of 3 hour averaging periods should provide a large enough data sample to average out the life cycles of individual clouds and produce robust statistics. This averaging time could be roughly converted to horizontal dimension using an average wind speed of 8 ms\(^{-1}\). The average 86.4 km horizontal domain is expected to comfortably accommodate the grid size of any modern General Circulation Model.

A total of 1003 3 h-periods (i.e. 3009 h; Table 1) satisfied this criterion, which is the largest available dataset to-date for oceanic shallow cumuli and should be sufficient to derive robust statistics. Using the radar Doppler velocity as vertical air motion indicator we estimate at each height above the 3 h-averaged cloud base height the mean updraft velocity (\(w_{up}\); estimated by all points with \(w > 0 \text{ ms}^{-1}\)), the updraft area (\(a_{up}\); estimated as the number of points in 3 h with \(w > 0 \text{ ms}^{-1}\) divided by the total number of observations cloudy or not in 3 h) and the updraft mass flux (\(M_{up} = a_{up} w_{up}\)). The corresponding profiles for downdrafts (\(w < 0 \text{ ms}^{-1}\)) are estimated. Note that traditionally mass flux is calculated using the draft velocity departure from the mean atmospheric velocity. However given that the radar only samples the cloudy part of the atmosphere, we have to rely on the usual assumption that mean atmospheric vertical air motion at every level is 0 ms\(^{-1}\) [Ghate et al., 2011; Kollias and Albrecht, 2010]. For this technique, hereafter named direct sampling, the median (black line) and IQR (shaded region) of the 2-year sample are presented in Fig. 6 as a function of height above cloud base. One panel is presented
per parameter (M, Fig. 6a; a, Fig. 6b; w, Fig. 6c). In each panel, the right portion (positive x-axis) is dedicated to the updrafts while the left portion (negative x-axis) is dedicated to the downdrafts.

Furthermore, the contribution of coherent (in time-height) updraft structures is investigated. To account for tilted updrafts as described in literature [Ghate et al., 2011], in time-height, each cluster of velocity echoes above the velocity threshold (w > 0 ms\(^{-1}\)) larger than 15/25 pixels is defined as a coherent structure. Two size thresholds (15+ and 25+) are used to illustrate the lack of dependency of our results on the size threshold used to define a coherent eddy. Two additional velocity thresholds (0.5 and 1.0 ms\(^{-1}\)) are included to derive the statistics of more intense coherent updrafts. For the coherent updraft sampling, only the median (colored lines) of the 2-year sample is presented in Fig. 6. The various velocity thresholds take different colors (0 ms\(^{-1}\), red; 0.5 ms\(^{-1}\), blue; 1 ms\(^{-1}\), pink). While, the various size thresholds are assigned different line types (15+ pixels, solid; 25+ pixels dashed). For references, table 2 contains tabulated values of draft mass flux, area and velocity for the direct sampling and the coherent updrafts (15 pixels, 0 ms\(^{-1}\)) at key heights.

From figure 6c, the direct draft velocity (updrafts and downdrafts, black lines, respectively on the positive and negative x-axis) is initially invariant with height in the first 150 m from the cloud base. Above this level, the draft magnitudes increase rapidly to reach their maximum value just below cloud top. Overall, at all levels, the updraft magnitude is higher to that of downdrafts and the width of their distributions increases with height. The shape of the coherent updraft velocity profiles (colored lines) is similar to that of the direct sampling but their magnitude is greater reflecting the velocity thresholds used.

The shape of the direct updraft fractional area profile (Fig. 6b) follows the n-p cumuli cloud fraction profiles (shown in Fig. 4II), where most of the clouds are concentrated within a 0.3 km thick layer. The updraft area profile peaks 0.09 km above cloud base and decreases rapidly to zero 0.57 km above the cloud base height. The area occupied by the coherent structures peaks higher as the velocity threshold chosen increases in magnitude (depicted by the vertical translation of the peak from the red curve to the blue curve to the pink curve). The 0.0,
0.5 and 1.0 ms\(^{-1}\) coherent updrafts profiles peak at heights 0.09, 0.09-0.15, 0.18-0.27 km with values 0.032, 0.012, 0.003 respectively. Notice how the coherent updraft structures do not span the entire depth of the cloud layer. At the cloud base and top, updrafts with magnitude higher than 1 ms\(^{-1}\) are not present. Alternatively, the direct downdraft area profile “maxima” is not very pronounced. The direct updraft area is systematically larger and more variable than that of the direct downdraft area (IQR ~0.05 vs. ~0.02). This supports the results of the 2D analysis of the previous section: In n-p clouds, downdrafts form a narrow shell of invariant width while updrafts occupy the central portion of the clouds that varies proportionally to cloud size.

