Turbulence in the Marine Boundary Layer and Air Motions below Stratocumulus Clouds at the ARM Eastern North Atlantic Site

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ABSTRACT: Marine stratocumulus clouds are intimately coupled to the turbulence in the boundary layer and drizzle is known to be ubiquitous within them. Six years of data collected at the Atmospheric Radiation Measurement's (ARM) Eastern North Atlantic (ENA) site are utilized to characterize turbulence in the marine boundary layer and air motions below stratocumulus clouds. Profiles of variance of vertical velocity binned by wind direction (wdir) yielded that the boundary layer measurements are affected by the island when the wdir is between 90° and 310° (measured clockwise from the north from where air is coming). Data collected during the marine conditions (wdir < 90° or wdir > 310°) showed that the variance of vertical velocity was higher during the winter months than during the summer months because of higher cloudiness, wind speeds, and surface fluxes. During marine conditions the variance of vertical velocity and cloud fraction exhibited a distinct diurnal cycle with higher values during the nighttime than during the daytime. Detailed analysis of 32 cases of drizzling marine stratocumulus clouds showed that, for a similar amount of radiative cooling at the cloud top, within the subcloud layer 1) drizzle increasingly falls within downdrafts with increasing rain rates, 2) the strength of the downdrafts increases with increasing rain rates, and 3) the correlation between vertical air motion and rain rate is highest in the middle of the subcloud layer. The results presented herein have implications for climatological and model evaluation studies conducted at the ENA site, along with efforts to accurately represent drizzle–turbulence interactions in a range of atmospheric models.

KEYWORDS: Marine boundary layer; Drizzle; Clouds; Remote sensing

1. Introduction

Marine boundary layer (MBL) stratocumulus clouds cover vast areas over eastern subtropical oceans and persist over very long periods (Klein and Hartmann 1993). These clouds reflect much greater amount of solar radiation back to space as compared with the underlying ocean surface, while emitting longwave radiation similar to the ocean surface due to their high cloud-top temperatures. Hence these clouds have a net cooling effect on Earth's surface and are an important component of Earth's radiation budget. MBL stratocumulus clouds are intimately coupled to the turbulence in the boundary layer that transports enthalpy, moisture, and momentum away from the surface. MBL turbulence is primarily modulated by surface turbulent fluxes, radiative cooling at the boundary layer top, entrainment of the free tropospheric air into the boundary layer, wind shear, and drizzle (LeMone et al. 2019). Drizzle is known to be ubiquitous within and below stratocumulus cloud systems and has been shown to affect transitions to cumulus

clouds, cause thermodynamic decoupling within the boundary layer, and decrease subcloud-layer turbulence through evaporative cooling (Wood 2005, 2012; Jones et al. 2011; Ghate and Cadeddu 2019; Yamaguchi et al. 2017; Borque et al. 2018). Drizzle has also been shown to produce surface density currents (cold pools) thereby affecting the mesoscale cloud structure (Wilbanks et al. 2015; Wang and Feingold 2009).

Due to the large radiative impact of MBL clouds, it is necessary to accurately simulate them in Earth system models (ESMs) used for predicting the future climate. As the processes controlling cloudiness and microphysical and radiative properties occur at spatiotemporal scales much finer than those of the ESM, MBL clouds are parameterized using resolved scales variables. However, it is challenging for the ESMs to accurately simulate stratocumulus clouds, drizzle within them, and the drizzle–turbulence coupling below them (Takahashi et al. 2017; Zheng et al. 2017, 2020; Dong et al. 2021).

Despite the importance of covariability of drizzle and vertical air motion, it is still unclear whether most of the drizzle falls within updrafts or downdrafts, and how do drizzle rate and vertical air motion covary for different turbulence forcing. To address these and other questions related to marine boundary layer clouds, the Atmospheric Radiation Measurement's (ARM) Mobile Facility was deployed at the island of Graciosa (Portugal) for a 19-month period under the auspices of the Cloud, Aerosol, and Precipitation in the Marine Boundary Layer (CAP-MBL) project (Wood et al. 2015; Miller et al. 2016). After the conclusion of this project, a new permanent site was deployed on the island that became fully operational at the end of 2014 and

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was named the Eastern North Atlantic (ENA) site. In this study we utilize the observations made at the ARM ENA site to 1) characterize the annual and diurnal cycle of turbulence in the marine boundary layer and 2) quantify the covariability between vertical air motion and rain rate below the stratocumulus cloud base.

The data and instrumentation used in this work are described in the following section, followed by a section reporting the annual and diurnal cycle of turbulence and its forcing parameters. The analysis of covariability between vertical air motions and rain rates below stratocumulus clouds is described in section 4, and the article is concluded with a summary, discussion, and conclusions section.

