Mean Structure and Diurnal Cycle of Southeast Atlantic Boundary Layer Clouds: Insights from Satellite Observations and Multiscale Modeling Framework Simulations

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ABSTRACT

The mean structure and diurnal cycle of southeast (SE) Atlantic boundary layer clouds are described with satellite observations and multiscale modeling framework (MMF) simulations during austral spring (September–November). Hourly resolution cloud fraction (CF) and cloud-top height \( H_T \) are retrieved from Meteosat-9 radiances using modified Clouds and the Earth’s Radiant Energy System (CERES) Moderate Resolution Imaging Spectroradiometer (MODIS) algorithms, whereas liquid water path (LWP) is from the University of Wisconsin microwave satellite climatology. The MMF simulations use a 2D cloud-resolving model (CRM) that contains an advanced third-order turbulence closure to explicitly simulate cloud physical processes in every grid column of a general circulation model. The model accurately reproduces the marine stratocumulus spatial extent and cloud cover. The mean cloud cover spatial variability in the model is primarily explained by the boundary layer decoupling strength, whereas a boundary layer shoaling accounts for a coastal decrease in CF. Moreover, the core of the stratocumulus cloud deck is concomitant with the location of the strongest temperature inversion. Although the model reproduces the observed westward boundary layer deepening and the spatial variability of LWP, it overestimates LWP by 50%. Diurnal cycles of \( H_T \), CF, and LWP from satellites and the model have the same phase, with maxima during the early morning and minima near 1500 local solar time, which suggests that the diurnal cycle is driven primarily by solar heating. Comparisons with the SE Pacific cloud deck indicate that the observed amplitude of the diurnal cycle is modest over the SE Atlantic, with a shallower boundary layer as well. The model qualitatively reproduces these interregime differences.

1. Introduction

The southeast (SE) Atlantic and Pacific low clouds have the largest fractional coverage and are the most spatially extensive low-cloud regimes in the subtropics (e.g., Klein and Hartmann 1993; Norris 1998). Their occurrence is concomitant with cool sea surface temperatures because of coastal upwelling, surface cold advection (Zhang et al. 2010), a shallow marine boundary layer, and strong temperature inversions enhanced by the climatological subsidence. These Southern Hemisphere clouds
reach their maximum development in austral spring (September–November), when the lower troposphere becomes more thermally stable. Unlike other marine boundary layer clouds, the meridionally oriented coastal topography, with elevations up to 5 and 2 km for the SE Pacific and SE Atlantic, respectively, play a key role in intensifying the lower tropospheric stability, which in turn yields larger cloud fractions than their Northern Hemisphere counterparts (e.g., Richter and Mechoso 2004, 2006; Xu et al. 2004).

In addition to their extensive spatial coverage, the SE Atlantic and Pacific boundary layer clouds have, on average, larger liquid water paths and optical thicknesses than other subtropical marine boundary layers (e.g., Jensen et al. 2008). Despite their similarities, the SE Atlantic cloud regime possesses unique features that differentiate it from its SE Pacific counterpart. For instance, the SE Atlantic coastal region is influenced by stronger near-surface coastal winds and stronger surface cold advection than observed in the SE Pacific (Zhang et al. 2010), accompanied by a shallower boundary layer, and less prominent precipitation occurrence (e.g., Leon et al. 2008).

Modeling and observational studies that analyze the SE Atlantic boundary layer variability are overwhelmingly scarce. In contrast, earlier ship cruises (e.g., Garreaud et al. 2001; de Szoeke et al. 2012) and an unprecedented recent field campaign in the SE Pacific, the Variability of American Monsoon Systems (VAMOS) Ocean–Cloud–Atmosphere–Land Study Regional Experiment (VOCALS-REx; Mechoso et al. 2014), revealed rich details of the marine boundary layer variability, cloud vertical structure and microphysics, and aerosol composition. Moreover, VOCALS-REx motivated a significant number of modeling studies and intercomparisons, which provided new insight into the physical processes that drive the cloud deck variability (e.g., Wyant et al. 2010).

While studies over the SE Pacific continue strengthening our understanding of subtropical clouds, there is an increasing recognition of the need to focus on the study of the Namibia–Angola stratocumulus clouds, the most poorly characterized marine boundary layer cloud system in the tropics.

A limited number of specific modeling studies over the SE Atlantic (e.g., Allan et al. 2007; Hu et al. 2008) show an overall underestimate of cloud cover amount and a misplacement of the stratocumulus cloud deck, further impacting the simulated feedbacks between the ocean and atmosphere. In fact, it is believed that the inaccurate prediction of boundary layer clouds contributes to the severe positive sea surface temperature bias in coupled models representing the eastern Atlantic (e.g., Gent et al. 2010). In contrast to a traditional climate model, an emerging modeling paradigm, the so-called multiscale modeling framework (MMF) or superparameterization (Khairoutdinov and Randall 2001; Randall et al. 2003), offers new opportunities to tackle the cloud representation problem in climate models. An MMF uses a two-dimensional cloud-resolving model (CRM) to replace the subgrid parameterization in each grid of a conventional global climate model. While MMF produces evident improvements, recent studies demonstrate that the simulation of low clouds can be further refined when upgrading the low-order turbulent closure of the CRM component of MMF by an advanced third-order turbulence closure (Cheng and Xu 2011). This novel MMF configuration is particularly successful in predicting the spatial and seasonal changes in low clouds over the sub-tropics (e.g., Cheng and Xu 2011; Xu and Cheng 2013b).

Given the satisfactory MMF performance, the model can be further utilized to understand the physical mechanisms that modulate cloud variability.