Finally, we find that the shape of the direct mass flux profiles (Fig. 6a, black lines), which is calculated each 3-hour as the product of the mean area and velocity profiles, is mainly influence by the area profile (Fig. 6b; black lines). The direct updraft mass flux peaks at 0.09 km above cloud base at a value of 0.020 kg m\(^{-2}\) s\(^{-1}\), which coincides with coherent updraft (w > 0 ms\(^{-1}\)) transport of 0.019 kg m\(^{-2}\) s\(^{-1}\) or 95% of the direct mass flux. The mass flux from other coherent updrafts peaks at different heights, 0.5 ms\(^{-1}\) at 0.09 km with 0.012 kg m\(^{-2}\) s\(^{-1}\) and 1 ms\(^{-1}\) at 0.24 km with 0.004 ms\(^{-1}\).
Profiles of a) mass flux, b) area, c) vertical velocity above cloud base for n-p cumulus using 3 h averaging periods. Results from direct sampling (considering all updrafts and downdrafts) are in black (solid line: median, envelope: 25th-75th percentiles). Colored solid lines show results from sampling small coherent updraft (15 pixels or more), the different colors represent the various vertical velocity thresholds used to capture stronger updrafts [solid lines: median; 0 ms\(^{-1}\) (red), 0.5 ms\(^{-1}\) (blue) 1 ms\(^{-1}\) (pink)]. Similarly, colored dashed lines represent results from sampling large coherent updrafts.

Table 2: Tabulated percentile values of mass flux, area and velocity at key height levels from the profiles illustrated in Fig. 6.

<table>
<thead>
<tr>
<th></th>
<th>Mass flux (kgm(^{-2})s(^{-1}))</th>
<th>Area</th>
<th>Velocity (ms(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Updraft 50(^{th}) Coherent Updraft 25(^{th})/75(^{th}) Downdraft 50(^{th})</td>
<td>Updraft 50(^{th}) Coherent Updraft 25(^{th})/75(^{th}) Downdraft 50(^{th})</td>
<td>Updraft 50(^{th}) Coherent Updraft 25(^{th})/75(^{th}) Downdraft 50(^{th})</td>
</tr>
<tr>
<td>At the updraft mass flux peak (0.09km)</td>
<td>0.020 0.019 0.003</td>
<td>0.008/0.036 0.000/0.007</td>
<td>0.037 0.032 0.010</td>
</tr>
<tr>
<td>At cloud base</td>
<td>0.012 0.011 0.001</td>
<td>0.005/0.024 0.000/0.004</td>
<td>0.021 0.019 0.005</td>
</tr>
<tr>
<td>Max. value (Height km)</td>
<td>0.020 0.019 0.005</td>
<td>0.09 0.15 0.15-0.24</td>
<td>0.037 0.032 0.010</td>
</tr>
<tr>
<td>Min. value (Height km)</td>
<td>0 0 0</td>
<td>0 0 0</td>
<td>0 0 0</td>
</tr>
</tbody>
</table>
Rather than the direct sampling percentiles (black lines and grey shading in Fig. 6), the full histograms of the parameters (M, a, w) at the height of the mass flux maximum (0.09 km) are shown in figure 7. Keep in mind that these results are produced using 3 h statistics over a 2-year dataset (Fig. 7; blue lines). In order to investigate the stability of the mass flux on various scales we also reproduced these statistics using 2 h and 1 h averaging periods (Fig. 7; green and red lines respectively). The three distributions are very similar; the only discrepancy is a slight variance increase with decreasing averaging period.

All together, the updraft mass flux exhibits significant spread (Fig. 7b). The downdrafts mass flux spread is smaller and contains a preferred mode at 0-0.005 kg m\(^{-2}\) s\(^{-1}\) (Fig. 7a). The observed difference in the spread of the updraft mass flux is also present in the updraft area coverage (Fig. 7c,d) and overall the updrafts are wider than downdrafts (Fig. 7c,d). The updraft velocity distribution is nearly symmetrical about 0.5 ms\(^{-1}\) while the downdraft velocity distribution peaks at -0.2 ms\(^{-1}\) and has a tail elongating to about -1.3 ms\(^{-1}\) consistent with some narrow downdrafts actually more vigorous than the wider updrafts (Fig. 7e,f).

