Simulating Radiative Fluxes through Southeastern Pacific Stratocumulus Clouds during VOCALS-REx

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(Manuscript received 26 September 2017, in final form 18 December 2017)

ABSTRACT

Time series of solar and thermal infrared radiative flux profiles are simulated with the Rapid Radiative Transfer Model (RRTM) using a hierarchy of constraints from radar reflectivity and passive microwave cloud remote sensing measurements collected over a ship in the southeastern tropical Pacific Ocean (20°S) during the second leg of the Variability of American Monsoon Systems (VAMOS) Ocean–Cloud–Atmosphere–Land Study Regional Experiment (VOCALS-REx). Incorporating additional constraints results in simulations of physically consistent radiative profiles throughout the atmosphere, especially within the cloud, where they are difficult to observe precisely. Simulated surface radiative fluxes are compared with those observed on the ship and by aircraft.

Due to the strong Rayleigh scattering of drizzle drops compared to cloud droplets that absorb, emit, and scatter natural radiation, cloud radar reflectivity overestimates cloud liquid water content (LWC). As a result, clouds are optically too thick and transmission ratios are too low in simulations using radar LWC. Imposing a triangular (increasing linearly with height from zero at cloud base) LWC profile in agreement with microwave liquid water path (LWP) improves the simulation of the transmission ratio. Constraining the corresponding microphysical cloud effective radius to that retrieved from optical depth, LWP, and cloud thickness results in additional improvements to the simulations. Time series, averages, and composite diurnal cycles of radiative fluxes, heating rates, and cloud radiative forcing are presented.

1. Introduction

The extensive subtropical stratocumulus cloud layer over the southeastern Pacific Ocean, west of the coasts of Chile and Peru (Klein and Hartmann 1993), has a strong influence on Earth’s radiation budget. These clouds have a strong net cooling effect as a result of their high albedo reflecting incoming solar radiation and a weak longwave effect as a result of their high temperatures. Their spatial scales are poorly resolved by weather and climate models, necessitating parameterization of processes important to the clouds. Cloud radiative forcing, as well as cloud microphysical and dynamical processes, remains as one of the largest sources of uncertainty in projecting future climate (Bony et al. 2006). Model simulations show differing responses by boundary layer clouds to such forcing factors as increasing sea surface temperatures (Zhang et al. 2013), greenhouse gases (Bretherton et al. 2013), and aerosol properties (Caldwell and Bretherton 2009); therefore, models require better understanding of the elements that control the life cycle (i.e., formation, maintenance, and dissipation) of marine boundary layer (MBL) clouds.

Warm marine stratiform clouds have long been a target for simulation with radiative transfer schemes (e.g., Liao and Sassen 1994; Mlawer et al. 1997; Mather et al. 2007; Ghate et al. 2014). Relationships derived between radar reflectivity, effective radius ($r_e$), and liquid water content (LWC) (e.g., Mather et al. 2007; Matrosov et al. 2004) can be provided to radiative transfer models to generate estimates of cloud albedo. Empirical relationships are accurate for nonprecipitating stratiform clouds, such as radar reflectivity increases as $Z \sim D^6$, while LWC increases more slowly as $LWC \sim D^3$, where $D$ is drop diameter. The relationships for cloud LWC are not valid when radar reflectivity is dominated by drizzle. Large-diameter drizzle drops radically increase the radar reflectivity, so the validity of radiative retrievals based on cloud radar is limited to specifically nonprecipitating clouds (Papatsoris 1994; Fox and Illingworth 1997).

Our aim is to model physically accurate radiative fluxes within stratocumulus clouds observed over the

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DOI: 10.1175/JTECH-D-17-0169.1

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southeastern tropical Pacific Ocean. Prior studies have used observations and models to study the impact of uncertainties when measuring cloud properties and to compare various retrieval methods of cloud radiative forcing and heating rate profiles (Comstock et al. 2013). In this study a hierarchy of methods is used to run the radiation retrieval in order to simulate the vertical structure of radiative fluxes in as physically consistent manner as possible. The resultant simulated fields are then examined. Data simulated by the Atmospheric and Environmental Research, Inc., Rapid Radiative Transfer Model (RRTM) include shortwave and longwave fluxes, for clear and cloudy columns, from which atmospheric heating rates and cloud radiative forcing can be inferred.

