Stratocumulus-Topped Marine Boundary Layer Processes Revealed by the Absence of Profiler Reflectivity

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ABSTRACT

Stratocumulus (Sc) clouds occur frequently over the cold waters of the southeastern Pacific Ocean. Data collected during two Pan American Climate Study research cruises in the tropical eastern Pacific illuminate many aspects of this Sc-topped marine boundary layer (MBL). Here the focus is on understanding gaps in detectable wind-profiler reflectivities during two boreal autumn cruises. After rigorous quality control that included applying the Riddle threshold of minimum signal-to-noise ratio (SNR) detectability, there are many instances with no measurable atmospheric signals through a depth of up to several hundred meters, often lasting for an hour or more. Rain gauge data from the autumn 2004 cruise are used to calibrate the profiler, which allows SNR to be converted to both equivalent reflectivity and the structure-function parameter of the index of refraction $C_2^n$. Profiles of $C_2^n$ statistics from the two profiler modes (resolutions) highlight the wide range of $C_2^n$ during a 24-h period and bound the atmosphere’s $C_2^n$ when low-mode gaps are not mirrored in the high-mode data. Considering the gaps in terms of $C_2^n$ allows them to be understood as indications of reduced “top down” buoyancy processes and/or reduced turbulent intensity, both of which have been demonstrated by previous researchers to be associated with decoupling within the Sc-topped MBL. A decoupling index calculated from surface and ceilometer data strongly suggests that decoupled conditions were common and that the MBL was coupled when gaps in profiler reflectivity were unlikely. Further study of data from other cruises may lead to a method of using profiler reflectivity as an indicator of decoupled conditions.

1. Introduction

Extensive marine stratocumulus (Sc) clouds over the southeastern Pacific Ocean play a critical role in the dynamics of the ocean–atmosphere system as well as the global atmospheric circulation in the eastern Pacific (Klein and Hartmann 1993). The tops of the Sc are coincident with the top of the marine boundary layer (MBL). Both the height $z_i$ of the inversion atop the MBL and the thickness of the Sc vary in space and time; these variations affect vertical mixing between the ocean and the atmosphere as well as radiative processes within the atmosphere. These attributes are poorly measured by satellites. It is possible that cloud depths could be calculated by using wind-profiler reflectivities to determine $z_i$ and ceilometers to measure cloud-base heights. Initial research along those lines (Piña 2010; Lujan 2011) has incidentally revealed deep, long-lasting voids in profiler observations of the Sc-topped MBL, which we explore here.

2. Basic profiler observations

Our data were collected during east Pacific cruises conducted as part of the Pan American Climate Study (PACS). Wind-profiling radars, ceilometers, and radiosondes were among the instruments on the research vessel (R/V) Ronald H. Brown during several boreal autumn deployments. Data from two cruises are used here: autumn 2000 (17 October–12 November) and autumn 2004 (29 October–27 November). The ship tracks and amount of average daytime Sc coverage are shown in Fig. 1.
Wind profilers are dwelling (not scanning) radars and measure signal-to-noise ratio (SNR), radial velocity, and spectral width. A 915-MHz profiling radar receives returns from refractive index fluctuations with a scale size of 16.5 cm, one-half of the transmitted wavelength (i.e., from Bragg scattering), and from Rayleigh scattering by falling hydrometeors. The index of refraction is a nonlinear function of pressure, temperature, and moisture (Gage and Balsley 1978); fluctuations on this scale are created by turbulence. The 915-MHz profiler deployed during these cruises was a five-beam, electronically stabilized, phased-array system (Law et al. 2002); the same radar was present in both cruises. Data used here come from the vertical beam, which operated in two modes primarily differentiated by their height coverage. The “low mode,” with gates spaced 60 m apart, collected data from relatively low altitudes but with higher resolution. The “high mode” had 105-m gate spacing and so provided less resolution but was pulse coded so as to have increased sensitivity and therefore the ability to collect data higher in the atmosphere. In autumn 2000, the dwell duration was about 30 s and the spacing between vertical dwells in the same mode was about 5.3 min. In autumn 2004, the dwell duration was about 40 s for the low mode and 30 s for the high mode and the spacing between vertical dwells of the same mode was about 6 min.