2. Instrumentation and data processing

The ARM ENA site is located on the island of Graciosa (39°N, 28°W, 25 m MSL) in the Azores and has been operational since 2015. Data from this site have been extensively used to study processes related to marine boundary layer clouds including stratocumulus (Wood et al. 2016). The site experiences a broad range of weather, cloud, and drizzle conditions including different mesoscale organizations of stratocumulus clouds (Rémillard and Tselioudis 2015; Mechem et al. 2018; Lamer et al. 2019), making it an ideal site to study MBL turbulence and its coupling with stratocumulus clouds.

a. Instrumentation

The ARM ENA site has several instruments to make detailed measurements of aerosol, cloud, precipitation, as well as dynamic and thermodynamic fields (Wood et al. 2015). Discussed here are only the instruments used in the study. A vertically pointing Ka-band Doppler cloud radar operating at 35-GHz frequency [Ka-band ARM Zenith Radar (KAZR)] records the raw Doppler spectrum and its first three moments in copolarization and cross-polarization at a 2-s temporal and 30-m range resolutions. Only the data from copolarized mode were used in this study. The KAZR was calibrated by comparing it with the Ka-band Scanning ARM Cloud Radar that is calibrated using a corner reflector. Hence, the KAZR calibration is accurate within 3 dB (Kollias et al. 2020). A laser ceilometer operating at 905-nm wavelength recorded the raw backscatter and the first three optical cloud-base heights at a 15-s temporal and 30-m range resolution. A Doppler lidar that operates at 1.5- μm wavelength is also present at the site and records the backscatter and mean Doppler velocity at a 1-s temporal and 30-m range resolution (Newsom and Krishnamurthy 2020). Surface sensible heat flux (SHF) and latent heat flux (LHF) are calculated using the eddy covariance (ECOR) technique at 30-min temporal resolution. The ECOR reported fluxes are affected by the island heating, while the focus of this study is on marine clouds. Hence the surface SHF and LHF as reported by the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis model (ERA5) at the first grid point north of the island are also used in this study. The sensible and latent heat fluxes from ECOR are referred to as ECOR-SHF and ECOR-LHF, respectively, while those from ERA5 are referred to as ERA5-SHF and ERA5-LHF, respectively. A radar wind profiler (RWP) operating at 1290 MHz frequency records raw Doppler

spectra and the first three moments in vertically pointing and two oblique beams. These RWP moments are then used to calculate boundary layer winds at 10-min temporal and 50-m vertical resolution. Surface meteorological instruments recorded surface air temperature, humidity, pressure, and winds at 1-min temporal resolution. Radiosondes are launched at the site twice daily at 0000 and 1200 UTC, and they provide profiles of temperature, moisture, pressure, and winds. The height of the planetary boundary layer, assumed corresponding to the base of the boundary layer inversion in potential temperature, was calculated from the radiosonde data (Heffter 1980; Riihimäki et al. 2015). Visible satellite imagery and cloud-top temperatures around the site location from the Spinning Enhanced Visible and Infrared Imager (SEVIRI) on board the Meteosat Second Generation satellite were also used (Schmetz et al. 2002). The SEVIRI data were available at ~ 3 -km horizontal and 30-min temporal resolution.

Data collected between 1 January 2015 and 31 December 2020 are used in this work. The surface meteorological station, ceilometer, and radiosonde data were available during the entire period (2193 days). The ECOR fluxes were unavailable for 124 days, the KAZR data were not available for 257 days, and the Doppler lidar (DL) data were not available for 288 days. The satellite and ECMWF data were available for the entire duration. Hence, the available data were of sufficient duration to characterize the annual and seasonal cycle of turbulence at the site. The figure showing uptime of these instruments is in the online supplemental material.

b. Doppler lidar and ceilometer calibration

The returns from Doppler lidar and ceilometer below the cloud base are affected by aerosols, and hydrometeors. The ceilometer returns corresponding to hydrometeors are later used to derive the drizzle properties by combining them with data from the cloud radar (O'Connor et al. 2005). Drizzle properties can also be retrieved by combining data collected by lidars at different wavelengths (Westbrook et al. 2010a; Lolli et al. 2017). The DL returns from aerosols can be used for tracking air motion because aerosols act as passive tracers within the boundary layer eddies. Drizzle has been shown to be present $\sim 50\%$ of the time below boundary layer clouds at the ENA site (Rémillard et al. 2012). Hence, the first step is to calibrate both lidars and to develop a technique to objectively separate drizzle returns from those of aerosols.

Prior to the calibration, data from both lidars were filtered for noise using the method described by Kotthaus et al. (2016). In addition, the parameters of the telescope focus function of the DL were derived using the technique described by Pentikäinen et al. (2020). The profile of the telescope focus function, profiles of raw, noise-filtered, and calibrated DL backscatter for 7 March 2016 are shown in the online supplemental material.