It is essential to understand the fundamental structure and properties of stratocumulus in order to further study boundary layer processes. The diurnal cycle of shortwave radiation plays an important role in diurnal cycles of turbulence and precipitation (Burleyson et al. 2013), and during daylight it influences sea surface temperatures and suppresses ocean mixing (Colas et al. 2012; Colbo and Weller 2007). Radiative flux divergence determines radiative heating rates and balances the divergence of buoyancy flux by cloud motions. This balance constrains the generation of turbulence kinetic energy (TKE) by buoyancy flux, which drives entrainment (e.g., Tjernström and Rogers 1996; Lock and Macvean 1999; Caldwell et al. 2005; Ghate et al. 2014). Cloud-top radiative heating rates and subcloud buoyancy flux affect feedbacks that stabilize and destabilize the atmospheric boundary layer, leading to boundary layers coupled or decoupled to the surface layer via turbulent mixing (Turton and Nicholls 1987; Bretherton and Wyant 1997). If the marine boundary layer is coupled, then turbulence and positive buoyancy flux driven by cloud-top radiative cooling are connected through the whole boundary layer to the surface (de Szoeke et al. 2012).

Improving knowledge of atmospheric turbulence and heating profiles throughout the depth of the boundary layer helps to inform numerical weather prediction models employed in forecasting cloud-base heights and surface visibility, quantities which have broad impacts on planning, safety, and economics for the marine and aviation communities (Baker et al. 2002; Keith and Leyton 2007). Given the proper structure of radiative fluxes as initializations, large-eddy simulations can be employed to study diurnal cycles of entrainment and decoupling (Mechem et al. 2012; Lock 1998). In addition, combining knowledge of boundary layer structure and processes allows for much needed improvement in cloud parameterizations for incorporation into global climate models (Hannay et al. 2009; Wyant et al. 2010) to improve the representation of marine stratocumulus and future climate projections.

Section 2 introduces the study location, instrumentation, and radiative transfer model, followed by four methods that progressively combine radar reflectivity and passive microwave cloud remote sensing measurements to test the sensitivity of the model to the representation of microphysical properties of clouds. Section 3 presents validation statistics for the sensitivity experiments discussed in the previous section. Section 4 presents average diurnal cycles and vertical profiles of simulated radiative fluxes, as well as cloud radiative forcing. Section 5 concludes with a summary.

2. Data and methods

a. The VOCALS-REx

Data for this study were collected as part of the Variability of American Monsoon Systems (VAMOS) Ocean–Cloud–Atmosphere–Land Study Regional Experiment (VOCALS-REx; Wood et al. 2011). VOCALS-REx was an international experiment during October and November 2008 in the southeast Pacific Ocean off the coastline of southern Peru and northern Chile. Skies were observed to be overcast 80% of the time during the experiment, providing an excellent opportunity to observe marine stratocumulus clouds (de Szoeke et al. 2012; Wood et al. 2011). This study focuses on observations collected during the second leg of the NOAA R/V Ronald H. Brown (RHB) cruise, 14–30 November 2008, when the W-band (94 GHz) Doppler cloud radar was operational. Clouds were continuously remotely sampled by sensors on the ship throughout the 16-day study period. Therefore, these data provide a robust picture of the stratocumulus cloud deck diurnal cycle.

Instrumentation of primary interest to this study were aboard the RHB and include the NOAA vertically pointing motion-stabilized W-band Doppler cloud radar (Moran et al. 2012), a laser ceilometer, a passive microwave radiometer (Zuidema et al. 2005), surface meteorology, and rawinsondes launched every 4 h during the cruise. The instrumentation aboard the RHB is described in de Szoeke et al. (2012, appendix A).

Shortwave and longwave flux profiles are computed every 10 min based on the framework of Ghate et al. (2014) using the Atmospheric and Environmental Research, Inc., RRTM (Mlawer et al. 1997; Mlawer and Clough 1998). The experiments showed results to be insensitive to higher temporal resolution. The vertical resolution of the simulations is 25 m from the surface to 3 km so as to match the cloud radar resolution, then 50 m from 3 to 15 km, and 1 km from 16 to 60 km. Carbon
dioxide concentration was set to 400 ppm, and other greenhouse gas concentrations were prescribed based on recent measurements. Rawinsonde data were interpolated between launches to provide continuous temperature and humidity profiles. Cloud-base height was determined using the 85th percentile from a laser ceilometer (de Szoeke et al. 2010). Cloud-top height was determined as the 10-min running average of the 1-min mean height above which the radar reflectivity falls below a threshold minimum detectable signal of the W-band cloud radar (de Szoeke et al. 2010). A minimum of three cloudy retrievals in 1 min was required for a cloud-top height to be retrieved. Regardless of when drizzle is present, the uncertainty of the cloud-top height is 25 m. Measurements are physically combined through four different methods (Fig. 1) as discussed in section 2b.