The diurnal cycle of the marine boundary layer properties is of particular interest as the physical processes that drive the diurnal variability, solar radiative heating, entrainment, and subsidence, among others, play key roles in the clouds’ synoptic and seasonal variability. Moreover, accurate descriptions of the cloud diurnal cycle are essential for estimating the radiative forcing of marine stratiform cloud regimes (Bergman and Salby 1997). Diurnal cycle processes appear to be particularly relevant over the SE Atlantic, where satellite observations produce broad diurnal amplitudes in cloud fraction and liquid water path (Rozendaal et al. 1995; O’Dell et al. 2008). This study extends the work of Cheng and Xu (2013, hereafter CHX13) concerning diurnal cycle modeling in the context of MMF by focusing on the SE Atlantic low clouds. Our two main goals here are 1) to document the mean structure and diurnal variability of low-cloud properties over the southeast Atlantic using several satellite climatologies, which are also used to evaluate the performance of MMF simulations, and 2) to further utilize the model results to understand the factors that drive the diurnal and spatial variability of the SE Atlantic boundary layer clouds. Finally, to gain a better perspective of the SE Atlantic cloud diurnal cycle magnitude in a global context, we carry out a preliminary comparison of the boundary layer properties between the SE Pacific and Atlantic cloud regimes.

2. Model and satellite dataset description

a. Model description

The model configuration has been extensively described in Xu and Cheng (2013a,b), whereas the experimental
setting is the same as that in CHX13. Here we briefly describe the most relevant model features. The host climate model corresponds to the Community Atmospheric Model (CAM), version 3.5 with a finite-volume dynamical core and a horizontal grid size of 1.9° × 2.5°. Of the 32 vertical levels, 12 are below 700 hPa in order to better simulate boundary layer clouds (Xu and Cheng 2013a). The embedded 2D CRM is the System for Atmospheric Modeling (SAM) (Khairoutdinov and Randall 2001), upgraded with an advanced third-order turbulence closure known as intermittently prognostic higher-order closure (IPHOC) (Cheng and Xu 2006, 2008). The sub-CRM grid variability is described by a double-Gaussian probability density function, which is used to diagnose cloud fraction, grid-mean cloud water mixing ratio, buoyancy terms, and fourth-order terms in the equations describing the evolution of the second- and third-order moments. As in the host climate model, SAM-IPHOC has the same 32 vertical levels and a horizontal grid spacing of 4 km. This specific configuration was proven to perform better than that from an MMF with a first-order turbulence closure and that from the standard CAM (Cheng and Xu 2011).

As in CHX13, the MMF is forced with prescribed climatological sea surface temperature and sea ice distributions with monthly-mean annual cycles. The model was integrated for 10 yr and 3 months, and we only analyzed September–November (spring) of the last integration year because the hourly output was not saved for the first 9 yr and 3 months of the integration. Good agreement between 1-yr (year 10) and 4-yr (years 6–9) averaged results of the MMF integration reported in Xu and Cheng (2013a), with low-cloud fraction correlations exceeding 0.99 and a bias of only 0.2%, demonstrates that the 1-yr simulations presented in this study are representative of multiyear simulations.

b. Satellite observations

The satellite dataset includes cloud retrievals from sun-synchronous Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO; Winker et al. 2007) and Aqua on the A-Train constellation of satellites and the geostationary Meteosat-9, along with liquid water paths derived from several satellite microwave radiometers. To provide direct measurements of cloud-top height and verify the passive satellite estimates of cloud-top height, the CALIPSO–CloudSat–Clouds and the Earth’s Radiant Energy System (CERES)–Moderate Resolution Imaging Spectroradiometer (MODIS) merged product (CCCM) of Kato et al. (2011) is used here. The CCCM matches the near-nadir Cloud–Aerosol Lidar with Orthogonal Polarization (CALIOP) 333-m data with the larger CloudSat and 1-km MODIS pixels included within the nearest ∼20 km × ∼20 km CERES scanner footprint. In the CCCM, cloud-top height is measured directly using CALIPSO’s CALIOP, whereas mean liquid cloud fraction is obtained from all of the Aqua MODIS pixels within the CERES footprint using the algorithms described by Minnis et al. (2011a). For this study, the CCCM CALIPSO and MODIS retrievals from six austral springs (September–November during 2007–12) are averaged to the 1.9° × 2.5° model grid. To be consistent with the CALIPSO footprint, the CCCM product only merges nadir-view observations; therefore, CCCM sampling is spatially sparse on a given day over the region of study. The 6 yr of data, however, should produce representative long-term averages. Since our focus is on low clouds, we only analyze single layer warm clouds with cloud tops below 3 km as determined from the CALIOP vertical mask feature included in CCCM.

Since the two Aqua/CALIPSO daily overpasses [0130 and 1330 local solar time (LST) at the equator] are inadequate to resolve the diurnal cycle, we also utilize hourly retrievals from the geostationary satellite Meteosat-9 and its Spinning Enhanced Visible and Infrared Imager (SEVIRI). Meteosat SEVIRI radiances were calibrated against MODIS and processed every hour at a 3-km pixel resolution as described by Minnis et al. (2008) and sampled every third scan line and pixel. The analysis utilizes algorithms adapted from MODIS for CERES Edition 2 (Minnis et al. 2011a). We employ observations during two austral springs in 2012 and 2013 covering the Southern Hemisphere domain. Meteosat-9 cloud fraction (CF) is calculated as the ratio between cloudy pixels and the total number of pixels within a 1.9° × 2.5° grid (model grid size).