Figure 2 illustrates the effects of our quality control on one day’s low-mode data from each cruise; SNR was converted to relative reflectivity before plotting. The original dataset contains some SNRs and spectral widths that are unrealistically large, and many observations have very low SNRs and large-magnitude velocities (Figs. 2a,b). Three criteria were used to further refine the data. First, whenever SNRs were above 80 dB all variables were excluded. Second, if spectral width was
Fig. 2. Vertical-beam low-mode 915-MHz profiler data from (left) 31 Oct 2000 and (right) 3 Nov 2004: reflectivity, vertical velocity, and spectral width as functions of time and height, together with ship position. Profiler data are plotted only to 2 km to enhance readability. Shown are (a),(b) original processed data; (c),(d) “partially thresholded” data, confined to instances in which spectral width < 3 m s^{-1} and SNR < 80 dB; and (e),(f) fully “thresholded” data, to which the Riddle threshold was also applied.
larger than $3 \text{ m} \cdot \text{s}^{-1}$ then all variables were excluded.\footnote{These empirical values can be justified on physical grounds. An 80-dB SNR at about 1000 m yields a relative reflectivity of about 85 dBZ, larger than that usually associated with hail, and a $3 \text{ m} \cdot \text{s}^{-1}$ spectral width corresponds to a $2.25 \text{ m}^2 \text{s}^{-2}$ vertical velocity variance.} The effect of this partial thresholding is shown in Figs. 2c and 2d. Also, a minimum threshold of detectability (Riddle et al. 2012) was used to exclude nonatmospheric data with very low SNRs, leaving mostly atmospheric signal.\footnote{We are confident that insects are not an issue in this marine environment.} Although the Riddle threshold is a good way to remove a great many nonatmospheric echoes fairly quickly, researchers doing detailed quality control on a dataset occasionally find that a slightly higher or lower value makes sense for their situation. Examination of scatterplots of SNR versus vertical velocity and before/after plots showed that, for the autumn 2000 data, subtracting 1.5 dB from this profiler’s Riddle threshold, $\text{SNR}_{\text{min}} = -14.55 \text{ dB}$, resulted in a beneficial trade-off of many more “good” data points and a few more “bad” data points (Piña 2010). No adjustment was deemed necessary for the autumn 2004 data (Lujan 2011). Therefore, when SNR was less than $\text{SNR}_{\text{min}}$ (autumn 2004) or $\text{SNR}_{\text{min}} - 1.5 \text{ dB}$ (autumn 2000) all three variables were excluded. The final “thresholded” data are shown in Figs. 2e and 2f.

The data in Fig. 2 are similar to those collected on several days when the ship was not in the ITCZ region. Obvious features are a thin layer of enhanced reflectivity between 1000 and 2000 m, occasional drizzle or rain (e.g., 1100–1200 UTC 3 November 2004), and a layer with small-amplitude vertical motion extending upward from the lowest gates to an undulating height below the enhanced-reflectivity layer. Comparable conditions are observed from 30 October through 3 November 2000 and from 1500 UTC 31 October through 0600 UTC 4 November 2004. (On many other days outside the ITCZ, the noted features were less simple or clear-cut; e.g., multiple layers were present.) For much of our analysis, we concentrate on the two days shown in Fig. 2, but in section 4 we employ data from the longer periods.

Comparison with ceilometer and radiosonde data indicates that the enhanced-reflectivity layer is within the relative humidity gradient atop the boundary layer (Lujan 2011; Hartten et al. 2012). Above that layer, there are few atmospheric returns. Of particular note is the lack of returns within the middle and upper MBL, which also occurs in the high-mode data but to a much lesser extent. \textit{What is the absence of detectable atmospheric signal telling us about this MBL?}

\section{Additional profiler-based measurements}

To explore what the frequent lack of atmospheric returns in the upper portion of these Sc-topped MBLs means, we examine the structure-function parameter of the index of refraction $C_n^2$, which can be obtained from a calibrated profiler. Wind-profiling radars are generally not calibrated to give absolute reflectivity, although their relative reflectivity is very good (Gage et al. 2000). If an independent source of rain data is available, however, then a postdeployment calibration is possible.

\subsection{Calibrating the profiler}

An Optical Scientific, Inc., model ORG-815 optical rain gauge (see online at http://opticalscientific.com/pdf/brochures/ORG/ORG815DS130405.pdf) was deployed on each cruise; data were postprocessed into 10-min rain rates. These can be used to calibrate the profiler, although ship-based measurements of rain are difficult (Yuter and Parker 2001) and there are temporal, spatial, and sampling-volume size issues associated with using ground-based instruments and vertically pointed radars (Tokay et al. 2009). We have adapted the disdrometer-based technique of Gage et al. (2000, 2002) to the ORG-815 data, using total rain accumulation over several hours to average out some of the temporal and spatial variations. We initially chose one day from each cruise with a clear, long-lived rain episode and good profiler data. As our work progressed, we lost confidence in the ORG-815 data from the autumn 2000 cruise and therefore only used autumn 2004 data. Our calibration constants can be used for both cruises, however; profiling radars are such stable devices that the same calibration factor can be used for a long time (Gage et al. 2000, 2002) and the radar operational modes from both cruises are very similar.

