Clouds over the Southern Ocean as Observed from the R/V Investigator during CAPRICORN. Part II: The Properties of Nonprecipitating Stratocumulus

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ABSTRACT

The properties of clouds derived from measurements collected using a suite of remote sensors on board the Australian R/V Investigator during a 5-week voyage into the Southern Ocean during March and April 2016 are examined. Based on the findings presented in a companion paper (Part I), we focus our attention on a subset of marine boundary layer (MBL) clouds that form a substantial portion of the cloud-coverage fraction. We find that the MBL clouds that dominate the coverage fraction tend to occur in decoupled boundary layers near the base of marine inversions. The thermodynamic conditions under which these clouds are found are reminiscent of marine stratocumulus studied extensively in the subtropical eastern ocean basins except that here they are often supercooled with a rare presence of the ice phase, quite tenuous in terms of their physical properties, rarely drizzling, and tend to occur in migratory high pressure systems in cold-air advection. We develop a simple cloud property retrieval algorithm that uses as input the lidar-attenuated backscatter, the W-band radar reflectivity, and the 31-GHz brightness temperature. We find that the stratocumulus clouds examined have water paths in the 15–25 g m\(^{-2}\) range, effective radii near 8 \(\mu\)m, and number concentrations in the \(10^{13}\) range in the Southern Ocean with optical depths in the range of 3–4. We speculate that addressing the high bias in absorbed shortwave radiation in climate models will require understanding the processes that form and maintain these marine stratocumulus clouds in southern mid- and high latitudes.

1. Introduction

Unlike the northern high latitudes where climate warming is unambiguous, the Southern Ocean (SO) poleward of the Antarctic Circumpolar Current (ACC) has enigmatically shown little warming over past decades (Armour et al. 2016). This is in contrast to the thinning and shoaling of ice sheets along the west Antarctic and substantial warming equatorward of the ACC. The reasons, while not entirely clear, seem to be derived from an interesting mix of oceanographic processes driven by external forcing but also governed by slow processes such as intermediate and deep ocean circulations. Warming of the high-latitude Southern Ocean will not occur, it is thought, until the heat stored in the North Atlantic and subducted into the abyss eventually shoals in the Southern Ocean on millennial time scales (Armour et al. 2016; Marshall and Speer 2012). This thinking is consistent with the recent findings of Frey and Kay (2018) who show that the Southern Ocean warming in a fully coupled general circulation model is substantially reduced compared to results derived from models that do not include a fully coupled ocean. However, decadal-scale changes are indeed thought to be occurring primarily in the air–sea fluxes of latent, sensible, and radiant heat. Trends seem to be driven by measurable changes to the large-scale meteorology of the SO (Ceppi and Hartmann 2015) that has driven detectable trends in surface winds (Hande et al. 2012). Korhonen et al. (2010) suggest that the surface wind changes may be modulating cloud properties through sea-spray aerosol.

Set against this complicated natural backdrop, climate models have found it challenging to capture the present-day cloud cover of the Southern Ocean. The models
tend to produce simulations that are strongly high biased in the amount of solar radiation that is absorbed at the ocean surface (Trenberth and Fasullo 2010) because of a low bias in cloud cover. Recently, these biases have been linked to processes in the cold quadrants of midlatitude cyclones (Bodas-Salcedo et al. 2016) and then to the partitioning between ice and liquid water in shallow clouds (Tan et al. 2016; Kay et al. 2016). In particular forcing the models to agree with the phase relationships implied by spaceborne lidar observations (Hu et al. 2010), Tan et al. (2016) illustrate significant feedbacks from large but compensating errors between low-level atmospheric feedbacks. Even at shorter (weather forecasting) time scales, a surface shortwave radiation bias is found in a regional forecast model, and this bias results from large but compensating errors between low-level cloud where the cloud coverage is too low and multilayer situations where the coverage is too high (Protat et al. 2017).

With the exception of global satellite measurements however, vertically resolved observations of cloud properties over the remote Southern Oceans, especially poleward of the ACC, are nearly nonexistent. In this and a companion paper (Mace and Protat 2018, hereafter Part I), we contribute initial studies derived from measurements collected from one of the first such forays into the Southern Ocean and ACC region with a modern suite of ship-based remote sensors. The voyage of the R/V Investigator was conducted in March and April 2016 on the second voyage of the Clouds, Aerosols, Precipitation, Radiation, and Atmospheric Composition over the Southern Ocean (CAPRICORN) program. Protat et al. (2017) described results from the initial short voyage with a reduced set of cloud observations. A third voyage was undertaken in January and February 2018.

In Part I, we examine cloud occurrence and phase partitioning and show that the high cloud fraction of 76% was significantly derived from tenuous non-precipitating clouds. The clouds observed during the 5-week voyage in 2016, while very often composed of supercooled liquid when the thermodynamics allowed, did contain ice-phase precipitation significantly more often than implied by spaceborne lidar, as discussed in Part I. We suggest that interpretation of spaceborne lidar statistics should be tempered by the knowledge that lidar is unable to characterize cloud properties below the first three optical depths in a cloud layer.

In this work, we expand on Part I by examining in more detail the properties of geometrically thin non-precipitating and often supercooled liquid-phase clouds by considering the structure of the marine boundary layer when such clouds were present and then by deriving their microphysical characteristics when a combination of upward-looking characteristics when a combination of upward-looking instruments were able to capture their properties. Our goals in this paper are specifically limited to describing the properties of these clouds and the observed characteristics of the environment in which they occurred. As such, this descriptive study allows us to raise important questions so that future work with additional data will allow us and others to explore in a more comprehensive manner.

In the next section, we describe our methodology for deriving the microphysical properties of these clouds and our approach to analyzing the structure of the marine boundary layer. In section 3, we examine the results, presenting two representative case studies, and then statistics of the supercooled liquid cloud properties. As a point of comparison, we also analyze data collected by the ARM Program on Graciosa Island (GRW), the Azores, in a deployment of remote sensors in 2009 and 2010 (Wood et al. 2015; Rémillard et al. 2012).