The ceilometer and the DL are calibrated using the technique described in O'Connor et al. (2004). The technique exploits the constant integrated backscatter and hence the lidar ratio measured by the Ceilometer and the Doppler lidar during nonprecipitating stratocumulus cloud conditions with drop diameters between 20 and 50 μm . The extinction and

backscatter efficiencies at 905-nm, 1.5- μm , and 8.6-mm wavelengths were simulated using the “MIEV” scattering code of Wiscombe (1980). The simulated extinction and backscatter efficiencies along with the calculated lidar ratio are shown in the online supplemental material. For these calculations, the refractive index of liquid water was $1.33 + j(5.61 \times 10^{-7})$ at 905 nm, $1.32 + j(1.35 \times 10^{-4})$ at 1.5 μm , and $5.22 + j(2.80)$ at 8.6 mm (Hale and Querry 1973; Kou et al. 1993). The lidar ratio was calculated by assuming a normalized Gamma drop size distribution (DSD). When the modal diameter of the DSD is between 20 and 50 μm the lidar ratio is 18.87 sr at 905-nm wavelength and 21.83 sr at 1.5- μm wavelength (shown in the online supplemental material). There is no significant sensitivity to the shape of the droplet size distribution for these modal diameters (O'Connor et al. 2004; Westbrook et al. 2010b), given the range of DSD shape parameters expected to be encountered in nature (Miles et al. 2000).

At 905-nm wavelength the lidar ratio decreases from its value of ~ 15 sr at 100 μm to a value of ~ 4 sr at 1 mm. At 1.5- μm wavelength, the lidar ratio increases from ~ 25 sr at 100 μm to 500 sr at 1 mm. Hence, larger drizzle drops (diameter $> 100 \mu\text{m}$) decrease the lidar ratio for the ceilometer and increase the lidar ratio for the DL. In addition, the absorption efficiency at the two wavelengths differs significantly with much higher absorption at 1.5 μm as compared with that at 905 nm. The ratio of Mie-to-Rayleigh backscatter cross section at 8.6 mm suggests that Mie scattering effects are important for drizzle drops larger than about 300 μm . The ceilometer lidar ratio and the Ka-band ratio of Mie-to-Rayleigh backscatter cross sections are used to derive drizzle microphysical properties below the cloud base, as shown later.

The vertically integrated lidar backscatter B (sr^{-1}) is related to the lidar ratio S (sr) through the following:

$$B = \frac{1}{2\eta S},$$

where η is the multiple-scattering factor. The lidar ratio and hence the integrated backscatter are constant for nonprecipitating stratocumulus clouds with drop diameters ranging between 20 and 50 μm . Data collected on 7 March 2016 were used to calibrate the lidars (Fig. 1). Nonprecipitating stratocumulus clouds with tops around 1.5 km were observed from 0200 to 0600 UTC, followed by light drizzle that completely evaporated around 500 m. Drizzle below the cloud was captured by both lidars because aerosol returns were much weaker (less than $1 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$) than returns from drizzle ($\sim 3.16 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$). The theoretical multiple scattering factor for liquid clouds at this altitude were calculated using the code described in Hogan (2008), with a value of 0.79 for the ceilometer and 0.95 for the DL. The combination of the variation in the lidar ratio for liquid water clouds and their multiple scattering factor can lead to an uncertainty of about 10% in the calibration value (O'Connor et al. 2004). The lidar ratio reported by the ceilometer was higher than the theoretically determined value during the nonprecipitating conditions and was adjusted to obtain the calibration constant of 1.3109. A similar procedure was followed for the DL that exhibited much lower lidar ratio during nonprecipitating conditions

yielding a calibration constant of 0.4031. Consistent with Mie calculations, the lidar ratio of the ceilometer decreased during precipitation, while that of the DL increased during precipitation.

c. Discerning DL returns from aerosols and hydrometeors

Drizzle is known to be frequently present below marine stratocumulus clouds at the ENA site. Here we propose an objective technique to distinguish between the DL returns corresponding to aerosols and those corresponding to hydrometeors. The ENA site experiences a broad range of aerosol conditions (Zheng et al. 2018), and it also experiences a range of drizzle conditions (Wu et al. 2020). The proposed technique is illustrated using the data collected on 26 November 2016 (Fig. 2). A wide range of drizzle rates and hydrometeor sizes were observed on this day and was also analyzed by Cadetdu et al. (2020). The DL and the KAZR data were gridded to a uniform 5-s temporal and 50-m vertical grid. The DL signal extinguished completely at the cloud base as reported by the ceilometer. Prior to distinguishing the DL returns between aerosols and hydrometeors, the cloud-base height is determined from the DL signal using the technique proposed by Westbrook et al. (2010a). The technique uses the vertical gradient (attenuation) of the DL backscatter (Fig. 2c) and determines the cloud base when this gradient is greater than $1 \times 10^{-7} \text{ m}^{-2} \text{ sr}^{-1}$. On average the cloud-base height determined from DL was 150 m lower than that reported by the ceilometer.

Cumulative distribution functions (CDF) of DL backscatter below the cloud-base height inside and outside the KAZR-reported hydrometeor occurrences were developed (Fig. 2e) using the data for the entire day. The 95th-percentile value of the DL backscatter below the cloud base in hydrometeor-free regions, as reported by the KAZR, was identified. This value was then chosen to develop a mask identifying DL returns corresponding to hydrometeors and those corresponding to aerosols (Fig. 3d). For this case, the threshold value was $1.44 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$, which is very close to the value of $1.5 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$ used by Westbrook et al. (2010a). DL returns lower than the threshold value are classified as aerosol; those higher than the threshold are classified as hydrometeors.