On 14 November 2004 (0000–1000 UTC) the gauge accumulated 12.818 mm, mostly from 0500 to 0900 UTC. We classified this rain as stratiform on the basis of the relative reflectivities and velocities (Fig. 3a) measured by the radar during this period\footnote{The high-mode profiler reflectivity shows a bright band at about 4.5 km during the 0500–1000 UTC rain, which rules out warm-rain processes during that time. Reflectivities above the bright band were too weak for convective precipitation; hence, the choice of the stratiform algorithm. The short-lived rain events between 0000 and 0230 UTC were shallow (below 4.4 km) and produced no bright band, and therefore they might have been warm rain. They made only a small contribution to the accumulated rain during the calibrating period, however.} (Williams et al. 1995). We used the Marshall–Palmer relationship $Z = 200R^{1.6} \text{ mm}^6 \text{ m}^{-3}$ (Marshall et al. 1955) to convert reflectivity $Z$ to rain rate $R$. This is the stratiform $Z$–$R$...
relationship from the National Weather Service’s Quantitative Precipitation Estimation system (Zhang et al. 2011); most other stratiform $Z-R$ relationships are a variation of it, and it therefore seems to be a reasonable choice for a basic calibration. We did the calibration separately for each profiler mode, only using range gates with linear power response and uncontaminated by sea clutter. The radar reflectivities contributed to the rain accumulation only when our postprocessing software identified rainfall [cf. the cluster analysis of Williams et al. (2000)], and we assumed that the rain was constant over the time between records. We are not concerned about loss to evaporation because, once fall times were accounted for, there was very good agreement between the profiler and ORG-815 time series.

We determined the radar calibration constants using an iterative process, matching the accumulated rain observed by the radar to the accumulated gauge rainfall (P. E. Johnston et al. 2014, unpublished manuscript). The calibration constant for the low (high) mode is the value for which the average of four (two) consecutive range gates is within 0.001 mm of the gauge accumulation. The accumulation from each gate closely matches the ORG-815 data, both during and after the calibration period (Fig. 3b).

![Fig. 3.](image-url)

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The high mode is more sensitive than the low mode because of its longer pulse length and use of pulse coding. Changing the radar constant by $1.5 \text{ dB}$ would change the accumulation by amounts from 3.11 to 2.49 mm. We believe the calibration to be within 1.5 dB, although this belief relies on our assumption that we have a good precipitation dataset.

After calibration, the profiler SNR can be directly related to equivalent reflectivity $Z_e$ (dBZ) through the generic equation

$$Z_e = \text{SNR} + 20 \log(r) + C,$$  

(1)

where $C$ is a constant and $r$ is the distance to the range gate (m). This equation is similar to Eq. (5.19) in Rinehart (2010) but is written in terms of SNR instead of logarithmic received power $P_r$. The last term $C$ is the calibration factor. For this profiler during these cruises, we find that Eq. (1) can be written as

$$Z_e = \text{SNR} + 20 \log(r) - 63.329$$  

(2)

for the high mode and

$$Z_e = \text{SNR} + 20 \log(r) - 51.324$$  

(3)

for the low mode.

**b. Finding the structure-function parameter of the index of refraction**

After calibration, the profiler reflectivity can also be related to the Bragg scatter from index-of-refraction fluctuations caused by turbulence (Gage et al. 1999). For this profiler’s high mode, this is expressed as

$$\log(C_n^2) = \frac{\text{SNR}}{10} + 2 \log(r) - 19.686.$$  

(4)

For the low mode, $C_n^2$ is calculated using

$$\log(C_n^2) = \frac{\text{SNR}}{10} + 2 \log(r) - 18.486.$$  

(5)

When precipitation is present in a profile, the signal comes from Rayleigh scattering by hydrometeors. A value can be calculated from Eqs. (4) or (5), but it is not the atmosphere’s $C_n^2$. 

By using Eqs. (2)–(5), the SNR thresholds applied in section 2 can be converted to $Z_e$ and $C_n^2$ to get minimum measurable values for these cruises. The resulting profiles (Fig. 4) outline bounds of detectability in terms of traditional atmospheric parameters. All profiles show strong range dependence, losing about 1.5 decades of sensitivity (15 dBZ) with height from 0.5 to 3 km. Figure 4a
reassures us that this profiler is not observing scatter from cloud particles, which generally have \( Z_r < -20 \) dBZ (Gossard and Strauch 1983). Figure 4b includes the maritime lower-tropospheric \( C_n^2 \) profile estimated by Gossard and Yeh (1980) from measurements made by Ames et al. (1955); other estimates of \( C_n^2 \) are found in Doviak and Zrnić (1993, their Fig. 11.7 and associated text). This profiler’s minimum detectable \( C_n^2 \) curves intersect the Gossard and Yeh (1980) mean maritime profile near 1100 m (low mode) and 2400 m (high mode). Comparison with other estimates of \( C_n^2 \) yields similar results. Depending on the variability of \( C_n^2 \), we expect \( C_n^2 \) to become less frequently detectable by this profiler near and above those intersections, at ranges that could be within or only slightly above the MBL.