2. Data and methods

CAPRICORN took place aboard the Australian R/V Investigator in collaboration with two companion oceanographic missions. The voyage took place from 13 March 2016 through 15 April 2016. Investigator spent roughly two weeks in the vicinity of the Southern Ocean Time Series (SOTS; Schulz et al. 2012) moorings southwest of Tasmania near 45°S and 142°E. Subsequently, approximately two weeks were spent in the region of the Antarctic Circumpolar Current near 50°S and 148°E. The measurement suite during CAPRICORN included an extensive aerosol measurement capability that we will not consider here along with a suite of ship-based meteorological instrumentation and a suite of cloud and precipitation active and passive remote sensors. More details on these other instruments are provided in Part I. The cloud instrumentation included a W-band millimeter radar (Delanoë et al. 2016), a cloud-aerosol Leosphere RMAN-511 mini-Raman lidar operating at 355 nm, and a two-channel (23 and 31 GHz) microwave radiometer. The instrumentation was not mounted on a stabilized platform. However, all instruments pointed relative to the ship’s local vertical so that the same sample volumes were observed by them.

Retrieving the properties of thin liquid clouds with CAPRICORN measurements

Much of the cloud cover during the CAPRICORN voyage was observed to be geometrically and often
optically thin liquid water clouds that, as we show, existed near the top of the marine boundary layer (MBL). Of the 76% total cloud cover during CAPRICORN, approximately 30% of the clouds below 4 km were below the detection threshold of the millimeter radar and observed only by the lidar. Another 30% of the low-level layers had a layer-mean dBZ less than −20 dBZ. It is these nonprecipitating layers observed by the radar with reflectivity less than −20 dBZ that we focus on here. We assume that these layers are composed of a single mode of Rayleigh scattering droplets at the W-band frequency. That these layers are often geometrically and optically thin suggest that they can be approximated assuming layer-mean quantities for radar reflectivity factor Z and lidar backscatter. With this assumption, the lidar backscatter can be treated simply by neglecting multiple scattering and parameterizing the backscatter cross section. Furthermore, these layers cause a measureable but small perturbation to the 31-GHz brightness temperature Tb that tends to be less than a few kelvins, allowing us to treat the microwave radiative transfer approximately in a first guess algorithm. Taken together, the relative simplicity of this subset of clouds allow us to explore their microphysical properties with a simple retrieval algorithm and the measurements available during CAPRICORN and with a similar subset of measurements from the ARM GRW deployment.

The approach we take begins with a semi-analytical approximation that is used as a first guess to an optimal estimation solver. Approximating a single-mode droplet-size distribution using a modified gamma representation, we can write the drop-size distribution (DSD) \( N(D) \), the distribution mass \( (q) \), the effective radius \( (r_e) \), and the total droplet number concentration \( (N_d) \) as follows:

\[
N(D) = N_0 \left( \frac{D}{D_0} \right)^\alpha \exp \left( -\frac{D}{D_0} \right), \quad (1)
\]

\[
q = \frac{\pi}{6} N_0 D_0^3 \Gamma(\alpha + 4), \quad (2)
\]

\[
r_e = \frac{D_0}{2} \Gamma(\alpha + 3), \quad (3)
\]

\[
N_d = N_0 D_0 \Gamma(\alpha + 1), \quad (4)
\]

where \( N_0 \) and \( D_0 \) are the proportionality and scale factors, respectively, and \( \alpha \) is the shape factor. All units unless otherwise stated are cgs. We treat the lidar backscatter near cloud base by assuming a geometric optics scattering approximation. Water droplets with size parameters in excess of 100 are safely in this large particle limit and Mie calculations (Bohren and Huffman 1983) show that the backscatter efficiency \( Q_b \) has a mean of approximately 0.12 with a scaled standard deviation of approximately 100%. Therefore, we express the backscatter cross section \( \beta \) as

\[
\beta = \frac{\pi}{4} Q_b N_0 D_0^3 \Gamma(\alpha + 3). \quad (5)
\]

By substitution of Eqs. (2) and (3) into Eq. (5) and simplification we can write

\[
\beta = \frac{3}{4} \frac{q}{r_e^3} \quad (6)
\]

where \( \rho \) is the density of liquid water. For the radar reflectivity factor \( Z \), we assume Rayleigh scattering at the W-band frequency and integrate Eq. (1) and substitute Eqs. (2) and (3) to derive

\[
Z = C_R q r_e^3 \times F_7, \quad \text{where}
\]

\[
F_7 = \frac{(\alpha + 7)(\alpha + 6)(\alpha + 5)(\alpha + 4)}{(\alpha + 3)^3} \quad (8)
\]

is derived from the recursion properties of the gamma function and \( C_R = 10^8 \times \lambda_R^4 / K^3 \pi^2 \) is used to convert to standard radar units of \( \text{mm}^6 \text{m}^{-3} \).

The thin liquid-phase clouds that we are considering tend to cause a small but measureable increase in \( T_b \) over the cloud-free sky in the 31-GHz channel of up to 5 K and generally less. We approximate the Rayleigh absorption efficiency, \( Q_{abs} = a_b D^3 \) where \( a_b \) can be derived using the refractive index of water and \( b_b \) can be assumed to be 1. From Mie theory we find that \( a_b \approx 1.71 \) and \( b_b \approx 1.076 \), which we assume to be 1. Thus we can express the 31-GHz volume absorption coefficient of the layer,

\[
A_{cld} = \frac{\pi a_b}{4} N_0 D_0^3 \Gamma(\alpha + 3). \quad (9)
\]

Then, the \( T_b \) of the layer can be approximated with

\[
T_b = T_{eff} \left[ 1 - \exp(-\tau_A) \right], \quad (10)
\]

where \( \tau_A = A_{cld} h \) is the absorption optical depth and \( h \) is the depth of the layer. In Eq. (10) we have neglected any downwelling energy from above the layer and any attenuation of the downwelling energy between the cloud and the surface. The quantity \( T_{eff} \) is the layer-mean physical temperature. Substituting Eqs. (9) and (2) into Eq. (10) we can write

\[
\ln \left( 1 - \frac{T_b}{T_{eff}} \right) h = a_b \frac{3}{2 \rho} \frac{q}{(\alpha + 3)} \quad (11)
\]

Equations (6), (7), and (11) then form a set of approximate forward model equations relating the

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measurements of the single-layer thin liquid cloud to the layer microphysical properties.