The DL returns near the cloud base within heavy drizzle shafts are affected by hydrometeors, while very few DL returns are affected by hydrometeors within weak drizzle shafts. This approach allows objective identification of DL returns corresponding to hydrometeors and those corresponding to aerosols during varying aerosol conditions. The technique was applied to all of the data collected between 2015 and 2020. Both DL and KAZR data were available for 1828 days (out of 2192). The average backscatter value for separating DL returns from aerosols and hydrometeors was $2.26 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$, with a median of $1.5 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$. The time series and histogram of the daily values of this threshold are shown in the online supplemental material.

d. Dataset

As aerosols have negligible fall velocity, the DL reported mean Doppler velocity during vertically pointing mode corresponding to aerosol returns can be used as a surrogate for the vertical air motion. Hourly profiles of variance and skewness of

FIG. 1. Time-height profile of (a) radar reflectivity, (b) ceilometer backscatter, and (c) DL backscatter. Also shown are time series of raw and calibrated lidar ratio from the (d) ceilometer and (e) DL. The data collected during nonprecipitating and weakly precipitating stratocumulus cloud conditions on 7 Mar 2016 were used in this figure.

vertical air velocity were calculated to characterize turbulence in the MBL. As the DL signal extinguishes completely shortly above the cloud base, these estimates reflect the nature of turbulence in the subcloud layer during cloudy conditions. Because precipitation affects the maximum height at which the

vertical air motion can be observed, the variance and skewness were only calculated at heights for which at least 80% of samples were available during the hour.

Various techniques have been proposed to derive vertical air motion from the data collected by vertically pointing cloud

FIG. 2. (a) Time–height profiles of (a) radar reflectivity and (b) DL backscatter (shading) and ceilometer cloud-base height (black dots). (c) Time–height profiles of (c) gradient of DL backscatter and (d) derived hydrometeor mask, where 0 corresponds to aerosol returns and 1 correspond to hydrometeor returns, (shading) and ceilometer cloud-base height (black dots) and DL cloud-base height (red dots). Note that the ceilometer cloud-base heights are identical in (a)–(d) as are the DL cloud-base heights in (c) and (d). Also shown is (e) the CDF of DL backscatter pertaining to aerosol and hydrometeor returns, the derived threshold for identifying hydrometeor returns (solid vertical black line), and the threshold used by Westbrook et al. (2010a) for identifying hydrometeor returns (dashed vertical black line). Data collected during weakly and strongly precipitating stratocumulus cloud conditions on 26 Nov 2016 were used in this figure.

FIG. 3. (a) Variance of vertical velocity during clear-sky convective conditions binned by the wind direction at the surface. (b) Elevation above mean sea level of Graciosa Island with the direction of the flow corresponding to marine conditions shown in the red sector.

radars as the mean Doppler velocity reported by the KAZR is the sum of vertical air motion and drop fall velocity. The techniques range from restricting the analysis only to non-precipitating cloud conditions (e.g., Ghate et al. 2010), subtracting the average drop sedimentation velocity from the mean Doppler velocity (e.g., Pinsky et al. 2010), and deconvolving the recorded Doppler spectra into cloud and drizzle mode (e.g., Luke and Kollias 2013). Although useful, none of these techniques have been applied to a large amount of data, such as used in this study, to seamlessly derive vertical air motion above the cloud base. In addition, the efficacy in retrieving vertical air motion of even the most sophisticated technique, such as the one proposed by Luke and Kollias (2013), is $\sim 50\%$ near the cloud top and only $\sim 15\%$ near the cloud base, thereby complicating its application to derive moments of vertical air motion in variety of cloud conditions. Hence, the turbulence profiles reported in this study are those derived from the DL, and they correspond to cloud-free regions of the boundary layer. This is further discussed in the last section.

Hourly profiles of KAZR echo fraction were calculated for the entire period using the technique proposed by Clothiaux et al. (2000). As the focus of this work is on the boundary layer, the hourly ceilometer cloud fraction and hourly cloud-base height were calculated only using the cloud-base height values below 3 km within each hour. Similarly, the number of instances of column maximum KAZR reflectivity below 3 km greater than -20 dBZ was used to calculate rain fraction. The data from the surface meteorological station and RWP were also averaged to hourly time scales, thereby generating the

entire dataset on a uniform hourly temporal resolution from 1 January 2015 to 31 December 2020.