4. Understanding “missing” profiler reflectivities

Our final thresholding step, application of a minimum SNR threshold, is not universal practice in the profiling community. Therefore we start our last analysis with a less-refined dataset to highlight the impact of the threshold and differences between the profiler modes. We focus on \( C_n^2 \) because it better lends itself to physical interpretations in a nonprecipitating atmosphere than does \( Z_r \).

a. Statistical view of \( C_n^2 \) profiles

Profiles of the minimum, median, and maximum \( C_n^2 \) from the partially thresholded low- and high-mode data on 3 November 2004 are shown in Fig. 5. Individual measurements of \( C_n^2 \) made by the profiler have a large range during this day. The values above \( 10^{-12} \) m\(^{-2/3}\) seen up to \( \sim 1.5 \) km by both modes are consistent with modeled results for the MBL (Burk 1978), although some could also be associated with the brief precipitation just before 1200 UTC (Fig. 2f). Below 1500 m the two median profiles have a similar shape, as do the maximum values. The first available high-mode data are at 727 m, and the last low-mode data whose median is above the Riddle threshold are at 690 m. The high- and low-mode medians at these heights are nearly equal, which is another indication that the calibration is good. The accompanying curves of minimum detectable \( C_n^2 \) reveal how often this profiler cannot truly measure \( C_n^2 \). The median \( C_n^2 \) in the low-mode data is above the minimum detectable value only up to 690 m and was even slightly below it at 1170 m, approximately the height of the thin layer of elevated reflectivity. In the high mode, the median exceeds the minimum detectable value up to 1462 m. Figure 5c shows the amount of data remaining after the Riddle threshold is applied. There is clearly a several-hundred-meters-deep region below the inversion in which the atmospheric \( C_n^2 \) is frequently undetectable by the low mode.

b. Relating \( C_n^2 \) to MBL processes

Parameter \( C_n^2 \) is viewed at least two ways in the literature. Sometimes it is approached as a linear combination of \( C_n^2, C_q^2, \) and \( C_{\theta q}^2 \), the structure-function parameters of potential temperature, moisture, and their covariance. Many researchers have made arguments, on theoretical or practical grounds, for measuring or parameterizing one or more of these subsidiary structure-function parameters, but sometimes \( C_n^2 \) is described as a product of the turbulent intensity and the vertical gradient of the potential refractive index. Both views can point us toward a possible atmospheric meaning to the gaps in profiler reflectivities.

Suppose we view \( C_n^2 \) as

\[
C_n^2 = a^2 C_q^2 + b^2 C_{\theta q}^2 - 2ab C_{q\theta q}^2
\]  

[Doviak and Zrnić 1993, their Eq. (11.143b)]. The scalar coefficients for this formulation are hard to come by (most studies focus on the details of \( C_n^2, C_q^2, \) and \( C_{\theta q}^2 \) and do not try to recombine them), but Gossard (1977) gives \( a^2 = 2.24 \times 10^{-12}, b^2 = 17.8 \times 10^{-12}, \) and \( 2ab = 12.6 \times 10^{-12} \) for mid-boundary layer (BL) tropical maritime air. Peltier and Wyngaard (1995) used a large-eddy simulation (LES) to derive similarity profiles of \( C_n^2, C_q^2, \) and \( C_{\theta q}^2 \) in a convective BL. Their \( C_n^2 \) and \( C_q^2 \) fall off as \( z^{-1/3} \) until just below the start of the interfacial layer, at which point they decrease more rapidly before increasing quickly to their well-documented peak near \( z_i \), the top of the BL. However, \( C_{\theta q}^2 \) departs from this form, being nearly constant from just above the surface layer (\( \sim 0.1z_i \)) to about 0.6\( z_i \) before beginning to increase through the lower interfacial layer toward a peak near \( z_i \). They demonstrated that this different form of \( C_{\theta q}^2 \) is due to “top down” processes, as described by Moeng and Wyngaard (1984) in LES results and by Fairall (1987) using observations. Weakening of these processes would cause \( C_n^2 \) to take a form more like \( C_q^2 \) and \( C_{\theta q}^2 \). The intermittent, deep, hours-long gaps in the reflectivity (and therefore in \( C_n^2 \)) seen in Figs. 2e and 2fc could, therefore, be a sign of a decrease in these top-down moisture processes. This possibility is consistent with the measured variances of equivalent potential temperature and specific total water content in Nicholls (1984), which have a local minimum below the cloud base.