In practice we iterate over a range of \( a \) values between 0 and 5. For a given \( a \), Eq. (11) is used to solve for \( q \), which is then used in Eq. (6) to solve for \( r_e \). The resulting DSD properties and the assumed \( a \) are used in Eq. (7) to find a value of \( Z \). This \( Z \) is then compared to the observed layer-mean \( Z \) and the \( a \) that produces the best agreement is chosen as the first guess DSD for the optimal estimation (OE) step.

OE is a means of applying Bayes’s theorem using Gaussian statistics and assumptions of linearity—that is, a first-order Taylor expansion of the measurement vector \( \{ y = [\ln(Z), \delta T_b = (T_h - T_{b,clcl})/T_{b,clcl}, \ln(\beta)] \} \) about the unknown microphysical parameter vector \( \{ x = [\ln(q), \ln(r_e), \ln(\alpha)] \} \) is sufficient to describe the problem. The OE allows us to account for the uncertainties in the measurements, the assumptions, and for taking account of prior knowledge (Rodgers 2000). See Mace and Benson (2017) for a more thorough discussion of our approach to OE algorithms such as the one we are considering here. Because we assume a single-mode liquid DSD and approximate the layer as a vertical mean (i.e., 1 layer), the OE problem is more straightforward than outlined in Mace and Benson (2017). Essentially our goal is to minimize the cost function

\[
\Phi = [y - F(x)]^T S^{-1}_y[y - F(x)] + (x - a)^T S_a^{-1}(x - a).
\]

(12)

Using Gauss–Newton iteration, that reduces to iteratively solving

\[
x_{i+1} = x_i + (S_a^{-1} + K_b^T S_y^{-1} K_b)^{-1} K_b^T S_y^{-1} [y - F(x_i)] - S_a^{-1}(x_i - a)
\]

(13)

until convergence is reached after \( i \) steps. In Eqs. (12) and (13), \( F \) represents the forward models that simulate \( y \) in terms of \( x \) and assumptions. We use Eqs. (6) and (7) for \( Z \) and \( \beta \) but we implement the Delta–Edington microwave radiative transfer model (Lebsock et al. 2013) for the simulated 31-GHz \( T_h \). Additionally, we define the measurement error covariance matrix \( (S_y) \), the covariance \( (S_y) \) of the prior state \( (a) \), the sensitivity of \( y \) to \( x \) \( (K_b) \), and of \( y \) to assumptions \( (K_y) \), and the retrieved state covariance \( (S_a) \), respectively, as

\[
S_a = \begin{bmatrix} \sigma_q^2 & \sigma_{qr}^2 & \sigma_{qr}^2 \\ \sigma_{qr}^2 & \sigma_{r_e}^2 & \sigma_{r_e}^2 \\ \sigma_{qr}^2 & \sigma_{r_e}^2 & \sigma_{a}^2 \end{bmatrix},
\]

(15)

\[
a = \begin{bmatrix} \gamma \\ \tau_e \\ \sigma \end{bmatrix},
\]

(16)

\[
K_b = \begin{bmatrix} \frac{\delta \ln(Z)}{\delta \ln(q)} & \frac{\delta \ln(Z)}{\delta \ln(r_e)} & \frac{\delta \ln(Z)}{\delta \ln(\alpha)} \\ \frac{\delta \ln(T_b)}{\delta \ln(q)} & \frac{\delta \ln(T_b)}{\delta \ln(r_e)} & \frac{\delta \ln(T_b)}{\delta \ln(\alpha)} \\ \frac{\delta \ln(\beta)}{\delta \ln(q)} & \frac{\delta \ln(\beta)}{\delta \ln(r_e)} & \frac{\delta \ln(\beta)}{\delta \ln(\alpha)} \end{bmatrix},
\]

(17)

\[
S_y = (S_y^{-1} + K_b^T S_y^{-1} K_b)^{-1}.
\]

(19)

In the above expressions, \( \sigma^2 \) represents the variance or covariance as indicated by the subscripts. Overbars are taken to signify means from prior data derived from airborne measurements as described in Mace and Benson (2017) while the variances and covariances of these statistics populate \( S_y \). In this study, we use measurements derived from recent campaigns such as offshore data collected during Olympex (Houze et al. 2017) in similar maritime synoptic situations. Notably, no prior datasets are used from the Southern Ocean. In \( K_b \) and \( K_y \), the derivatives are computed analytically using Eqs. (6), (7), and (11). Note that \( S_a \) above is given a lowercase symbol because the actual observational error covariance is computed by accounting for the uncertainty due to our assumptions listed in \( K_b \) (Austin and Stephens 2001; Mace et al. 2016):

\[
S_y = S_y + K_b S_y K_b^T,
\]

(20)

where \( S_y \) includes the variances of \( a_b \) and \( Q_b \) noted earlier.