3. Annual and diurnal variability

As the focus of this work is on marine boundary layer and the observations are made on an island, we first identify measurements that are unaffected by the island, and then characterize the annual, and diurnal cycle of turbulence in the MBL as well as its controlling factors in the region.

a. Island effect

The ECOR observations are made on the island and hence are affected by the land heating. During clear-sky convective cases over land we expect the variance of vertical velocity to be affected by the land heating. This is used to identify direction of the airflow where the collected observations are unaffected by the island heating. Clear-sky convective hours were identified when the ECOR-SHF was greater than $25 W m^{-2}$, and cloud cover from the ceilometer was lower than 20%. The average ERA5-SHF at the grid point north of the island was $\sim 6.29 W m^{-2}$ during these conditions. The average profiles of variance of vertical velocity during these hours segregated by surface wind direction are shown in Fig. 3. The wind direction is measured clockwise from the north denoting the direction from which the air mass is coming from. Hence values of 0° , 90° , 180° , and 270° denote air coming from the north, east, south, and west, respectively. Figure 3 suggests that the island heating is affecting MBL turbulence during conditions with wind directions between

FIG. 4. Monthly average profile of (a) variance of vertical velocity and (b) skewness of vertical velocity, and (c) monthly average marine fraction. The monthly average values and standard deviations of the cloud-base height (black) and PBL top (red) are shown in (a).

90° and 310°. Although, the island might affect the atmospheric boundary layer through drainage flows and surface roughness, due to their minimal impact on the surface fluxes, these effects are not considered in this analysis. The highest peak on Graciosa is 408 m above mean sea level, and hence it is possible for air masses advected between north of 90° and 310° to be affected by the terrain before reaching the site (Holton 1972). However, apart from the 2-m friction velocity measurements made by the ECOR, the site lacks detailed observations of wind shear at the relevant scales necessary to quantify this effect. Hence for the rest of the analysis, we consider measurements made at wind directions between 90° and 310° to be affected by the island, and between north of 90° and 310° to be unaffected by the island (marine conditions). The measurements unaffected by the island reflect the nature of marine MBL, and hence are referred as “marine” conditions in the rest of the article. This is further discussed in the last section.

b. Annual cycle

On average ~30% of the measurements made at Graciosa (Fig. 4c) are unaffected by the island, with the marine fraction being highest (~38%) during summer and autumn months and lowest (~18%) during the winter months. The strength and location of the high pressure system largely governs the surface winds at the ENA site (Ilotoviz et al. 2021). Hence the data suggest that, for about 30% of the time, the center of the high pressure system is northwest of Graciosa thereby yielding northerly wind conditions at the site, and the system is

northeast of the site yielding southerly winds. The variance of vertical velocity is higher during the winter months than during the summer months (Fig. 4a), while the skewness of vertical velocity below the cloud base is positive all year suggesting that updrafts are stronger than downdrafts (Fig. 4b). The average cloud-base height is ~1 km, and the PBL height is ~1.5 km. In summary, the figure suggests the presence of a shallower boundary layer and weaker turbulence during the summer months than during the winter months.

To gain insights on the causes and effects of changes in turbulence on monthly time scales, we look at the annual cycle of the factors affecting MBL turbulence. In particular the annual cycles of clouds, drizzle, surface fluxes, and winds are analyzed (Fig. 5). The MBL was cloud topped year-round with ceilometer cloud fraction higher than 80% all year. The column maximum KAZR echo fraction suggests cloudiness to be ~20% during the summer months and ~40% during the winter months. Rain was also ubiquitous during the entire year with the rain fraction peaking in the winter months (~50%) and presenting minimum of ~30% during the summer months. The boundary layer winds were also higher during the winter months than during the summer months. The ocean was always warmer than the air denoting transfer of heat and moisture from the ocean surface to the atmosphere. The average ERA5-SHF was $20 W m^{-2}$, and the average ERA5-LHF was $123.50 W m^{-2}$. Both the ERA5-LHF and ERA5-SHF exhibited a distinct annual cycle with maximum in the winter months and minimum in the summer months. Collectively, the figure suggests higher cloudiness, stronger winds, abundant drizzle, and higher surface fluxes

FIG. 5. (a) Monthly averaged profile of KAZR echo fraction, (b) monthly average and standard deviation of ceilometer cloud cover (black) and KAZR rain fraction (red), (c) monthly average profiles of wind speed, and (d) monthly average and standard deviation of sensible heat flux and latent heat flux. All of the values are subsampled for marine conditions only.

during the winter months relative to the summer months. The higher turbulence during winter also explains the lack of thermodynamic decoupling despite deeper boundary layers as reported by Wang et al. (2021).

c. Diurnal cycle

The diurnal cycle of turbulence and its controlling factors are examined next. The local time at Graciosa is 2 h behind UTC. The variance of vertical air motion during marine conditions exhibited a distinct diurnal cycle (Fig. 6) similar to that observed in stratocumulus topped marine boundary layers over the open oceans, with higher values during the nighttime and lower values

in the daytime (e.g., Caldwell and Bretherton 2009). The skewness of vertical velocity was positive during the entire day with higher values below 500 m between 0800 and 1600 local time. The average diurnal cycle during different seasons is similar to that shown in Fig. 6 and is included in the online supplemental material. A version of Fig. 6 with height normalized by the cloud-base height is also included in the online supplemental material.