Suppose instead that we view \( C_n^2 \) as

\[
C_n^2 = a^2 e^{-1/3} K_\phi \times 10^{-12} \left( \frac{d\phi}{dz} \right)^2
\]

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Fig. 5. Profiles of minimum, median, and maximum $C_n^2$ (circles and whiskers) on 3 Nov 2004 for the (a) low and (b) high modes of the profiler, together with the appropriate minimum detectable $C_n^2$ from Fig. 4 (dashed line for low mode; solid line for high mode). (c) Profiles of the percent of low-mode (dashed line) and high-mode (solid line) observations on 3 Nov 2004 that were above the minimum detectable level. All values were calculated using the partially thresholded data.
[Doviak and Zrnić 1993, their Eq. (11.143a)]. (Here $a^2$ is a dimensionless constant, $e$ is the eddy rate of turbulent dissipation, $K_b$ is the ensemble-average potential refractive index.) One way to reduce $C_n^2$ would be for the gradient term to become smaller, but it is not clear that such changes could be as deep and long-lived as those seen in Fig. 2. The gaps in reflectivity alternatively could be caused by a reduction in the turbulent intensity in the upper portions of the MBL.

Either view of $C_n^2$ suggests that the reflectivity gaps reveal a widely observed feature of stratocumulus-topped MBLs: the frequent temporary decoupling of the upper, cloud-containing portion from the lower, surface-driven portion. For example, Nicholls (1984) and Nicholls and Leighton (1986) documented very small negative buoyancy fluxes below the cloud layer in decoupled cases, and Tjernström and Rogers (1996) found very small turbulent kinetic energy in what they called the “transition layer.”

c. Other evidence of MBL decoupling

Section 4b was theoretical and somewhat hypothetical. It would be pleasing if we were able to link other observational indications of decoupling with the gaps in the profiler data. There were a limited number of balloon soundings during these two multiday cruise segments, but launch times were determined by clock rather than by conditions, and therefore sampling with respect to profiler gaps is not robust. We therefore consider other BL metrics with higher time resolution.

Decoupling within Sc-topped MBLs is identified in various ways throughout the literature. One commonly noted feature of decoupled MBLs is a great discrepancy between the cloud-base height $z_b$ and the lifting condensation level (LCL) height $z_{LCL}$ computed from lower BL conditions. Jones et al. (2011, hereinafter JBL2011) obtained $z_b$ and $z_{LCL}$ from data collected during 88 subcloud flight legs (~150 m MSL) over the subtropical eastern Pacific during the Variability of American Monsoon Systems (VAMOS) Ocean–Cloud–Atmosphere–Land Study Regional Experiment (VOCALS-REx; October–November 2008). They computed the difference between high-resolution (1 Hz) values and then computed the leg-average$^5$ difference $\Delta z_b$, which they called a subcloud decoupling index. Their $\Delta z_b$ values fell within the range 0–1100 m. After comparing $\Delta z_b$ with adjacent aircraft soundings that they had already identified as well mixed or decoupled, they set $\Delta z_b \leq 150$ m as the criterion for well-mixed Sc-topped MBLs in their study region (70°–85°W, 17°–30°S). According to this criterion, about 45% of the subcloud legs were in well-mixed conditions; for comparison, 28% of the aircraft profiles collected were identified as well mixed. All of the legs with heavy drizzle and 44% of the legs with light drizzle were decoupled.

We have used ship-based observations to compute a similar measure of decoupling. The R/V Ronald H. Brown instrumentation included a ceilometer and near-surface meteorological instruments; from these we have obtained mean cloud-base and LCL heights at 10-min resolution during the multiday periods identified in section 2. Details of the datasets and computations are given in the appendix. We calculated the height difference between contemporaneous cloud bases and LCLs (see Figs. A1a and A1b) and then averaged the differences into hourly values $\Delta z = z_b - z_{LCL}$. This is a longer time average than that used by JBL2011, but it allows for a wide range of conditions to be sampled by a ship moving slower by at least a factor of 10 than the VOCALS-REx aircraft.

The distributions of $\Delta z$ are shown in Fig. 6. The 2004 values are clearly bimodal in distribution; the 2004 values are less clearly so and are perhaps even trimodal. The maximum values are considerably smaller than those of JBL2011. Two of the hourly 2004 values (and 27 of the 418 10-min values) are negative; this fact might be due to our use of surface values instead of 150-m observations, to our location 5°–8° from the equator, or to different synoptic conditions. What constitute the criteria for “well mixed” is open for debate and may even vary from year to year. If we assumed that 28% of the observed MBLs were well mixed, corresponding to the JBL2011 aircraft profiles, the cutoff would be ~300 m in 2000 and ~250 m in 2004; if we assumed a value of 45%, corresponding to the JBL2011 $\Delta z_p$, the cutoff would be ~300 m. Something in the range from 150 (as per JBL2011) to 400 m (the break point in the autumn 2000 distribution) seems a reasonable estimate. In any event, MBLs that are not well mixed were clearly common and fall into a continuum of decoupled states.