The behavior of the first guess liquid-phase algorithm [Eqs. (1)–(11)] is shown in Fig. 1 using the approximate measurement ranges observed during CAPRICORN. Figures 1a–d show how the radar reflectivity [Eq. (7)] varies as functions of \( a \), \( \delta T_b \), and \( \beta \). As expected, \( Z \) increases with \( a \) and \( \delta T_b \) although the increase in \( Z \) asymptotes for a given \( a \) as \( \delta T_b \) increases. This behavior is most pronounced at smaller values of \( \beta \). An opposite tendency in \( Z \) is found with respect to \( \beta \) in Figs. 1c and 1d.
for a given $a$. This is because when keeping $a$ constant and fixing $\delta T_b$, increasing $\beta$ can only be accomplished by moving the particle-size distribution (PSD) to smaller sizes and higher values of $N_d$; $Z$ then decreases in response. It is also noteworthy that the dynamic range of $\beta$ in this context is rather limited within the reasonable range, shifting as a function of the other parameters. It is also interesting that for the values of the parameters for these clouds (especially $\delta T_b$), the radar reflectivity remains small. The sensitivity of CloudSat, for instance, begins where the colors turns from yellow to red, suggesting that CloudSat would be challenged to observe these clouds under most circumstances. Figure 1 also shows that for reasonable physical ranges of $N_d$ and $r_e$,
the range of $\beta$ for particular $\delta T_b$ values is well constrained. While the valid ranges shift to some extent with $\alpha$, the range remains narrow. For instance, for $\alpha$ of 5, $r_c$ in the 10–20-μm range could only be found for $\beta$ between 0.2 and 0.5 km$^{-1}$ sr$^{-1}$.

The uncertainties in the retrieved quantities, taken as the diagonal elements of $S$, in Eq. (19), depend on the assumed measurement uncertainties, the sensitivities of our measurements to the parameters we are trying to retrieve ($K_a$), and on our model parameters and assumptions ($S_b$ and $K_b$). See Mace et al. (2016) for a discussion of the behavior of Eq. (20) relative to these parameters. Given the challenging marine environment in which the measurements were collected and the fact that no prior measurement statistics for the specific clouds of interest in the region are available, we take a conservative approach to estimating uncertainty. For the radar reflectivity, we assume an uncertainty of 2 dB since there was no capacity for routine calibration during the voyage. The $\delta T_b$ term is relative to clear-sky profiles that tended to occur relatively often in broken clouds and is, therefore, insensitive to the absolute calibration of the radiometer although tipping curve calibrations were conducted when clear skies were present (approximately 5 instances during the CAPRICORN voyage). Therefore, we assume an uncertainty of 20% in $\delta T_b$. Since we track the lidar calibration during the voyage (see Part I) when we had a view of the free troposphere, we are reasonably certain of $\beta$ and therefore assume a conservative uncertainty of 25% since we had to use returns from the free troposphere for calibration (see Part I, appendix B). Taking these uncertainties along with the uncertainties in the model parameters $a_0$ and $Q_b$, from above, the uncertainties in $q$, $r_c$, and $\alpha$ tend to be substantial for individual retrieval instances and typically are 120%, 80%, and 60%, respectively, as we show in more detail in section 3. Also, while $N_{hi}$ is not directly retrieved in the algorithm, we take the retrieved quantities ($q$, $r_c$, and $\alpha$), their uncertainties, and the prior data, and conduct a straightforward optimal estimation inversion for $N_{hi}$ that returns uncertainty for single retrievals in the range of approximately 90%.

Ideally, validation of such algorithms would be provided by measurements collected in situ. However, since no in situ validation was available we conduct radiative closure calculations and compare the broadband infrared fluxes measured at the surface with the fluxes calculated from the retrieved microphysical properties and the MERRA thermodynamic profiles supplemented with measured surface temperature, pressure, and relative humidity measurements. Under many circumstances, boundary layer clouds are optically thick blackbodies and infrared fluxes are primarily a function of cloud-base temperature and not very useful for validating retrievals. While certainly temperature matters here, the clouds we examine are often semitransparent in the infrared and the downwelling infrared fluxes provide an interesting constraint on their radiative properties, which are, in turn, dependent on the microphysics. This will be discussed in more detail in section 3.

In total we have applied the algorithm to 7500 retrieval instances during CAPRICORN each relating to ~55 s of data. That amounts to 3.0 days of retrievals. During CAPRICORN, the cloud-coverage fraction was 0.76 (Part I). In 30 days of data collection that amounts to 22.8 days of clouds; 30% of these had layer-mean dBZ between −20 and the minimum detected by the radar but only 80% of these indicated no ice. That amounts to 5.4 days of retrievable clouds. Thus, we applied our algorithm to just over 56% of the possible cloudy profiles. The remainder were not addressed for various reasons such as a wet radiometer, layers that were too thick suggesting precipitation, failed inversions, etc.

3. Results

Our goal with this analysis is to provide an initial observational characterization of a subset of clouds that composed a significant fraction of the cloud cover observed during the 2016 CAPRICORN voyage. As we report in Part I, the overall cloud cover during the voyage was 76%, dominated by clouds in the marine boundary layer. Furthermore, compared with clouds measured during the cold season at the Azores during the ARM GRW deployment where the cloud cover from low-level clouds was 56%, the boundary layer clouds measured during CAPRICORN by the millimeter radar had significantly lower radar reflectivities and were less likely to be precipitating. At both locations, however, a large fraction of the observed cloud cover (30% during CAPRICORN and 20% during the examined GRW period) was not observed by the radars and only observed by the lidars. The lidar-only clouds combined with the subset of clouds that were observed by the radars but were otherwise nonprecipitating (assumed to have dBZ < −20) together dominate the overall cloud fraction—especially in the CAPRICORN data where they represent 60% of all cloud layers. In this section, we consider the conditions under which these clouds were observed during CAPRICORN and we examine the microphysical and radiative properties of nonprecipitating clouds that were observed by the millimeter radars using the technique described in section 2.