The diurnal cycle of turbulence-forcing parameters is examined in Fig. 7. On average the surface turbulent fluxes, and winds did not exhibit a diurnal cycle during the marine conditions. However, the KAZR echo fraction exhibited a distinct

FIG. 6. Average diurnal cycle of (a) variance of vertical velocity and (b) skewness of vertical velocity during marine conditions. The local time is 2 h behind UTC.

diurnal cycle with higher cloudiness during the nighttime and lower cloudiness during the daytime, similar to that observed in other stratocumulus regions (Eastman and Warren 2014). The cloud-base height did not exhibit a diurnal cycle, but consistent with the KAZR echo fraction, the ceilometer cloud fraction was higher during the nighttime than during the daytime. The KAZR rain fraction also exhibited a distinct diurnal cycle with maxima in the nighttime and minima in the daytime. Collectively, Figs. 6 and 7 suggest that turbulence, cloudiness, and rain at Graciosa exhibit a distinct diurnal cycle during marine conditions with higher values during nighttime and lower values during the daytime.

4. Air motions below stratocumulus clouds

Now we focus on air motions below stratocumulus clouds observed during the marine conditions with the goal of understanding the covariability of drizzle rates and vertical air motions in the subcloud layer.

a. Case selection and data

A total of 32 cases of marine stratocumulus clouds were identified from the dataset using the following criteria, similar to those used by Jensen et al. (2021): 1) had closed or open cellular stratocumulus clouds below 3 km based on the KAZR, ceilometer and satellite data, 2) were devoid of heavy precipitation originating from mid or high-level clouds, 3) had cloud-top temperatures above 0°C , and 4) had wind direction either less than 90° or greater than 310° . Most of the cases were in summer and autumn, one case was in winter, and five cases were in spring. The list of cases is included in the online supplemental material. Several

techniques have been proposed to retrieve drizzle properties from the data collected by vertically pointing cloud radar (e.g., Kollias et al. 2011; Luke and Kollias 2013), lidars (Westbrook et al. 2010a), and by combining the data from the radar and lidar (O'Connor et al. 2005). Here we retrieved below-cloud drizzle properties by combining data from the KAZR and ceilometer using the technique proposed by O'Connor et al. (2005) at 1-min temporal and 50-m range resolution. Profiles of radiative fluxes and heating rates were also calculated for these cases using the Rapid Radiative Transfer Model (RRTM) with input as described by Ghate and Cadet (2019). The covariability of rain rates, and vertical air motions is studied for similar amount of radiative cooling. For this analysis, the 5-s vertical velocity data were averaged to 1-min temporal resolution. The average wind speed at 500 m was $9.03 \pm 2.67 \text{ m s}^{-1}$; hence the minute-averaged vertical velocity corresponds to eddies with length scales of $\sim 540 \text{ m}$. The analysis is done on minute-averaged data because (i) the drizzle retrievals were made at 1-min temporal resolution and (ii) the analysis reflects on drizzle–dynamics interactions at broad scales, roughly about one-half of the depth of the subcloud layer.

A total of 44 609 one-minute profiles were used in this analysis, out of which 38 554 (86.43%) profiles were cloudy and 17 326 (38.84%) profiles had drizzle below the cloud base. The average cloud-base height during the cases was $1212 \pm 391 \text{ m}$, with cloud-top radiative cooling of $-48 \pm 35 \text{ W m}^{-2}$. The average rain rate during precipitating conditions was $2.29 \pm 6.89 \text{ mm day}^{-1}$, with drizzle modal diameter of $227 \pm 133 \mu\text{m}$. Most of the drizzle evaporated in the subsaturated environment below the cloud base. Hence the rain rate is highest at the cloud base, reducing to zero somewhere within the subcloud

FIG. 7. Average diurnal cycle of (a) KAZR echo fraction and (b) ceilometer cloud fraction (black) and rain fraction (red).

layer. This complicates studying the rain rate-vertical velocity relationship. Hence, we first study their covariability regardless of their location below the cloud base.

The PDF of below-cloud vertical air motion is positively skewed with a zero mean at all values of radiative cooling at the cloud top (Fig. 8). This suggests that updrafts are stronger than downdrafts at a minute temporal resolution. In addition, the width of the distribution increased with increasing cloud-top radiative cooling, which is consistent with having higher variance of vertical velocity (turbulence) associated with stronger forcing. The average and median values of rain rates, as well as their range increased with increasing cloud-top radiative cooling, consistent with an increase in microphysical process rates due to turbulence. A version of Fig. 8b with the outliers is included in the online supplemental material. It shows minute-averaged rain rates in excess of 100 mm day^{-1} during cloud-top cooling stronger than -80 W m^{-2} .

b. Rain rate and air motion covariability

The number of samples containing drizzle below cloud base increased with cloud-top radiative cooling (Fig. 9). Only about 5% of samples below the cloud base with radiative cooling of -40 W m^{-2} contained drizzle, whereas $\sim 45\%$ of samples below the cloud base with radiative cooling of -100 W m^{-2} had drizzle in them. This suggests that even at higher radiative cooling rates most of the air below the cloud base ($\sim 55\%$) is hydrometeor free. The percent of 1-min vertical velocity samples within drizzle areas with different rain rates were segregated by cloud-top radiative flux divergence. The updraft fraction was $\sim 60\%$ in drizzle-free areas below the cloud base independent of

the radiative flux divergence and decreased with increasing rain rates. At low rain rate ($<1 \text{ mm day}^{-1}$), the updraft fraction was $40\%–50\%$. On the other hand, for rain rates greater than 3 mm day^{-1} , when the radiative cooling at the cloud top was -100 W m^{-2} , the updraft fraction was less than 20% . There were very few samples (<200) of stronger rain rates at radiative cooling higher than -60 W m^{-2} . Collectively, the figure shows that weak drizzle ($<1 \text{ mm day}^{-1}$) falls in both updrafts and downdrafts, and the updraft fraction decreases with increasing rain rates.