A quick visual comparison between time series of $\Delta z$ (not shown) and plots such as Figs. 2e and 2f indicated some correspondence between large $\Delta z$ and deep gaps and between small $\Delta z$ and minimal gaps. The visual results were equivocal enough that we wanted a more objective and methodical comparison between the decoupling index $\Delta z$ and the MBL gaps, but quantifying the gaps proved to be difficult. We decided to use very

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$^4$ Although it can be possible to extract information on turbulent intensity from radar observations of spectral width, this particular profiler was unfortunately configured in a way that renders its data unsuitable for such work.

$^5$ Legs were approximately 12 min (70 km) long.
small $C_n^2$ as a proxy for gaps and to take advantage of the fact that the high profiler mode is more sensitive than the low one, and therefore we returned to $C_n^2$ in the thresholded low- and high-mode data from the multiday cruise segments. We excluded all profiles with precipitation in the lower 2500 m, because we are interested in clear-air estimates of $C_n^2$, and we chose 600 m as the lowest altitude of interest, because below that altitude range-dependent gain in the high mode and sea-clutter contamination make the SNR data unsuitable for reflectivity work. The top of the thin elevated layer of reflectivity was first estimated by software that found the peak reflectivity in each low-mode dwell. This initial determination was hand edited to remove outliers, and the height of the layer top was interpolated across up to five contiguous profiles if there was local enhancement of reflectivity. Following this, the minimum $C_n^2$ occurring between 600 m and the top of the layer height was identified in each high-mode profile; if the minimum occurred multiple times, the highest altitude was chosen. Several of the minima are fairly large (greater than the median $C_n^2$ shown in Fig. 5b), perhaps because of very light precipitation or other contaminants, but we decided to forego further refinement of the data.

Each high-mode profile’s $C_n^2$ minimum was compared with the low-mode $C_n^2$ threshold at that height; from this comparison we derived a 6-min time series indicating whether the minimum high-mode $C_n^2$ was above or below the low-mode threshold at the corresponding height. If the minimum high-mode $C_n^2$ were below the low-mode threshold, we would expect there to be a gap in the low-mode profile near that height because we would have measured with the high mode a very small atmospheric $C_n^2$ that we could not detect with the low mode. This analysis effectively excludes times when drizzle was occurring, either because profiles with precipitation were removed or because contamination by unidentified precipitation yielded $C_n^2$ that was larger than purely atmospheric $C_n^2$, large enough to be observed by both profiler modes.

For each hour, we counted the number of profiles in which the high-mode $C_n^2$ was above and below the low-mode threshold and computed the “percent below” for that hour. The distributions of this metric, shown in Fig. 7, are fairly similar in the two years. At the left end of the distributions are hours in which very small atmospheric $C_n^2$ was rarely observed and so low-mode gaps are unlikely. In 2000, eight hours fell into the 0%–20% bins; in 2004, only two hours did. At the right end of the distributions are hours in which very small $C_n^2$ was frequently observed by the high mode but not by the low, and therefore low-mode gaps are very likely. In 2000, one-half of the hours fell into the 80%–100% bins; in 2004, it was two-thirds of them. The question is whether the MBL is decoupled during these hours when atmospheric $C_n^2$ was so small that the low mode had frequent gaps.

To address this question, we compared these hourly percent-below values with the hourly $\Delta z$ (Fig. 8). The correlation for all paired points in 2000 (Fig. 8a) is 0.44, indicating that hours with a larger percentage of very small atmospheric $C_n^2$ that was undetectable by the low mode (i.e., more expected low-mode gaps in reflectivity) are somewhat associated with a larger decoupling index. The correlation for all points in 2004 (Fig. 8b) is 0.01, indicating no relationship between the two metrics. Despite the lack of any strong linear relationship between

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6 This exclusion was accomplished through a combination of our postprocessing algorithm and hand analysis and removed 16 of 1318 profiles in 2000 and 68 of 871 profiles in 2004.
Δz and persistent small $C_n^2$, other relationships can be seen in the scatterplots.

Hours with no more than 40% of $C_n^2$ minima that are so small as to be undetectable by the profiler’s low mode (i.e., points in the bottom two-fifths of the plots) are associated with $Δz < 350$ m. These $Δz$ values indicate well-mixed or slightly decoupled conditions, and the percent below indicates that long-lived gaps are unlikely. Hours with all profiles having atmospheric $C_n^2$ minima that are undetectable by the profiler’s low mode (i.e., points lying along the top axis) are coincident with $Δz$ values ranging from 146 to 731 m in 2000 (all but three of them above 313 m) and with an almost full range of $Δz$ in 2004. Put another way, during hours for which it was likely that there were low-mode gaps for the full hour, the MBL was decoupled (positive $Δz > 300$ m) two-thirds of the time in 2000 and one-half of the time in 2004. Asterisks to the right of each panel show the

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**Fig. 7.** Distributions of the percentage of high-mode $C_n^2$ minima in an hour falling below the low-mode $C_n^2$ threshold during the (a) autumn 2000 ($N = 120$) and (b) autumn 2004 ($N = 119$) cruises.