![Graphs showing distributions of updraft and downdraft parameters](image)

Fig. 7. Distribution at the mass flux peak (0.09 km above cloud base) of updraft a) mass flux, c) area, e) velocity and downdraft b) mass flux, d) area, f) velocity for non-precipitating cumulus periods using three different averaging periods [1 h (red), 2 h (green), 3 h (blue)].
3.4. Relationships to environmental thermodynamic profiles

Now that we have described cumulus clouds on various scales and quantified their variability, we wish to identify some drivers of this variability. Given that vertical and lateral mixing of cloudy and environmental air is an accepted mechanism in cumulus clouds, it is reasonable to hypothesize that environmental variability could explain some of the cloud variability.

First, the environment is represented using profiles of relative humidity and temperature estimated within a one-hour window of each precipitating cloud (Nighttime only; Fig. 8). The temperature profiles are virtually identical and the temperature inversion is not pronounced. Adversely, the relative humidity inversion is more distinct and this is where differences between the two cloud types arise. The presence of stratiform layers in the 1.5-2 km region coincides with the presence of increased environmental moisture.

Fig. 8. Thermodynamic profiles averaged within an hour of a precipitating cumulus [red] or precipitating cumulus with stratiform outflow [blue] of a) relative humidity and b) temperature. Solid lines are median values and shaded regions represent the 25th and 75th percentiles. Also included are horizontal dotted lines and dashed lines respectively most occurring cloud top and base.
Second, mass flux has an inherent relationship to both thermodynamics and cloud layer depth in parameterization schemes. In order to isolate the effect of environmental conditions, de Rooy and Siebesma [2008] proposed to study non-dimensional mass flux as a function of non-dimensional height.

\[ \bar{M} = \frac{M}{M_b} \quad (4) \]

\[ \bar{Z} = \frac{Z - Z_b}{Z_t - Z_b} \quad (5) \]

Where the subscript \( b \) refers to the height with maximum updraft mass flux and \( t \) refers to the first height with zero mass flux (i.e. cloud top).

Additionally, de Rooy and Siebesma [2008] parameterized the updraft mass flux in two parts, below and above the normalized center (\( \bar{Z} = 0.5 \)). To emulate this technique with simplicity, we used a linear relationship from the maximum updraft mass flux (\( \bar{Z} = 0 \)) to the center of the profile (\( \bar{Z} = 0.5 \); lower slope) and from the center of the profile to the top (\( \bar{Z} = 1 \); upper slope). As a result, the updraft mass flux magnitude in the center of each profile dictates both slopes. We find that this linear approximation is accurate (\( R^2 > 80 \% \)) in 47.8 \% of the cases for the upper slope and 58.8 \% of the cases for the lower slope. However, even if often imprecise, this approximation conserves the inherent trends of the data (i.e. concave profiles have the highest linear slope magnitude).

Recall that our analysis was previously limited to periods containing no precipitating clouds since they were responsible for a reduction in the potential n-p draft horizontal area coverage. This reduction in draft area would directly impact the potential maximum mass flux magnitude. However, in this section, since the profiles are non-dimensional (the maximum mass flux is always 1), this bias is alleviated. Thus, all 2,390 3 h-periods (i.e. 7,717 h; Table 1) of radar observations were used to produce a joint probability function of n-p cloud bulk (3 h averaged) normalized updraft mass flux with normalized height (Fig. 9b). The normalized updraft mass flux profile slopes are computed as the change in normalized updraft mass flux per
0.1 unit normalized height for $0<\bar{z}<5$ ($d\bar{M}/0.1d\bar{z}$). The distribution of the normalized updraft mass flux slope in the lower half of the cloud layer is presented in Fig. 9a. Note that the lower and upper slope distributions would be symmetrical since both are linear trends of the normalized mass flux encroached at the center of the cloud layer.

The probability density function, its median and IQR (black circle and whiskers) give us a comprehensive visual of the normalized mass flux profiles of the entire population and can be compared to work presented in previous studies (Fig. 9b; Also see chapter 4). The normalized updraft mass flux at the center of the cloud layers, critical to determine the slopes, varies around a median value of 0.38 (IQR [0.24 0.53]). The majority (71.8%) of the observed non-dimensional mass flux profiles are concave ($d\bar{M}/0.1d\bar{z} < -0.1$ in the lower part of the cloud layer) and as a result the lower slope distribution peaks at $-0.13$ $d\bar{M}/0.1d\bar{z}$ (Fig. 9a). Alternatively, 8.7% of the observed cases are linear ($-0.1$ $d\bar{M}/0.1d\bar{z}$ slopes) and the remaining 19.5% convex.