Box-and-whisker plot of vertical air motion as a function of rain rates for similar amount of radiative cooling at the cloud top are shown in Fig. 10. Relative to the clear-sky values (rain rate = 0 mm day^{-1}), on average the downdrafts increased in strengths with increasing rain rates, however, the updrafts did not exhibit similar behavior. The average vertical air motion was dominated by the downdrafts due to their large numbers, and hence also decreased (more negative) with increasing rain rates. The differences observed at different ranges of radiative cooling were all statistically significant. For weak radiative cooling (between -50 and -30 W m^{-2}) there were very few instances of rain rates stronger than 2 mm day^{-1} and of downdrafts stronger than -1 m s^{-1} . Collectively, these results show that for a similar amount of radiative cooling, the downdrafts strengthened with increasing rain rates, while the updrafts did not exhibit any changes with increasing rain rates.

To further understand the covariability between vertical air motion and rain rate, the correlation coefficient between the two was calculated at the hourly time scales (60 samples). The profiles of these correlation coefficients were then normalized

FIG. 8. (a) Histograms of vertical air motion for different cloud-top radiative cooling rates. (b) Box-and-whisker plot of rain rates for different cloud-top radiative cooling rates.

by the cloud-base height values. The two-dimensional histogram of the correlation coefficient between vertical air motion and rain rate is shown in Fig. 11a. A total of 768 h of data were used in this analysis. During most of the hours the correlation was negative suggesting an increase in the rain rate was accompanied by a decrease in vertical air motion. Most of the samples had a correlation coefficient of -0.25 in the middle of the subcloud layer, suggesting that the effect of rain rate on vertical air motion is higher in that region. A similar analysis done for specific ranges of cloud-top radiative cooling showed similar patterns (see the figure in the online supplemental material). The fact that the peak in the correlation is found in the middle of the subcloud layer suggests that drizzle evaporation is responsible for the decrease in the vertical air motion (increase in downdraft strength). To probe this further, profiles of drizzle-induced evaporative cooling were calculated from the profile of rain rate by taking difference in rain rates at pixels above and below the given height (100 m). Hence similar to rain rates, the clear-air evaporative cooling was 0 W m^{-2} . The mean evaporative cooling of drizzling samples was $-3.32 \pm 7.36 \text{ W m}^{-2}$. Further, the profiles of evaporative cooling were averaged to hourly time scales and normalized with respect to cloud-base height for comparison with the correlation coefficient between rain rates and vertical air motion. The box-and-whisker plot between hourly averaged drizzle-induced evaporative cooling and correlation coefficient between rain rate and vertical air motion (Fig. 11b) show stronger evaporative cooling during instances of lower correlation coefficient. Little to no evaporative cooling was present for hours with the correlation coefficient greater than -0.2 . However, the evaporative cooling was lower than -5 W m^{-2} for

hours with correlation between vertical air motion and rain rate lower than -0.2 . Collectively, the figure suggests an impact of drizzle-induced evaporative cooling on vertical air motion predominantly in the middle of the subcloud layer.

5. Summary, discussion, and conclusions

Marine boundary layer clouds are intimately coupled to the turbulence in the boundary layer that is in turn maintained by surface turbulent fluxes, radiative cooling at the PBL top, wind shear, entrainment, and precipitation. In this work six years of data collected at the ARM ENA site were analyzed to characterize turbulence in the cloud free regions of the MBL, and the air motions below stratocumulus clouds. The backscatter recorded by the Doppler lidar was first calibrated and then used to objectively identify instances of returns corresponding to aerosols or hydrometeors. Due to their negligible fall velocity, the DL returns from aerosols were used to quantify the turbulence properties.

The profiles of variance of vertical velocity segregated by wind direction during clear-sky convective conditions show that the turbulence measurements are impacted by the island heating when the wind direction is between 90° and 310° . Aerosol and cloud measurements made at the site during periods with these wind directions might not correspond to open ocean conditions because terrain-induced turbulence affects the MBL aerosol and cloud fields. Graciosa is the northernmost island in the Azorean chain of islands with an area of 60.25 km^2 , a peak altitude of 408 m, and population of about 4000. However, islands south of Graciosa have peaks with altitude in excess of 2 km and total population of $\sim 242\,000$.

FIG. 9. (a) Drizzle echo fraction below cloud base for different cloud-top radiative cooling rates. (b) Updraft fraction for different values of rain rates for particular cloud-top radiative cooling.