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**Fig. 8.** The hourly decoupling index $Δz = z_b - z_{LCL}$ vs the percentage of high-mode $C_n^2$ profile minima that were below the low-mode threshold in that hour during the (a) autumn 2000 ($N = 51$) and (b) autumn 2004 ($N = 84$) cruises. A gray long-dashed line indicates $Δz = 300$ m, and a gray small-dashed line indicates the 150-m criterion used to differentiate well mixed from decoupled by JBL2011. The percentage below during each hour that $Δz$ could not be calculated because no $z_b$ were observed is indicated outside the right axes by asterisks.
percent-below values for the few hours in which $\Delta z$ could not be computed because no cloud bases were observed; in the absence of any MBL cloud bases, we would expect decoupling, and in all of these cases the percentage of minimum $C^2_n$ that was undetectable by the low mode would lead us to expect gaps. The two negative values of $\Delta z$ in 2004 are associated with 90% and 100% of the profiles in those hours having atmospheric $C^2_n$ minima that are undetectable by the low mode, indicating that gaps are expected. We do not understand the physical meaning of $\Delta z < 0$ m and so leave further examination of those situations for future work.

5. Final thoughts

Because wind profilers observe returns from several different types of phenomena [clear air ($C^2_n$), precipitation ($Z_r$), airplanes, birds, insects, sea clutter, etc.], the use of reflectivity can be problematic without intensive quality control of the data. In contrast, the absence of signal is a robust measurement. Although researchers usually want as many numbers as possible from their instruments, the thoughtful removal of recorded values that are theoretically and/or physically unrealistic may also yield useful information. In this case, the absence of valid low-mode profiler reflectivities within the stratocumulus-topped MBL, lasting an hour or more and sometimes extending to hundreds of meters deep, appears to be a marker of some of the processes that cause decoupling: reduction of buoyancy fluxes and/or reduction of turbulent intensity. We make this statement about processes on theoretical grounds, but with some supporting evidence of decoupling from nonprofiler observations. An hourly decoupling index computed from surface and ceilometer data showed that when gaps were unlikely the MBL was coupled and that when gaps were likely for the full hour the MBL was coupled one-third to one-half of the time and decoupled the rest of the time. These relationships are more clear and robust during the autumn 2000 cruise than during the autumn 2004 one; we take this as a reminder that the ocean–atmosphere system has many degrees of freedom and that what is sufficient sampling for the study of one problem may be insufficient for another. It is fortunate that similar data are available from several other cruises in this part of the eastern Pacific, and analysis of those data might clarify the relationship between decoupling in Sc-topped MBLs and reflectivity gaps in profiler observations. Also, a different profiler with greater low-mode sensitivity might be able to observe lower $C^2_n$ and have fewer gaps; in such a case, some $C^2_n$ threshold related to decoupling might be discernable. Regardless of the low-mode sensitivity, because information about very small atmospheric $C^2_n$ can be discerned from quality-controlled reflectivity alone, profiler data may be able to be used to identify the duration and depth of decoupling in Sc-topped MBLs.

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APPENDIX

Cloud-Base Measurements, Surface Meteorological Conditions, and Calculation of Lifting Condensation Level

Our time periods of interest for this portion of the study are from 30 October through 3 November 2000 (5.0 days) and from 1500 UTC 31 October through 0600 UTC 4 November 2004 (3.625 days). During these times, the ships were located outside the ITCZ region and within the approximate bounds 4°–8°S, 92°–110°W, and profiler reflectivities showed the same features seen in Figs. 2e and 2f and discussed in section 2.

a. Cloud-base heights

Cloud-base measurements were made during both cruises with a Vaisala, Inc., CT25K. During the autumn 2000 cruise, there were no ceilometer data available from 1850 UTC 29 October until 1750 UTC 1 November. The original 15-s data were processed into 10-min values of 15th-, 50th-, and 85th-percentile heights. We used the median cloud bases, discarding all values above 2000 m because our interests were confined to levels near or below the thin layer of reflectivity seen between 1000 and 2000 m. The 10-min values are variable (Figs. A1a,b), but exhibit more short-term variability (scatter) in 2004 than in 2000.
b. Surface meteorological conditions

During both cruises, sea surface temperature $T_s$ was measured at a nominal depth of 0.05 m with a thermistor A1 mounted inside a Brookhaven National Laboratory “sea snake” in a fashion similar to that originally described by Fairall et al. (1997). Air temperature $T_a$ and specific humidity $q$ were measured with an aspirated Vaisala HMP-235 mounted 15.5 m above the ship’s local sea level. Humidities were increased by 3% after intercomparison with psychrometer values. The 10-s samples were averaged into 1-min values; these data were postprocessed into 10-min values, which are shown in Figs. A1c and A1d. Temperatures $T_s$ and $T_a$ showed greater long-term variability and $q$ showed a larger range in 2000 than in 2004; $T_s$ ranged from 21.6°C to 24.3°C in 2000 and from 21.9°C to 22.7°C in 2004, and $T_a$ ranged from 20.5°C to 24.9°C in 2000 and from 19.7°C to 23.4°C in 2004. Specific humidity ranged from 15.7 to 18.6 g kg$^{-1}$ in 2000 and from 16.1 to 16.9 g kg$^{-1}$ in 2004.