In addition to remote and in situ measurements collected from the ship, imagery from the Himawari-8...
geostationary satellite, and synoptic maps created from the MERRA-2 reanalysis, we also evaluate the thermodynamic structure of the MBL using soundings launched from Investigator. On average, one sounding per day on an irregular schedule was attempted depending on interest in the meteorology and local weather restrictions. An exception was a campaign on 22 March where 6 soundings were launched over a 24-h period in order to capture a frontal passage. On several days when Investigator was in the vicinity of the ACC (31 March through 12 April), access to the weather decks was restricted and not all launch attempts were successful.

It is of interest to know the structure of the MBL and where within the MBL the clouds were occurring. While the structure of the subtropical and low-latitude MBL has been extensively studied in the context of understanding marine stratocumulus in the eastern ocean basins and in the transition regions from shallow to deep convection, the higher-latitude marine boundary layer structure has not been extensively examined. Wood and Bretherton (2004) provide a convenient summary of the components of the MBL and Wood (2012) reviews the properties of stratocumulus. As shown in Wood and Bretherton’s Fig. 1, the MBL in the subtropical eastern ocean basins consists of a surface-based well-mixed layer within which cumulus clouds are typically found rising, perhaps, into a layer that is essentially thermodynamically decoupled from the surface. Near the top of this layer, typically near the base of the marine inversion, a layer of stratocumulus, often but not always drizzling, is found.

Using the techniques outlined in Yin and Albrecht (2000) as implemented by Rémillard et al. (2012), we evaluate the nature of the soundings and the structure of the MBL looking specifically for evidence of decoupling as determined from the parameter $\mu = 0.61\theta/(1 + 0.61w) \times (\partial w/\partial p) - (\partial \theta/\partial p)$ where $w$ is mixing ratio and $\theta$ is potential temperature. As discussed in Yin and Albrecht (2000) this parameter attempts to identify the presence of decoupled boundary layers by accounting for the effects of $\theta$ gradients and changes in $w$ both of which are important features of decoupled layers. In our analysis, we first identify the presence of a marine inversion characterized by an increase in temperature and decrease in water vapor over at least a 250-m layer. Of the 30 soundings collected during the 2016 CAPRICORN voyage, a marine inversion was found to be present in 13 of them. The rest of the soundings were associated with frontal features and rain or cold-air vorticity maxima with well-mixed soundings and convective clouds. Recall that the radiosonde launch schedule was irregular and somewhat biased by interesting weather. Of the 13 soundings with identifiable marine inversions, 11 of them had evidence of decoupling where, following Yin and Albrecht, we identified $\mu > 1.5\overline{\theta}$ where the overbar represents the average from the surface to the inversion base. The layers where $\mu > 1.5\overline{\theta}$ were marked by identifiable breaks in the $\theta$ vertical gradient with a well-mixed surface layer below the decoupled transition layer. Note that the factor 1.5 is modified from that used in Yin and Albrecht who used a factor of 1.3. We found by inspection that this modification captured situations with identifiable decoupled layers more faithfully. It is a conservative choice and does not significantly change our results. All of the decoupled soundings had broken to overcast stratocumulus within the transition layers (infrequently drizzling) and with scattered cumulus below. Importantly, the clouds that characterized the majority of the cloud fraction during these periods and most of the lidar-only clouds during this voyage were composed of stratocumulus within transition layers below the marine inversions. When the marine inversion existed at temperatures below freezing, these cloud layers were supercooled and rarely indicated the presence of the ice phase (see Part 1). Next we examine two case studies of these transition-layer stratocumulus.

a. Case study 1: 26 March 2016

On 26 March, Investigator was in the vicinity of the SOTS buoys near 45°S and 145°E. The regional cloud field imaged by the Himawari-8 geostationary satellite (Fig. 2) shows an open-cellular low-level pattern with the downstream frontal band passing over Tasmania. This frontal system had passed over the ship during the previous day and the region was experiencing cold-air advection in a southwesterly flow (Fig. 3). The synoptic-scale meteorology remained active with upper-level jet maxima oriented on either side of the trough axis at 250 hPa. Cirrus associated with the jet stream is just noticeable in the IR imagery south of the ship location. Two soundings were launched during the local day on 26 March and they were similar in most respects. The sounding at 0624 UTC (1524 local time), shown in Fig. 4, has a marine inversion near 2.4 km (the deepest boundary layer observed during the campaign) and was decoupled from the surface with a transition between the surface mixed layer near 1.2 km, which also happened to be the freezing level. The lifting condensation level (LCL) was near 1 km. We note on the lidar-attenuated backscatter time series that a reasonably sharp transition to lower values occurs at the transition-layer boundary. This suggests that the stratocumulus layer existed in a distinctly different aerosol environment than the cumulus clouds that were emanating from the surface layer.
Two distinct layers of cloud were observed during the 2-h period centered on the sounding. The lower layer near the LCL was associated with widely scattered cumulus convection, some of which was precipitating, that was imaged by the W-band radar at 0630 and 0700 UTC. The convection was reaching the base of the marine inversion but not penetrating through it. The cloud fraction during the 2-h period was approximately 70% primarily because of a layer of stratocumulus that existed near the marine inversion near where the cumulus reached their maximum vertical extent. It is reasonably straightforward to gauge the approximate optical depth of this stratocumulus layer even though the radar only observed it occasionally. Note that the lidar beam tends to penetrate the layer most of the time but with noticeable attenuation. Thus we can infer that the 355-nm optical depth of the layer was on the order of approximately 3.

The temperature of the stratocumulus layer was about $-7^\circ C$ and, for the most part, the lidar depolarization ratios suggest that the layer was liquid although there were occasional departures to higher depolarization ratios suggesting pockets of occasional ice formation in the stratocumulus. We do note, while it is not obvious in the radar images shown here (see Part I for a lengthier depiction of this period), that the precipitation in the cumulus convection was glaciated at heights above the freezing level since a reasonably obvious melting layer along with a noticeable acceleration in the Doppler velocity (not shown) below the melting layer was observed several times during showers. Lidar depolarization also indicates the presence of the ice phase near cloud base.