Fig. 9. a) Distribution of the normalized updraft mass flux profile slope in the lower half of the cloud layer. b) Joint probability function of normalized updraft mass flux per normalized height. Included are the 0.1 slope [dotted line], the normalized updraft mass flux median [circle] and the interquartile range [whiskers] at each normalized height.
In order to investigate the relationship between the shape of the updraft mass flux profiles and the environmental thermodynamic profiles, the lower slope distribution is separated in 3 sections: -0.3 to -0.7, -0.9 to -0.11 and -0.14 to -0.19 \( \frac{d\bar{M}}{0.1d\bar{Z}} \). The corresponding thermodynamic profiles retrieved by the Raman lidar during nighttime operations are averaged to produce Fig. 10. Figure 10a shows that there is a progressive increase in mean relative humidity with decreasing slope, which corroborates the theory that the higher the ambient relative humidity the less effective mixing is in reducing the updraft mass flux profile. More robust results could potentially be achieved once the complete Raman lidar dataset is released.

Fig. 10. Thermodynamic profiles for different intervals of normalized updraft mass flux profile slope in the lower half of the cloud layer a) relative humidity and b) temperature. Solid lines are median values and shaded regions represent the 25th and 75th percentiles. Also included are horizontal dotted lines and dashed lines respectively approximate cloud top and base of the n-p cumuli.
Chapter 4
Discussion and relevance to previous studies

In this study, we take advantage of a long record (403 days across 2 years) of zenith pointing cloud radar observations at the Barbados Cloud Observatory (BCO) to describe the oceanic shallow cumulus cloud field. The description spans various physical and temporal scales (ranging from individual clouds to cloud ensembles and from daily to hourly statistics) and encompasses radar and “model-like” parameters (e.g. reflectivity, velocity, cloud fraction, and mass flux). Furthermore, we investigate the relationship between some of the cumulus cloud field variability and environmental thermodynamics as observed by the Raman lidar (3-month dataset). Throughout the analysis, we distinguish three marine boundary layer cloud types: Cumulus, stratiform and cumulus with attached stratiform segments and identify if individual clouds contain precipitation size droplets, which impacts the radar observables.

At the BCO, non-precipitating (n-p) cumulus contribute the most to the observed hydrometeor fraction (peak 4.5 % at 710 m) relative to precipitating cumulus without (peak 2.3 % at 809 m) and with attached stratiform segments (peak 1 % at 1679 m). All MBL cloud types experience similar temporal variability (IQR ~3%). A review of relevant literature indicates the lack of, consensus on the non-precipitating subtropical cumulus cloud fraction magnitude, and significant variability has been reported (e.g. Azores 5.7 %, [Ghate et al. 2011]; Nauru 15 %, [Kollias and Albrecht 2010]; Barbados 10 %, [Nuijens et al. 2014]). In part, the observed discrepancies could be attributed to the instruments and methodologies used in the various studies. The BCO Ka-band radar is one of the most sensitivity ground-based radars and thus is expected to detect most if not all radiatively significant shallow cumuli. It is plausible that our conservative definition of n-p cloud (section 2.3.) is partially responsible for the modest reported cloud fraction.

The shallow (maximum thickness ~ 750 m) n-p cumuli maintain a structure of elevated reflectivity and a positive mean vertical velocity core surrounded by a weak narrow subsidence shell. This is consistent with most efficient (less diluted by lateral mixing) moisture transport in
the center portion of the cloud and subsequent liquid water accumulation at the top of the cloud layer caused by condensational growth. The reported 15 dBZ reflectivity increase from cloud base to cloud core (reflectivity maxima) at the center of individual clouds is consistent with previously observed cloud averaged gradient [Ghate et al., 2011]. The observed positive median velocity (0.25 ms$^{-1}$) across the cloud approaches what Ghate et al. [2011] observed at the Azores (mean velocity around 0.33 ms$^{-1}$) and the description by Kollias and Albrecht [2010]: Clouds in Nauru experience rising motions across 80% of their height (mean velocity between 0.10-0.35 ms$^{-1}$). Also, the existence of subsiding shells was investigated in a modeling study by Heus and Jonker [2008] and further observed by aircraft-based studies [Heus et al., 2009]. Considering updrafts and downdrafts separately, their rapid increase in magnitude with height is also consistent with the literature and could be explained by strengthening of updrafts with height within each cloud as potential energy is converted to kinetic energy. In their BOMEX modeling study, Siebesma and Cuijpers [1995] succeeded at reproducing this feature in updrafts but generated velocity gradient steeper than reported here. Similarly, observations by Kollias and Albrecht [2010] indicate higher downdraft magnitude at cloud base and top (almost a factor of 2 stronger). Finally, and not surprising, we observed increasing width of the velocity distributions with height reflecting the increase in turbulent activity near the cloud tops [Kollias and Albrecht, 2010, Ghate et al., 2011].