Since direct estimates of cloud height from passive instruments such as SEVIRI are often unreliable in marine stratocumulus regimes (e.g., Garay et al. 2008), we calculate cloud-top height \( H_T \) using a linear fit between sea surface and cloud-top temperature differences \( \Delta T \) and \( H_T \) (Painemal et al. 2013). This empirical relationship was derived from in situ observations over the SE Pacific by Painemal et al. (2013) and expressed as

\[
H_T = \frac{\Delta T + 1.35}{0.0095}. \tag{1}
\]

The sea surface temperature is taken from Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) and produced by Remote Sensing Systems (Wentz et al. 2001). TMI SST primarily relies on the 10.65-GHz channel, and it has the advantage to see through clouds. A comprehensive validation analysis of TMI SST against buoy observations, including buoys...
over the SE Atlantic, revealed excellent agreement, with a small TMI bias of $-0.07\,^\circ C$ (Gentemann et al. 2004).

Equation (1) applied to the SE Pacific captured the low-cloud $H_T$ diurnal variability, yielding mean biases of 5.4 m and root-mean-square differences of 135 m relative to aircraft observations (Painemal et al. 2013). Although the lack of in situ observations prevents any error quantification of Eq. (1) in the SE Atlantic, the Meteosat $H_T$ mean agrees well with its CALIPSO counterpart, as shown later.

Liquid water path (LWP) values are taken from the University of Wisconsin (UWisc) climatology (O’Dell et al. 2008), which was derived from microwave sensors onboard seven polar-orbiting satellites ($F8$, $F10$, $F11$, $F13$, $F14$, $F15$, and $Aqua$) and one equatorial-orbiting satellite (TRMM). The UWisc climatology reports monthly-mean diurnal cycles, represented by the amplitude and phase of the 24- and 12-h cosine harmonic. An important shortcoming of this climatology is the inability to accurately separate cloud water from precipitation; therefore, the dataset is less reliable in convective clouds and near the ITCZ. In stratocumulus clouds, the precipitation bias is modest (<5%). Another source of error is clear-sky contamination within the instrument field of view. For typical cloud fractions reported here (50%–90%), Seethala and Horváth (2010) found satellite microwave LWP differences smaller than 15 g m$^{-2}$ that can be attributed to cloud fraction. Finally, statistical errors in the climatology cosine fit are typically less than 15% over the region of study.

Since the CRM embedded in the MMF has a single-moment microphysics scheme, droplet size and number concentration cannot be simulated. The microphysical characterization of the Namibia–Angola cloud regime from satellite and models will be the subject of our future endeavors.

Finally, it should be noted that, in addition to the temporal coverage differences there are also large differences in the spatial coverage between the CCCM and Meteosat-9. The latter offers spatial coverage of cloud fields comparable to that of models, while the former yields a very sparse coverage as mentioned above.

3. Results

a. Mean structure

The simulated near-surface winds are strong along the coast south of $24^\circ S$, blowing southerly parallel to the coast and turning into southeasterly trade winds far offshore (Fig. 1a, colors and black arrows). The model is consistent with National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis winds (Fig. 1a, white arrows). Moreover, the model correctly reproduces the location of the Benguela coastal low-level jet over the region where sea surface temperature is a local minimum ($20^\circ$–$25^\circ S$; Fig. 1a, blue contours) and contiguous with an elevated continental region with heights up to 1.4 km (Fig. 1b). These results are further confirmation that the model accurately reproduces large-scale parameters, such as subsidence, wind fields, and precipitable water (Xu and Cheng 2013a).

Mean liquid-phase cloud fraction during the period of study is depicted in Fig. 2. The simulated cloud fraction shows excellent agreement with MODIS in terms of magnitude and spatial distribution of the marine stratocumulus clouds (colors in Figs. 2b and 2a, respectively), even though MODIS cannot resolve the diurnal cycle.
We observe some disagreement over the southwesternmost part of the domain, where the model underestimates the observations. However, the Meteosat-9 CF shows a sharp westward cloud fraction decrease consistent with the MMF contours in Fig. 2a. As noted earlier, the Meteosat-9 and model cloud fractions are calculated in a similar manner using data from the entire diurnal cycle, whereas the MODIS results from Kato et al. (2011) were computed from twice-per-day data taken along a 20-km swath centered on the narrow CALIPSO field of view and then averaged to the model grid mesh. Although the long-term CCCM swath averages are quite comparable to those derived using the full MODIS swaths (not shown) computed as in Minnis et al. (2011b), it is expected that the Meteosat-9 CF would be more consistent with the model mean because the two sources use similar temporal sampling and averaging methods. A specific model–Meteosat-9 comparison, depicted in Fig. 2b (contours), shows that the model bias reaches $0.15$ at the edge of the stratocumulus cloud deck but remains within $0.1$ over regions with cloud fraction higher than $0.65$.

In terms of liquid water path, the model qualitatively reproduces the cloud spatial variability in observations (Fig. 3a), with a relative maximum around $12^\circ$S, $5^\circ$W and an absolute maximum associated with the intertropical convergence zone (ITCZ; Fig. 3b). On the other hand, a sharp LWP coastal decrease south of $20^\circ$S is concomitant with a cloud fraction decrease (Fig. 2). Overall, the qualitative consistency in Figs. 2 and 3 indicates that the model correctly predicts the cloud formation and occurrence. A significant model shortcoming is over-prediction of the LWP. For instance, LWP values over the ITCZ and at $12^\circ$S and $5^\circ$W surpass by $50 \text{ g m}^{-2}$ (35% and 50% overestimate, respectively) those from the climatology. We will return to this issue in the discussion section of this paper.