During this 2-h period there were approximately 10 min when we were able to implement the retrieval algorithm discussed in section 2. Recall that this algorithm is developed specifically for nonprecipitating stratocumulus although the radar and microwave radiometer must detect the clouds. Figure 5 shows the statistics of the microphysics derived during this short period. The optical depth calculated from the retrievals agrees with our heuristic evaluation of the lidar with a mean value near 2.7. Notable also is the very low number concentrations inferred from the data (16 cm$^{-3}$) with effective radii in the 8-$\mu$m range.

![Image](https://example.com/image.png)
We show this example because it is rather typical of the cold-air advection cases observed in this region during the CAPRICORN voyage where the cloud-coverage fraction is derived primarily from stratocumulus in thermodynamically decoupled layers. While the satellite imagery suggests an open-cellular cloud pattern, the majority of the cloud cover is due to relatively thin nonprecipitating supercooled liquid stratocumulus clouds near the base of the marine inversion. The cumulus penetrating to near the top of the marine inversion occupies a comparatively much smaller portion of the fractional cloud cover.

**b. Case study 2: 6 April 2016**

From March 29 through 12 April 2016 the R/V Investigator conducted operations in the vicinity of the ACC where cyclonic and anticyclonic mesoscale ocean eddies were studied (Part I). The shift southward by approximately 6° of latitude marked a considerable change in the meteorology. Referencing Fig. 2 of Part I, the SST and mean air temperatures dropped approximately 5°, and surface pressure and wind, respectively, decreased and increased noticeably. While cloud cover remained relatively constant, the frequency of supercooled liquid clouds increased.

We focus in this case study on a 2-h period centered on a sounding launched at 0400 UTC (1300 local time) on 6 April 2016 depicted in Figs. 6–9. A significant change in the synoptic weather pattern had taken place since the 26 March case study. During the first two weeks of the voyage the pattern was rather high amplitude and slow moving while by the second two weeks of the voyage that encompasses the 6 April case, the pattern deamplified and was more progressive. This is evident comparing, for instance, the 250-hPa plots in Figs. 3 and 7. The region experienced a rapid sequence of

![Fig. 3. Meteorological fields as indicated at the top of each panel derived from the MERRA-2 reanalysis at 0600 UTC 26 Mar 2016.]
Fig. 4. Analysis of soundings launched at 0624 UTC 26 Mar 2016. (a) The frequency distribution of lidar-derived cloud base during the 2-h period centered on the sounding launch time; (b) the profile of mixing ratio; (c) the vertical profile of potential temperature and equivalent potential temperature; (d) the temperature and dewpoint. Shown on these figures are the LCL, the base of the transition layer where the lower mixed layer is decoupled from the surface, and the base of the marine inversion. The vertical dashed line in (d) marks the 0°C isotherm. (e) Cross section of vertically pointing W-band radar reflectivity (white circles indicate lidar base); (f) copolar lidar-attenuated backscatter (a white line shows the base of the transition layer); (g) the lidar depolarization ratio at cloud base where a value > 0.03 denotes the likely presence of the ice phase at cloud base, whereas below a value of 0.02, we diagnose the liquid phase (see Part I); (h) the 31-GHz brightness temperature measured by the upward-looking microwave radiometer.
storms during the first two weeks of April with sharply defined frontal systems followed by transient high pressure ridges. On 6 April underneath a progressive ridge, an extensive layer of supercooled, primarily liquid phase, stratocumulus clouds was sampled. Imagery shows that the ship was on the northern boundary of a region of uniform closed-cellular clouds between the frontal systems with an open-cellular structure north of 50°S. While the IR image suggests that the layer was uniform, the visible image shows a pillowy spatial structure with pockets of higher reflectance.

The supercooled stratocumulus layer was observed at the ship (which was more or less stationary because of weather during this period) from roughly 2000 UTC 5 April when the downstream front passed until 1200 UTC 6 April when the upstream system began to influence the ship. From 2000 UTC 5 April until approximately 0200 UTC 6 April, the layer was occasionally drizzling with radar reflectivity in excess of −15 dBZ that rarely reached the surface. In these heavier precipitation features, lidar depolarization occasionally suggested the presence of the ice phase although the precipitation was liquid phase during most of this period even though the cloud bases were at −10°C.

By the time of the sounding launch at 0400 UTC 6 April, the layer had thinned considerably. The sounding shows that the stratocumulus clouds existed just below a strong marine inversion that was near 1800 m within a layer that was decoupled from the surface. The freezing level was at 400 m and the cloud layer persisted near −10°C. Satellite imagery shows a north–south-oriented break in the layer that seemed to mark a boundary in its texture at least in the visible imagery. As this boundary passed over the ship at around 0400 UTC (at about the time of the balloon launch), the layer was mostly lost to the cloud radar although the lidar continued to find a mostly opaque layer. Following passage of this break, the layer thickened, became once again visible to the radar, and showed some very light drizzle based on the structure of the radar reflectivity.

While it is less well defined than on 26 March, the lidar-attenuated backscatter (Fig. 8f) shows a shift to lower values above the transition layer indicating that the environment within which the clouds existed was
decoupled from the surface layer. For the most part, the layer was liquid according to the lidar with the exception of isolated pockets of the ice phase indicated by excursions of depolarization ratios above 0.03.