Hence, the aerosol and cloud fields observed during southerly wind conditions are not only affected by the terrain, but also potentially by aerosol emissions from human activities, thereby complicating the interpretation of long-term averages produced when disregarding the flow conditions.

The annual cycle of the variance of vertical velocity during marine conditions reveals higher values during winter than during summer. A similar analysis demonstrated greater cloud cover, surface fluxes and winds during the winter months as compared with the summer months suggesting a causal link between clouds and surface fluxes and the variance of the vertical velocity on monthly time scales. The variance of vertical velocity and cloudiness for the marine subset exhibited a distinct diurnal cycle with higher values in the nighttime hours as compared with the daytime hours. This diurnal cycle was similar to that observed in purely marine stratocumulus cloud conditions over the open oceans with turbulence largely driven by cloud-top radiative cooling. These results suggest that by identifying the marine conditions at the site, it is possible to study the diurnal variability of marine clouds. The thermodynamic advection at the site was not analyzed in this work, however, the diurnal changes in cloudiness and turbulence suggests that the changes in cloudiness are locally driven rather than governed by large-scale processes. The annual and diurnal cycles of radiative fluxes and cooling rates were not analyzed in this work due to the limitation of retrieving vertical profiles of condensate loading during periods that lack measurements of liquid water path from the microwave radiometer. It is plausible that the higher turbulence observed during the winter

months is a reflection of the diurnal cycle with stronger radiative cooling during the winter months than during the summer months.

The covariability between the vertical air motion and rain rate below stratocumulus cloud conditions was studied. For a similar amount of radiative cooling at the cloud top, (i) the updraft fraction decreased with increasing rain rates, and (ii) the strength of the downdrafts increased with increasing rain rates. This suggests that weak drizzle ($<1 \text{ mm day}^{-1}$) can fall under both updrafts and downdrafts, while stronger drizzle primarily falls in the downdrafts. On average the vertical air motion exhibited a weak negative correlation with rain rate with higher instances of negative correlation in the middle of the subcloud layer. In addition, the samples with low (more negative) correlation were associated with higher drizzle-induced evaporative cooling. These results point toward an increase in the downdraft strength below the cloud base due to drizzle evaporation within them. Although downdrafts tend to accompany strong drizzle when it falls below the cloud base, presumably the drizzle formation occurs within the cloud during the updraft phase. As the updraft changes to a downdraft later within the cloud, the evaporating drizzle can drive the downdrafts making them stronger.

The results presented here are also relevant for ESM cloud parameterizations and studies focused on model evaluation using data collected at the ENA site. Ghate and Cadeddu (2019) showed that drizzle and its evaporation affected the MBL turbulence and strength of the eddies (downdrafts). Downdrafts within drizzle were stronger than those within

FIG. 10. Box-and-whisker plot of average downdraft velocity (blue), average vertical velocity (black), and average updraft velocity (red) as a function of rain rates for radiative cooling values (a) between -110 and -90 W m^{-2} , (b) between -90 and -70 W m^{-2} , (c) between -70 and -50 W m^{-2} , and (d) between -50 and -30 W m^{-2} . A rain rate of 0 corresponds to clear air.

drizzle-free regions; hence, inaccurate representation of drizzle can lead to inaccurate representation of the mass transport within the boundary layer. Further, the impact of drizzle on vertical mass transport is absent in ESMs that treat subgrid-scale precipitation as a diagnostic variable. Purely

marine flow unaffected by the island is only observed $\sim 30\%$ of the time at the ARM ENA site, with minimum during the winter months and maximum during the summer months. Graciosa and most of the islands in the Azores are much smaller than the ESM grid resolution, and hence are not

FIG. 11. (a) A 2D histogram of correlation coefficient between rain rate and vertical air motion on hourly time scales as a function of height normalized with respect to cloud-base height. (b) Scatterplot (black dots) and box-and-whisker plot (red) of drizzle-induced evaporative cooling as a function of correlation coefficient between vertical air motion and rain rate.

represented in the ESM. This warrants caution for studies aimed at evaluating ESM output using the data collected at the ENA site, as southerly flow conditions will be affected by the island effect.

Turbulence estimates reported in this work are derived from the data collected by the Doppler lidar and hence correspond to the cloud-free regions of the boundary layer. Similarly, the estimates of drizzle properties were derived from lidar and radar data below the cloud base. High-resolution estimates of vertical air motion in cloudy regions (above the cloud base), derived from the collocated cloud radar would enable seamless estimates of air motion within the entire boundary layer. Such estimates together with retrievals of cloud and drizzle properties will enable studying covariability of drizzle and vertical air motion above the cloud base, and shed insights on key questions like whether drizzle predominantly forms within updrafts or downdrafts, the effects of entrainment on drizzle water content and size etc.

Last, previous studies have shown drizzle evaporation below stratocumulus clouds to cause thermodynamic decoupling and possibly being a harbinger to transition to cumulus cloud regime. Due to lack of reliable meteorological and flux measurements at the ocean surface and high-resolution thermodynamic profiles at the ENA site, this issue was not explored in this study. Measurements made from a buoy located north of the island within the marine flow will allow further exploration of this critical science issue.

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