The CALIOP $H_T$ shows shallow boundary layer clouds alongshore, with a westward deepening and $H_T$ increase over ITCZ (Fig. 4a, colors), as in Sun-Mack et al. (2014). In addition, Meteosat-9 $H_T$ mean is consistent with CALIOP (Fig. 4a, contours), yielding differences smaller than $100 \text{ m}$ and corroborating the adequacy of the linear regression method [Eq. (1)]. We calculate the $H_T$ from the model as the highest vertical level below $700 \text{ hPa}$ in which at least 3 CRM horizontal grids of a total of 32 embedded in the climate model grid box are identified as cloudy. A more restrictive threshold does not change our main findings. Cloud-base height $H_B$ is estimated in a similar manner but in terms of the lowest vertical level instead. The MMF simulations in Fig. 4b (colors) capture the zonal and equatorward gradients in $H_T$ but yield overestimates of $200 \text{ m}$ west of $20^\circ$S and underestimates of $400–500 \text{ m}$ along the coast (Fig. 4b, contours). This negative coastal bias has also been reported over the SE Pacific in climate and mesoscale models during austral spring. It is believed that underestimates over the SE Pacific are linked to the topography representation in the model (Wyant et al. 2010). As elevations in southern Africa are relatively modest (<2 km; Fig. 1b) compared to the Andes cordillera, one should expect a limited effect in the simulations over the Atlantic. In fact, the simulated $H_T$ average is approximately $300 \text{ m}$ along the Namibia–Angola coast, whereas the underestimated $H_T$ along the Chile coast south of $20^\circ$S reaches near-surface values, below the lifting condensation level (CHX13). On the other hand, the model overestimates $H_T$ over the westernmost area of the domain (Fig. 4b, contours; $0.2 \text{ km}$) and over the ITCZ, where the low-cloud cover decreases substantially.
The simulated cloud geometrical thickness $\Delta H$ is depicted in Fig. 5 and was calculated as the difference between the cloud-top and cloud-base height ($\Delta H = H_T - H_B$) for each hourly output. The $\Delta H$ correlates spatially well with $H_T$ (Figs. 5, 4a), whereas the minimum $\Delta H$ of 200 m found south of 20°S and along the coast occurs over a region with a local minimum in CF, LWP, and $H_T$ in both the model and observations. As previously discussed, this littoral zone is characterized by the presence of an alongshore low-level jet, coastal upwelling, and cold sea surface temperature associated with the Benguela Current (Fig. 1a; Nicholson 2010). Moreover, meteorological reanalyses and satellite wind data from the Quick Scatterometer (QuikSCAT) show that strong surface divergence is responsible for a subsidence enhancement (Nicholson 2010). Strong subsidence south of 20°S is reproduced in the model, as evinced in the vertical velocity maps ($\omega = dp/dt$) at 700 hPa in Fig. 6 (colors). This feature is responsible for the sharp boundary layer shoaling and cloud thinning in Figs. 3–5 and ultimately accounts for cloud dissipation along the Benguela Current (Fig. 2). Farther north, subsidence weakens and the boundary layer deepens, changes that typically favor increasing CF and LWP.

We define the inversion strength as the difference in potential temperature between the levels above and at the cloud top, divided by the layers' vertical separation. We note that the use of cloud top in this definition simplifies the calculation over regions with weak inversions, while $H_T$ matches the inversion base height in profiles having a well-defined temperature inversion, which typically occurs in a single vertical grid. The simulated inversion strength shown in Fig. 6 (contours) is greatest over the region where CF and LWP reach a regional peak, between 8° and 15°S. This is consistent with expectations of reduced entrainment below strong inversions. Interestingly, the maxima in subsidence and inversion strength in Fig. 6 are meridionally separated by more than 10°.

Boundary layer mixing is another element to be taken into account. Numerical modeling indicates that turbulence decoupling of the cloud layer from the boundary layer can eventually lead to cloud breakup and dissipation (e.g., Bretherton and Wyant 1997). Several mechanisms can lead to decoupling such as drizzle or afternoon shortwave heating of the cloud layer. In a regional-scale context, decoupling can primarily occur when boundary layer clouds are advected over regions with warmer sea surface temperatures or through boundary layer deepening driven by entrainment (Bretherton and Wyant 1997; Wood and Bretherton 2004). Following previous studies (e.g., Jones et al. 2011), we calculated a decoupling metric defined as the $H_B$ and lifting condensation level (LCL) difference ($H_B - \text{LCL}$), which is depicted in Fig. 7 (contours). For comparison, we also show in Fig. 7 the mean simulated CF (colors). The spatial correspondence between cloud variability and decoupling is remarkable, where the region with the highest CF coincides with decoupling lower than 200 m, while the decoupling is higher than 350 m for CF < 0.6. The exception to this positive correlation is the coastal region south of 20°S where low magnitudes of CF occur with well-coupled conditions ($H_B - \text{LCL} = 50$ m). As previously explained, the Benguela low-level jet and its effect on the coastal
shoaling is likely responsible for this decrease in CF. In addition, the model also yields more decoupling in deeper boundary layers (higher cloud tops)—that is, a westward deepening consistent with an increase in $H_B - LCL$—in agreement with observations over the SE Pacific (e.g., Jones et al. 2011).

b. Diurnal cycle

Hourly CF anomalies relative to the daily mean from MMF and Meteosat-9 are shown in Figs. 8 and 9, respectively. The model predicts the morning increase and afternoon dissipation, with maxima and minima around 0500–0800 and 1400 LST, respectively. Moreover, regions with larger diurnal changes near the equator and alongshore are also correctly reproduced in the model, although the amplitudes are overestimated south of 10°S. We observe a subtle disagreement around 1700 LST, when Meteosat-9 yields a cloud cover increase over the 0°–5°S, 10°W–5°E quadrant (Fig. 9f) not reproduced by the model (Fig. 9f). An afternoon local maximum is typical over regions with tropical/continental convective clouds (Hendon and Woodberry 1993). Thus, the afternoon peak in Fig. 9f suggests that part of the afternoon near-equatorial low clouds are originated from deep convective systems, which become more pervasive in spring and summer when the ITCZ moves southward.