Overall, we can say that the Barbados n-p cumuli are very organized and active [Stull, 1985]. In-cloud drafts are 78% updraft in the lower part of the cloud layer similar to that reported by Kollias and Albrecht [2010] (70%) but lower than modeled by Siebesma and Cuijpers [1995] (51%). Moreover, throughout most of the cloud layer, 86% of them are coherent which is significantly higher than previously found by Ghate et al. [2011] (~62%) using a slightly different definition. This finding is in agreement with the use of Large Eddy Simulations that attempt to explicitly represent mass transport in shallow cumulus by resolving only the main eddies. Our 15/25 radar pixels coherent structures correspond to about 3 x 5 (time x height) radar pixels or 240 m x 150 m (assuming a 8 ms$^{-1}$ horizontal wind). The location of the updraft area peak rather than the velocity maxima determines the location of the updraft mass flux peak which magnitude we report is median 0.020 IQR [0.008 0.036] kgm$^{-2}$s$^{-1}$. This value is directly comparable to that measured by Ghate et al. [2011] at the Azores. It is however quite lower than
the 0.09 kgm$^{-2}$s$^{-1}$ at Nauru reported by Kollias and Albrecht [2010] (using a density of 1.1 kgm$^{-3}$). It should be noted that cloud fraction is much larger in Nauru (~15%) and scaling the updraft mass flux with cloud fraction reconciles both results. Considering modeling studies, our results resemble that of Siebesma and Cuijpers [1995] using a grid resolution of 125 x 125 x 50 m. In this case, the combination of the aforementioned updraft fraction deficit and updraft velocity overestimation enabled the simulation of comparable mass flux. This shows the importance to compare all mass flux contributors to validate parameterizations. Moreover, our analysis indicates that the in-cloud downdrafts have insignificant mass flux and very weak vertical and temporal variance. The downdraft mass flux does not compensate for the in-cloud updraft mass flux. This suggests that most of the compensating downdraft motions occur in clear air and that in-cloud downdraft mass flux can be neglected [Siebesma and Cuijpers, 1995]. Additionally, the updraft mass flux profiles normalized to isolate the effect of environmental thermodynamic from cloud layer depth (Fig. 9b) may be compared to that de Rooy and Siebesma [2008, Fig. 9]. Even if the simulated cases were for continental fair cumuli, notice that both studies contain more concave than convex mass flux profiles and show a similar spread of values. Moreover, they share a similar mid-cloud layer normalized mass flux (0.3 vs. 0.38).

Part of the changes in cumulus cloud fraction and depth seem directly related to the previously acknowledged oscillation of the large-scale subsidence rate from positive in December-May to negative in June-November [Brueck, 2013]. Our analysis suggests that in December-April (during periods of large scale subsidence) deep cloud systems are suppressed at the expense of more and deeper cumulus clouds and higher occurrence of stratiform outflows. It is possible that the large-scale subsidence helps maintain the compensating clear air downdrafts around the vigorous non-precipitating cumulus [Siebesma and Cuijpers, 1995]. Moreover, this subsidence may help sustain the observed moisture inversion and underlying relative humidity enhancement that helps maintain the detrained stratiform layers up top deep precipitating cumulus.

Another factor that may impact the shape of the precipitating clouds is their growth stage. Our results show that precipitating cumulus clouds with a stratiform outflow are deeper, have
higher reflectivity and velocity magnitudes and gradients all of which are consistent with more
developed mature clouds.

Finally, using the Raman lidar observations a plausible relationship between the
environmental humidity and the rate of vertical decrease of the updraft mass flux in n-p cumuli is
observed. This relationship is consistent with the theory that clouds present in a moister
environment evaporate (i.e. reduce their mass flux) more slowly upon mixing with the
environment than the clouds present in dryer conditions.
References

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