The microphysical properties of the layer observed during this 2-h period are shown in Fig. 10. Compared to the clouds observed on 26 March, these clouds are similar in terms of particle size (~8.5 um) although the water path is higher driven likely by a higher \( N_d \) near 25 cm\(^{-3} \). The higher water paths also result in larger optical depths (~4) that seems to agree more or less with the lack of lidar penetration evident in Fig. 8f.

c. Overall properties of transition-layer marine stratocumulus

The extended layers of supercooled mostly liquid-phase stratocumulus near the tops of decoupled MBLs were common during the period that Investigator spent south of 50°S. Similar conditions were observed from 1700 UTC 2 April until 1400 UTC 3 April (21 h), the period considered above from 2000 UTC 5 April until 1900 UTC 6 April (23 h), 0600 UTC 7 April until 1900 UTC 7 April (37 h), and 2100 UTC 10 April until 0000 UTC 13 April (52 h). The intervening times were when active frontal systems were passing over the ship or the atmospheric profiles were well mixed and convectively unstable, such as 1100 UTC 9 April until 1700 UTC 10 April. Thus, it seems reasonable to advance the hypothesis that the cloud coverage of this region, when not governed by frontal systems and deeper convection is dominated by stratocumulus clouds that form at the tops of decoupled MBLs. When the thermodynamics allow, these clouds tend to be composed of supercooled liquid water with occasional ice precipitation (about 20% of the time as discussed in Part I). In the region of the SOTS operations near 45°S, the stratocumulus layers were accompanied by widely scattered cumulus convection emanating from the unstable mixed layer, while in the region of the ACC south of 50°S, cumulus convection below the stratocumulus was not as frequently observed.

We complete this study by examining the overall statistics of the transition-layer stratocumulus clouds observed during the voyage. Recall that we are
examining a subset of the clouds observed during CAPRICORN because we require that the radar, microwave radiometer, and the lidar are able to sense them and that the layer appears as liquid phase based on lidar depolarization ratio. This subset includes approximately 6800 individual retrievals [4100 (2700) layers warmer (colder) than freezing] that each represent 35-s intervals. As we demonstrate in the case studies, a substantial fraction of these stratocumulus clouds are below the sensitivity threshold of the radar and have water paths that are quite small and not well sensed by the radiometer even though the lidar suggests that the layers are often optically opaque. We find that even though the modeled absorbed shortwave biases are concentrated in regions where supercooled liquid clouds are prevalent, stratocumulus in decoupled boundary layers are prevalent throughout the region. If the absorbed shortwave bias is largely a cloud-cover bias and these stratocumulus layers dominate the cloud cover, it seems reasonable to hypothesize that understanding processes associated with the supercooled version of these stratocumulus layers may be an important part of the solution.

We first address the radiative closure that was discussed in section 2 (Fig. 10). Because these clouds tend to be semitransparent in the thermal IR most of the time (see optical depth on Fig. 10l), comparing the calculated downwelling longwave IR flux with measurements is a useful metric for evaluation of the retrieved properties. Figure 10g shows that these stratocumulus clouds increase the downwelling infrared flux on average by 50 to 80 W m\(^{-2}\) with an average near 70 W m\(^{-2}\). Comparing the calculated flux with the broadband IR flux observed by the pyrgeometer on the ship shows that our calculations, while varying rather widely as would be expected from 1) uncertainties in individual retrieval instances as shown in Figs. 10d,e,f and 2) because we are comparing a hemispheric observation with a vertically pointing measurement, tend to have a mean value near zero, suggesting that our characterization overall is reasonable and essentially unbiased. We conducted a similar comparison with data collected at Graciosa with the ARM data and we find a similarly unbiased comparison of IR flux. While we look forward to more quantitative evaluation of the retrieval algorithm that can be supplied by aircraft in situ measurements, we take these
results as a reasonable, if not preliminary, validation of the technique described in section 2.

Overall the subset of stratocumulus clouds that we examine here tend be tenuous with mean water paths in the 16 g m$^{-2}$ range, droplet effective radii in the 8-μm range with a narrow, approximately normal, distribution. The characteristics of the water path distribution arise from the lognormal distribution of $N_d$ that, on
average, tends to be in the $20 \text{ cm}^{-3}$ range. These $N_d$ values are remarkably small. In only a few cases do we find results of $N_d$ in excess of $60 \text{ cm}^{-3}$.

Comparing the warm and cold versions of these clouds raises interesting questions. We find that the water paths of the cold clouds tend to be larger. With very similar $r_e$, the $N_d$ of the cold clouds tend to be slightly larger with an optical depth distribution and downwelling IR forcing that is noticeably shifted to larger values. This result is derived from reflectivity measurements that are slightly smaller on average but brightness temperature deviations that are slightly larger. As we note from the water path histograms in Fig. 10a, the difference in water path from cold to warm clouds is systematic with the frequency distribution clearly shifted to higher values for the subfreezing clouds.

We next compare the mean results derived from data collected during the 2016 CAPRICORN voyage with 3 months of cold-season data from Graciosa Island, the Azores, during the ARM deployment there in 2009 and 2010 (Wood et al. 2015). We applied exactly the same criteria and algorithm to the GRW data as applied to CAPRICORN. Recall from Part I that the low-level cloud occurrence fraction observed by the ARM remote sensors was substantially smaller (56%) although, when present, the liquid thermodynamic phase statistics were similar with some notable exceptions. One of these exceptions can be seen in Table 1. At GRW, we find slightly larger effective particle sizes and marginally optically thicker clouds on average. The main difference in the two datasets is found in the $N_d$ that is approximately a factor of 2 greater at GRW. This, combined with the slightly larger effective sizes, results in marginally higher water paths at GRW and clouds that are optically thicker on average.