The explicit diurnal cycles for two regions along the 12.25°S parallel at 7.5°E and 22.5°W (magenta squares in Figs. 10b,c) are depicted in Fig. 10a. The agreement between the model and observations is remarkable in terms of magnitude and phase. As anticipated in Figs. 8 and 9, CF maximum and minimum occur at 0500–0600 and 1500 LST, respectively. The spike near 1800 LST at 22.5°W is the result of greater uncertainty in the satellite retrieval at near-terminator conditions. The Aqua MODIS CF values partially corroborate the diurnal cycle derived from Meteosat-9, although the mean coastal MODIS CF is slightly larger than its Meteosat-9 counterpart and agrees better with the modeled cycle (Fig. 10a, asterisks).

To gain a deeper regional insight, we simultaneously fitted 24- and 12-h cosine functions to the hourly composited CF and computed the respective amplitudes and phases, following Painemal et al. (2013). Unlike that over the SE Pacific, the amplitude of the semidiurnal cycle is negligible (e.g., O’Dell et al. 2008) and therefore will not be discussed. The CF amplitude of the diurnal cycle from the MMF and observations is depicted in Figs. 10b and 10c, respectively. Overall, as previously anticipated, the model overestimates the diurnal cycle amplitude, but it shows skill in simulating a negligible cycle at 12.25°S, 7.5°E along with the approximate location of two regions with maximum amplitudes near the equator and alongshore south of 15°S. The Aqua MODIS morning–afternoon CF differences, the lower bounds of the diurnal amplitude (Fig. 10a), are smaller than 0.1 over most of the domain (not shown), consistent with the Meteosat-9 results.

MMF hourly LWP anomalies are shown in Fig. 11. The overall diurnal evolution is similar to that for CF, with minimum and maximum around 0600 and 1600 LST. Specific MMF diurnal cycles along 12.25°S, as in Fig. 10a, yield contrasting differences with observations (Fig. 12b). Although the phases are consistent, the model substantially overestimates LWP, as initially anticipated by the mean field in Fig. 3. MMF morning–afternoon LWP differences reach 150 g m$^{-2}$ near the coast, compared to 60 g m$^{-2}$ in the UWisc climatology. A closer look at the
24-h harmonic amplitude reveals the details of the model overestimation (Figs. 12b,c). This is particularly severe over the northern part of the domain, where the MMF 24-h harmonic amplitude is near 72 g m$^{-2}$ but the observations yield amplitudes of only 24 g m$^{-2}$. Additionally, the largest diurnal cycle amplitudes in the model occur 10° northward relative to the observations. Although one should expect biases in the microwave retrievals, the model overestimates are larger than the typical uncertainties of the UWisc climatology described in section 2 (O’Dell et al. 2008). Since large diurnal cycle amplitudes in the model are accompanied with substantial overestimates of the mean LWP, we also assess the model performance by analyzing the LWP fractional amplitude relative to the mean (Figs. 12d,e). MMF relative amplitudes show better spatial agreement with observations and the overestimation decreases to some extent. It is also interesting that the LWP relative amplitudes spatially correlate with CF diurnal amplitudes in Figs. 10b,c, with linear correlations of 0.76 and 0.74 for MMF and observations, respectively. Large LWP relative amplitudes imply that clouds are more likely to break up during the afternoon because of solar heating.

Examples of $H_T$ diurnal cycle and regional amplitude of the 24-h harmonic are presented in Fig. 13. As observed in the mean regional map (Fig. 4), westward boundary layer deepening is apparent (Fig. 13a), and the model better reproduces the mean observed $H_T$ far offshore (Fig. 13a, red lines). Mean amplitudes are modest in the model and observations, particularly near the coast, in which the maximum–minimum differences are smaller than 100 m. Independent satellites estimates from CALIPSO corroborates the occurrence of amplitudes smaller than 100 m (Fig. 13a, asterisks), although the coastal morning–afternoon $H_T$ differences are around 20 m larger than that from Meteosat-9. As expected, cloud heights reach a peak during the early morning and decrease during the afternoon. This is qualitatively consistent with cloud fraction and LWP diurnal cycles (Figs. 10a, 12a) and in agreement with a typical marine boundary layer (Turton and Nicholls 1987). Regional amplitudes of the 24-h harmonic (Figs. 13b,d) are below 100 m over most of the domain in both the model and Meteosat-9 observations. Magnitudes over the center of the stratocumulus cloud deck at 10°S, 0° are particularly modest in both model and observations, reaching amplitudes smaller than 60 m. Model departures occur near the equator, where the clouds transition to the deep convection regime. Far offshore, reduced cloud fraction can make Meteosat-9 retrievals less reliable (e.g., Painemal et al. 2013). In agreement with the rather small $H_T$ diurnal cycle, the mean cloud geometrical thickness diurnal amplitude from the model (Fig. 13c) reaches typical values smaller than 60 m south of 10°S and up to 100 m over the equator. These zonal differences are insignificant because the model vertical grid spacing is larger. As in Fig. 13a, CALIOP (CCCM) observations also confirm the diurnal amplitude derived from Meteosat-9, with morning–afternoon differences smaller than 120 m south of 10°S (not shown) or equivalent to cosine amplitudes smaller than 60 m.