Surface pressure $P_{sfc}$ and sea level pressure (SLP) were not recorded during the cruises. During the times of interest, the ship was generally between the 5° and 8°S Tropical Atmosphere-Ocean array (TAO) buoys at 95° and 110°W (McPhaden 1995); the only pressure data available are from the 95°W buoys in 2000, however (Cronin et al. 2002). Analysis showed that the 8°S surface pressures were about 0.7 hPa higher than the 5°S values from 29 October to 4 November and the mean daily cycle (from 29 October through 1 November) at 8°S had a range of almost 4 hPa about a mean value of 1015.0 hPa. Composite mean maps of SLP from the Twentieth Century Reanalysis project (Compo et al. 2011), averaged over those same four days, were about 0.5 hPa less than the TAO mean at 95°W, 8°S. A 6-day mean (from 29 October through 3 November 2000) was within 1.0 hPa of the 4-day TAO mean and showed a gradient at 8°S of 0.5 hPa between 95° and 115°W. A 5-day mean from 2004 (from 31 October through 5 November) showed that gradient to be 0.2 hPa across a field of pressures that were approximately 2.0 hPa less than in the 2000 composite.

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Confident that pressure gradients were low across the ship’s relevant cruise tracks, we tested the effect of using $P_{sfc} = 1010 \text{ hPa}$ versus $P_{sfc} = 1015 \text{ hPa}$ (the high pressure dropped the height of the LCL by about $10 \text{ m}$) and the effect of incorporating a 4-hPa daily cycle (this increased the standard deviation of LCL heights from $\sim 0.1 \text{ m}$ to a value of $\sim 2 \text{ m}$). We ultimately decided to use a constant $P_{sfc} = 1010 \text{ hPa}$ in our LCL calculations.

c. LCL computations

Computing LCL from $T_a$, $q$, and $P_{sfc}$ required several steps, which we briefly outline for completeness. Specific humidity was converted to mixing ratio and then to vapor pressure; saturation vapor pressure was computed from $T_a$ using the formula of Hyland and Wexler (1983). Dewpoint temperature was computed using the formula of Bolton (1980), and the temperature of the LCL $T_{LCL}$ was computed from the formula of Barnes (1968). This was converted to the pressure $P_{LCL}$ using Poisson’s equation. Next, we followed the methods of Wilde et al. (1985). Virtual temperature $T_v$ was computed as per Wallace and Hobbs [1977, their Eqs. (2.17) and (2.14)]. By assuming that the mixing ratio was constant between the surface and the LCL, we were able to compute the vapor pressure at the LCL $e_{LCL}$; we used this, together with $P_{LCL}$ and $T_{LCL}$, to compute the virtual temperature at the LCL $T_{v,LCL}$. The height of the LCL $Z_{LCL}$ was computed from the hypsometric equation with $P_{sfc}$ required several steps, which we briefly outline for completeness. Specific humidity was converted to mixing ratio and then to vapor pressure; saturation vapor pressure was computed from $T_a$ using the formula of Hyland and Wexler (1983). Dewpoint temperature was computed using the formula of Bolton (1980), and the temperature of the LCL $T_{LCL}$ was computed from the formula of Barnes (1968). This was converted to the pressure $P_{LCL}$ using Poisson’s equation. Next, we followed the methods of Wilde et al. (1985). Virtual temperature $T_v$ was computed as per Wallace and Hobbs [1977, their Eqs. (2.17) and (2.14)]. By assuming that the mixing ratio was constant between the surface and the LCL, we were able to compute the vapor pressure at the LCL $e_{LCL}$; we used this, together with $P_{LCL}$ and $T_{LCL}$, to compute the virtual temperature at the LCL $T_{v,LCL}$. The height of the LCL $Z_{LCL}$ was computed from the hypsometric equation with $R_p g_0 = 29.3$ [Wallace and Hobbs 1977, their Eq. (2.30)] and $z_0 = 3 \text{ m}$, the height of the pressure measurements on the Autonomous Temperature Line Acquisition System (ATLAS) and TAO moorings. The resulting $Z_{LCL}$ for both cruise segments are shown in Figs. A1a and A1b; they are highly variable on subhourly and subdaily scales, ranging from $\sim 300$ to $\sim 1100 \text{ m}$, although the character of that variability is somewhat different in each cruise.

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