The interesting differences in the cold and warm clouds of the two regions causes us to consider the climate relevant feedback mechanisms associated with mid- and high-latitude maritime low-level clouds. From theoretical studies (i.e., Rieck et al. 2012) and evaluation of general circulation models and satellite imagery (i.e., Gordon and Klein 2014) we know that differences in cloud temperature in liquid boundary layer clouds generate competing climate feedbacks. Because warmer
FIG. 10. Observed and derived properties of nonprecipitating stratocumulus during CAPRICORN. Blue histograms denote supercooled layers colder than freezing, and red histograms denote layers warmer than freezing.
temperatures result in steeper mixing ratio lapse rates, liquid clouds that form in warmer environments tend to have higher water contents and, therefore, have higher albedos than liquid clouds forming in colder environments all else being equal. This brightening with warming results in a negative feedback on a warming climate. A competing positive feedback, however, results when warmer environments can more efficiently mix dry free-tropospheric air into the MBL causing cloud coverage to decrease. Rieck et al. (2012) and Gordon and Klein (2014) agree that the mixing feedback is predominant in subtropical clouds although Gordon and Klein suggest that the mixing ratio lapse-rate feedback may be dominant at higher latitudes.

It seems reasonable to hypothesize that increased dry-air mixing at warmer temperatures would also cause cloud layers to be geometrically thinner, and the lapse-rate feedback would tend to shift the liquid water content statistics to higher values in the warmer clouds. Summarized in Table 1, we find that the LWP increase in the cold clouds in the CAPRICORN data is due entirely to the layers being geometrically thicker in the cold environments with the liquid water content statistics nearly identical between the cold and warm clouds. At GRW we find that the warm and cold stratocumulus layers are overall geometrically thinner. However, the colder layers at GRW tend to be relatively geometrically thicker compared to the warmer layers at GRW as for CAPRICORN, but the GRW liquid water contents tend to be slightly smaller resulting in offsetting effects in the GRW LWP statistics. While clearly, this evaluation needs to be supplemented with additional data, it is an intriguing piece of unique observational evidence and suggests a line of inquiry for examining questions regarding cloud feedbacks by mid- and high-latitude boundary layer clouds.

4. Summary and conclusions

In this second of a two-part set of studies, we examine the properties of clouds observed by the R/V Investigator during a 5-week voyage into the Southern Ocean south of Tasmania, Australia. For comparison, we also examine clouds observed during winter in the northeast Atlantic on Graciosa Island, the Azores, during an ARM campaign there in 2009 and 2010 (Wood et al. 2015). The R/V Investigator focused on the region near 45°S and 142°E during the latter two weeks of March 2016, and in the region of 51°S and 147°E during the first 12 days of April. We refer to these as the SOTS and ACC portions of the voyage. Even though the meteorology evolved as would be expected with the advancing season and due to latitude changes, the cloud cover in both locations was dominated by clouds in the marine boundary layer (MBL). Using a limited set of soundings collected from the ship, we place the predominant cloud cover into the thermodynamic context of the MBL and find that the majority of the cloud cover existed near the base of marine inversions in layers that were decoupled from the surface. These extensive stratocumulus clouds were often found with lower-level cumulus rising into the transition layers. The transition-layer stratocumulus clouds rarely showed signs of ice-phase precipitation down to temperatures as cold as −10°C although the ice phase was present more often than would be suggested by the CALIPSO lidar (Part I).

Overall, the transition-layer stratocumulus tended to be tenuous when compared to their subtropical counterparts (Wood 2012). A large fraction of the layers were below the detection threshold of the cloud radars (Part I and the case studies above) although the lidar observations suggest that these layers were optically thick enough to significantly attenuate the lidar beam much of the time implying optical depths near 3. Examining the microphysical properties of a subset (~30% of cloud layers observed during CAPRICORN) of these clouds that were observed by the cloud radar and microwave radiometer with a technique described in section 2, we find water paths in the 15–20 g m⁻² range—slightly higher at GRW. Effective radii were in the 8-μm range. Droplet concentrations were unambiguously larger at
GRW near 50 cm$^{-3}$ while in the 20 cm$^{-3}$ range in the 2016 CAPRICORN data. As an initial gauge on their radiative significance, we found that these clouds tended to enhance the downwelling IR flux at the surface by approximately 70 W m$^{-2}$.

Especially in the ACC region of the CAPRICORN voyage, we find that supercooled stratocumulus clouds associated with decoupled boundary layers seem to dominate the cloud-coverage fraction between periods of deeper convective cloud systems and fronts (i.e., Fig. 6). From the CAPRICORN data we ascertain that ~60% of the cloud cover observed during the 5-week voyage was derived from these layers. While certainly a 5-week surface dataset should not be considered a statistically significant sample of the Southern Ocean, the data as summarized in Part I and here suggest the following hypothesis. It has been established that the shortwave radiation bias in climate models is associated with a dearth of supercooled clouds in the cold quadrants of the migratory low pressure systems (Bodas-Salcedo et al. 2016). Analyzing ISCCP data, Bodas-Salcedo et al. (2012) found that cloud systems with top pressures in the midlevels (less than 680 hPa) are implicated in this bias. We suggest that the largely nonprecipitating and largely supercooled stratocumulus layers found near the bases of the marine inversion in decoupled boundary layers may form a substantial portion of the clouds that drive the shortwave radiation bias in climate models. In datasets like ISCCP, clouds existing near the marine inversion tend to have their cloud-top pressures biased to lower pressures and given the deep boundary layers found in the soundings during CAPRICORN, the findings of Bodas-Salcedo et al. (2012) are consistent with an ISCCP analysis finding an occurrence bias in midlevel clouds.

The mechanism for maintaining the stratocumulus layers in the decoupled MBL is a topic we leave largely unaddressed in this study. While it appears that shallow convection from the surface layer seems to be present (see the 26 March 2016 case), it is difficult with vertically pointing radar observations to characterize the degree to which this convection resupplies water vapor to the decoupled layer. Future work should focus on the extent to which these tenuous stratocumulus layers contribute to the overall coverage fraction of these regions and investigate the processes that act to form and maintain them as well as the climate feedbacks associated with them.

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REFERENCES


Hu, Y., S. Rodier, K.-M. Xu, W. Sun, J. Huang, B. Lin, P. Zhai, and D. Josset, 2010: Occurrence, liquid water content, and fraction