4. Discussion

a. Cloud variability and regional circulation

Since MMF is particularly accurate in simulating the mean CF, one can further use the model outputs to understand the factors that explain CF spatial variability.
Consistent with eastern Pacific simulations analyzed by Xu and Cheng (2013b), a decoupling index $H_B - LCL$, is the single parameter that best spatially correlates with CF (Fig. 7). Nevertheless, disagreements were observed south of 20°S along the coast, where the boundary layer shoaling, driven by the strong subsidence, is responsible for coastal cloud thinning and therefore decreases in cloud fraction. On the other hand, cold air advection may be slightly weak in the model (Fig. 1a) south of 20°S along the coast, where cold advection is at a maximum in observations (Zhang et al. 2010), as the simulated surface flow is more parallel to the coastline than the observed flow. This reduces the inversion strength at the boundary layer top (Fig. 6, contours).

While one should expect a link between subsidence and warming, and therefore increases in the lower tropospheric stability (LTS), traditionally defined as the potential temperature difference between 700 and 1000 hPa (Klein and Hartmann 1993), the effect of subsidence-driven boundary layer shoaling and cloud thinning explains the cloud fraction decrease south of 20°S. Additionally, the inversion strength and the region with the highest CF are coincident (Figs. 2b, 6), suggesting weak entrainment over the stratocumulus cloud center. Because the cloud diurnal cycle amplitude is a minimum over the region with the strongest inversion, one can hypothesize about the regional control of the atmospheric stability over the diurnal variations in the boundary layer. Because the nighttime boundary layer deepens through entrainment, when the water content buildup enhances the cloud-top longwave-cooling-driven turbulence, a strong inversion would limit the entrainment and therefore would dampen the nighttime boundary layer growth and thus reduce the diurnal cycle of $H_T$. During daytime, solar heating is the primary driver of cloud thinning, whereas the boundary layer decoupling/coupling strength controls the water vapor turbulent fluxes available for condensation. Because a highly coupled boundary layer more efficiently connects water vapor fluxes to the in-cloud turbulence, the LWP recovery during the late afternoon is enhanced over areas with strong (weak) coupling (decoupling) (Fig. 6, contours), producing a large absolute amplitude in LWP (Fig. 12c). A counterintuitive finding is that the area with the strongest inversion is located almost 10° north of the maximum subsidence region (Fig. 6). Similarly, Medeiros (2011) reported a northward shift of the Namibia–Angola stratocumulus cloud deck relative to the region with the strongest LTS, whereas Myers and Norris (2013) found similar spatial disagreements with another stability metric. The meridional inconsistency between CF and LTS is also reproduced by the model (Cheng and Xu 2011). This mismatch could be related to the northward lowering of the Namibia–Angola topography north of 10°S (Fig. 1b), which favors the occurrence of stronger offshore winds (Krishnamurti et al. 1993) and above-inversion continental warm advection that contributes to the strength of the temperature inversion. Farther south, the blocking effect of coastal topography (Fig. 1b) should limit the horizontal temperature advection in the lower troposphere. Moreover, as previously mentioned, the boundary layer height lowering and its impact on cloud thinning might be a relevant component in explaining a cloud fraction decrease despite increases in LTS south of 20°S.

b. Namibia–Angola and Chile–Peru stratocumulus clouds

Assessing the significance of the SE Atlantic diurnal cycle merits an integrated analysis of the two major Southern Hemisphere marine boundary layer clouds. Although an in-depth analysis deserves the extent of a stand-alone paper, our goal here is to draw attention to the need of understanding specific regional traits, as they are key to understanding the subtropical boundary layer response to future climate scenarios.

The LWP climatological mean and 24-h harmonic amplitude are depicted in Fig. 14 (colors and contours, respectively). Spring maximum mean LWP in the SE Pacific is 20 g m$^{-2}$ (20%) larger than those in the SE Atlantic. It is also interesting that the minimum coastal LWP in the SE Atlantic has a broader extension. This is consistent with a low-level coastal jet that is more intense and meridionally extended than its Chilean counterpart (Ranjha et al. 2013). In terms of diurnal cycle, the SE Pacific has a LWP diurnal harmonic that is
20–25 g m$^{-2}$ larger than equivalent maxima in the SE Atlantic: that is, the amplitude in the SE Atlantic is only half that observed off the coast of Chile–Peru. Moreover, as reported in Painemal et al. (2013) for October–November 2008, the $H_T$ amplitude of the 24-h harmonic over the SE Pacific reaches magnitudes up to 140 m. This is around 40 m higher than the diurnal harmonic over the core of the stratocumulus cloud deck in the SE Atlantic east of 15°W (Fig. 13d). For a more specific comparison, we select two regions, at 20°S, 84°W (SE Pacific) and 12°S, 5°W (SE Atlantic), and compare their diurnal cycles. These two regions have climatological peaks in CF and LWP, as well as broad amplitude of the LWP 24-h harmonic (Fig. 14a). For the SE Pacific, we used Geostationary Operational Environmental Satellite-10 (GOES-10) $H_T$ retrievals for October–November 2008 calculated in a similar manner as we did with Meteosat-9, following Painemal et al. (2013). Since both satellite instruments were calibrated against MODIS channels, the retrievals are deemed to be comparable (Minnis et al. 2008). Finally, although some differences might arise because of period differences between Meteosat-9 and GOES-10, GOES-10 retrievals used here and in Painemal et al. (2013) accurately capture all the main cloud features in terms of mean $H_T$, phase, and amplitude of the diurnal cycle during spring (Painemal et al. 2013). The overall mean $H_T$ over the SE Pacific is 250–300 m higher than that in the SE Atlantic, in agreement with Leon et al. (2008) and Zuidema et al. (2009). Moreover, consistent with LWP diurnal amplitudes, the mean amplitude of the $H_T$ diurnal cosine is also larger over the SE Pacific (Fig. 14b). The $H_T$ maximum–minimum difference is 200 m for the SE Pacific, whereas in the SE Atlantic this does not exceed 120 m. Overall, the SE Pacific stratocumulus cloud deck tends to be thicker and the boundary layer deeper, with a more prominent diurnal cycle as well.

In agreement with the aforementioned differences, 24-h harmonic cloud fraction amplitudes are less than 6% (CF $\times$ 100) off the coast of Angola, whereas their Chilean counterparts are around 10%–12% (Painemal et al. 2013). The blocking effect of the zonal flow by the Andes cordillera is likely the most determinant factor that explains the boundary layer deepening (Xu et al. 2004). This deepening is also important for yielding cloud thickening and liquid water path increases. The dominant control of the Andes cordillera is also manifested in a strong diurnal cycle of subsidence, with a perturbation that propagates westward as a gravity wave. Observational evidence and numerical models show that the diurnal cycle of subsidence can enhance the diurnal cycle of LWP and $H_T$ offshore from the South American coast (Garreaud and Muñoz 2004; Painemal et al. 2013). This subsidence wave also gives

![Fig. 8. Austral spring simulated hourly low-level CF anomalies relative to the daily mean.](http://journals.ametsoc.org/doi/abs/10.1175/JCLI-D-14-00368.1)
rises to a semidiurnal cycle in LWP and $H_T$ (Painemal et al. 2013) and modifies the amplitude and phase of the solar-driven diurnal cycle. In contrast, the semidiurnal cycle over the SE Atlantic is negligible relative to the dominant amplitude of the 24-h cycle.

A preliminary short-term forecast using CAM, version 5, by Medeiros (2011) results in cloud-top height differences that are qualitatively in agreement with Fig. 14, although, unlike Fig. 14a, the model predicts less cloud water content over the SE Pacific than that in the SE Atlantic. SE Pacific/Atlantic differences simulated by MMF are difficult to quantify because the model places the Chile–Peru stratocumulus cloud deck 10° northward compared with observations. In addition, the Andes cordillera poses a major problem when simulating near-coastal clouds (CHX13). For these reasons, we evaluate the model ability to reproduce interregional differences by comparing two regions in the SE Atlantic and Pacific where the model predicts maximum liquid water path and cloud fraction. For the SE Pacific, we selected a model grid centered at 6.6°S, 95°W, whereas a 10.4°S, 2.5°W grid was chosen for the SE Atlantic (Fig. 15). These two areas are situated approximately 15° west from the coast. Aside from the model overestimation of LWP, the model is qualitatively consistent with the observations: that is, SE Pacific LWP possesses larger mean and diurnal cycle amplitude (Fig. 15a).

Concomitant with LWP, cloud-top height off the Peru coast is on average 210 m higher than its SE Atlantic counterpart, with a broader amplitude as well (Fig. 15b, black solid and dashed lines). Based on these results, one can hypothesize that the deeper boundary layer over SE Pacific should be more conducive to decoupling. Nevertheless, the decoupling strength ($H_B - LCL$) is similar in both regions, with values near 150 m (see Fig. 10 in Xu and Cheng 2013b; see also Fig. 7 herein). This implies that SE Pacific clouds are sufficiently thicker to counterbalance a potential decoupling due to boundary layer deepening, as indicated by Fig. 15b (gray solid and dashed lines). A deeper and well-coupled boundary layer would explain the occurrence of larger LWP in the SE Pacific (Fig. 15a).

c. Liquid water path bias

Although MMF qualitatively reproduces the main structure and diurnal cycle in CF, LWP, and $H_T$, the main drawback of the model is the substantial overestimation of LWP and its diurnal cycle. This overestimation also produces a larger shortwave radiative forcing than satellite observations at the top of the atmosphere, especially where the modeled CF matches the observations (Xu and Cheng 2013a). Determining the factors that account for the LWP positive bias is difficult as many intermingling processes can affect the
MMF performance. The model vertical resolution is certainly a factor that needs to be taken into consideration because higher-resolution simulations would better resolve the temperature inversion and cloud-top entrainment, helping to reduce the cloud water content. In this regard, sensitivity tests by Xu and Cheng (2013a) suggest that modest improvements are attained when the vertical levels below 700 hPa increase from the original 10 to the current 12 levels. It would be interesting to revisit this analysis by testing the vertical-resolution sensitivity of the simulations to more diverse environmental conditions. While sensitivity analysis suggests that the water content overestimate problem can be ameliorated with the use of a finer spatial resolution, especially in cumulus regions with weak inversions, it is not clear that improvements can be achieved over regions with strong temperature inversions (Cheng and Xu 2008).

The turbulence predicted by IPHOC is another plausible process that might account for the liquid water path overestimation. For instance, when compared to MMF-IPHOC with a standard MMF that included a first-order turbulence closure, Cheng and Xu (2011) found that the vertically integrated subgrid-scale turbulent eddy kinetic energy within the boundary layer in the IPHOC simulations was at least double that from the standard first-order turbulence closure simulation over stratocumulus–cumulus regimes and in the extratropics. These regions also had the most dramatic liquid water path increase in the IPHOC simulations, due in part to the increase of cloudy areas, with magnitudes 30–40 g m$^{-2}$ larger than those from a standard MMF over the Namibia–Angola stratocumulus regime core. While one can point to IPHOC for the LWP overestimation, IPHOC is also a key component in the overall improvement of marine boundary layer cloud simulations in terms of extent, location, and cloud fraction (Cheng and Xu 2011), so the problem of determining the optimal model configuration is difficult. An aspect that requires future research is the representativeness of the
probability density function to account for the subgrid-scale variability.

A more challenging issue from a modeling perspective is the representation of light precipitation and drizzle. Several large eddy simulation-based studies indicate that precipitation can reduce the liquid water path (e.g., Ackerman et al. 2009). Moreover, light precipitation can generate boundary layer decoupling and thus diminish the generation of turbulent kinetic energy (Stevens et al. 1998). Surface precipitation comparisons between MMF and the Global Precipitation Climatology Project (Adler et al. 2003) reveal a model underestimate of around 0.05 mm day$^{-1}$ over the Namibia–Angola stratocumulus cloud deck (Xu and Cheng 2013a). Additionally, drizzle proxies derived from the satellite CloudSat radar show a nighttime frequency of drizzle occurrence of 35%–40% over the Namibia–Angola stratocumulus cloud core (Leon et al. 2008). The small radar signal suggests that a significant amount of drizzle evaporates before reaching the surface. Unfortunately, the lack of in situ observations prevents a more specific assessment of the precipitation prevalence and its control of the boundary layer. Without more specific details, it is difficult to assess the model representation of light precipitation. Nevertheless, given the simplified microphysical treatment used here, it is unlikely that the details of drizzle are well captured in the model. The use of a double-moment microphysics scheme in the model is a promising approach (Wang et al. 2011), although its implementation in the context of this MMF-IPHOC requires further evaluation as this might greatly increase the simulation’s computational cost. Alternatively, the implementation of a subgrid-scale microphysics scheme, such as that proposed by Cheng and Xu (2009), would
enable the extra liquid water to be removed by the subgrid-scale precipitation.

5. Summary

We have described the mean structure and diurnal cycle of marine boundary layer clouds over the SE Atlantic, the most poorly documented stratocumulus cloud region in the subtropics. This work follows a series of studies using MMF simulations with an emphasis on low clouds by Xu and Cheng (2013a,b) and CHX13 but extended to analyze the unique characteristics of Namibia–Angola marine stratocumulus clouds. Satellite retrievals corroborate the main spatial–temporal structure of a typical marine boundary layer cloud regime that is characterized by westward boundary layer deepening and diurnal cycles in phase among \( H_T \), \( C_F \), and \( LWP \) with maxima during the early morning and minima in the afternoon around 1500 LST. The model captured all these features, along with a weak diurnal cycle near \( 10^\circ S, 0^\circ \). A clear improvement of MMF compared with other climate models is the excellent simulation of the mean cloud fraction over the region. Moreover, the model also captures the westward variation in cloud-top height, along with the region where \( LWP \) maximizes. The main model shortcoming is the 50% overestimation...
of LWP relative to satellite observations. As the model accurately reproduces the mean CF, we used the model to understand the factors that account for cloud variability. The parameter that best spatially correlates with cloud fraction is the decoupling strength ($H_B - LCL$).

The core of the stratocumulus cloud deck occurs under a region where the temperature inversion is the strongest. In contrast, a coarser stability metric, LTS (Klein and Hartmann 1993), estimated from both MMF-IPHOC and reanalysis (Fig. 5 in Cheng and Xu 2011), shows stable conditions that extend farther south relative to the core of the stratocumulus deck and the temperature inversion strength. Although it is possible that a subtle underestimate of cold advection by the model (Fig. 1a) plays a role in the small CF underestimate south of 25°S, thus slightly degrading the LTS–CF correlation, a southward extent of LTS relative to the stratocumulus cloud core was also observed in satellite climatologies and meteorological reanalysis (Myers and Norris 2013).

A first-of-its-kind comparison between the two most climatically important Southern Hemisphere stratocumulus cloud regimes reveal substantial differences. A shallow boundary layer yields reduced diurnal cycles in the southeast Atlantic, with substantial coastal shoaling driven by the dynamics of the low-level jet. The model suggests that the boundary layer deepening in the SE Pacific occurs under relatively well-coupled conditions, which could explain why a deeper boundary layer depth in the SE Pacific yield larger liquid water path as well. It is also encouraging that the model qualitatively reproduces the cloud regime differences, indicating that

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**Fig. 14.** (a) UWisc LWP (color) and amplitude of the 24-h harmonics (contours) during spring. (b) SE Pacific $H_T$ from GOES-10 (red) as in Painemal et al. (2013) and SE Atlantic $H_T$ from Meteosat-9 (black). White crosses in (a) indicate the regions selected to create (b).

**Fig. 15.** Modeled austral spring diurnal cycles of two regions with maximum mean CF and LWP over the SE Pacific (6.6°S, 95°W) and SE Atlantic (10.4°S, 2.5°W): (a) LWP and (b) $H_T$ (left y axis) and $\Delta H$ (right y axis).
the model will be a valuable tool for a more in-depth evaluation of the processes that govern the SE Atlantic and Pacific marine low-cloud variability.

The SE Atlantic cloud microphysical features were not addressed in this investigation. They are particularly complex over this region because, unlike other marine boundary layer clouds, seasonal biomass burning and the subsequent massive transport of aerosols that overlie the stratocumulus clouds deck during June–September can yield strong interactions between cloud and aerosols (e.g., Kaufman et al. 2005; Painemal et al. 2014). Moreover, the rather strong land–sea breeze circulation (e.g., Gille et al. 2005) should modulate the diurnal cycle in cloud droplet size and number concentration. A modeling approach will require a more complex microphysics scheme than the one used here; it will need to account for the aerosol chemical composition. While the microphysical representation is an aspect that should be improved, it is encouraging the MMF simulations are able to reproduce the boundary layer main features, despite the oversimplification of the cloud microphysics.

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