b. Representation of clouds to the radiative transfer model

To simulate radiative fluxes to support the study of turbulence in the MBL, we first specify the cloud to the RRTM using a commonly employed relationship between cloud radar reflectivity and LWC. Two more methods sequentially incorporate additional measured quantities, thereby reducing tuning. The fourth and final method prescribes a height-dependent cloud droplet effective radius. The effective radius is retrieved using liquid water path (LWP), cloud depth, solar zenith angle, transmission ratio, and cloud optical depth (Dong et al. 1998).

1) METHOD Z: LWC AS A FUNCTION OF REFLECTIVITY

The first method uses observations of cloud radar reflectivity. LWC is parameterized as a function of reflectivity $Z$ following Liao and Sassen (1994),

$$\text{LWC}_Z = 2.4Z^{1/2}.$$  

Effective radius is calculated as in Eq. (2).

2) METHOD ZL: CONSTRAIN COLUMN–INTEGRATED LWC BY MEASURED LWP

The second method for preparing the cloud data prescribed to the RRTM incorporates LWP derived from the ship-based passive microwave radiometer (LWP_{mic}). Since LWP is the column-integrated LWC, the LWC$_Z$ predicted by the radar reflectivity [Eq. (1)] is scaled everywhere by the ratio of retrieved LWP$_{mic}$ to the radar LWP$_Z$ in order to match the LWP retrieved from measurements made by the MWR,

$$\text{LWC}_{Z,\text{adj}} = \frac{\text{LWP}_{mic}}{\text{LWP}_Z} \times (2.4Z^{1/2}).$$

3) METHOD TL: IMPOSED LWC TRIANGLE

The third method imposes a triangular LWC profile within the cloud layer (Fig. 2). LWC ($\text{g m}^{-3}$) is constrained by retrieved LWP$_{mic}$ and has a maximum at cloud top decreasing linearly with cloud depth to zero at cloud base. A representative profile is created for every 10-min block of time employing 10-min averaged retrieved LWP$_{mic}$, cloud-top heights, and cloud-base heights. Because of the imposed LWC profile, this profile is insensitive to drizzle in the cloud radar reflectivity that can be found throughout the cloud, particularly at lower levels. The contribution to LWP by drizzle for clouds similar to those included in this study has been shown to be below the accuracy of the microwave radiometer (Zuidema et al. 2005), and thus it is considered negligible. A triangular profile is supported by modeling (Liao and Sassen 1994) and observational (Albrecht et al. 1995; Fox and Illingworth 1997) studies of marine clouds. Effective radius is calculated as in Eq. (2).

4) METHOD TR: CONVERT UNIFORM EFFECTIVE RADIUS PROFILE TO A VERTICALLY VARYING EFFECTIVE RADIUS PROFILE CONSISTENT WITH TRIANGULAR LWC PROFILE

The fourth method transforms the uniform-in-height effective radius retrieval developed by Dong et al. (1997, 1998, hereafter D97 and D98, respectively) to a vertically varying effective radius consistent with a triangular LWC and constant $N_d$ throughout the cloud layer. The
effective radius is calculated as an empirical function of LWP, transmission ratio $\gamma$, and cosine of the solar zenith angle $\mu_0$ [D98, their Eq. (9)],

$$r_e = -2.07 + 2.49 \frac{\text{LWP}}{\text{mic}} + 10.25 \gamma - 0.25 \mu_0 + 20.28 \frac{\text{LWP}}{\text{mic}} \gamma - 3.14 \frac{\text{LWP}}{\text{mic}} \mu_0,$$

where the units of LWP are 100 g m$^{-2}$ and $r_e$ are in micrometers ($\mu$m). The transmission ratio is calculated as a ratio of ship-based pyranometer-observed surface shortwave radiation to RRTM-simulated clear-sky surface shortwave radiation.

To achieve an optically accurate radiative retrieval, the cross section $\pi \int r_e^2 dz$ should be proportional to the cloud optical depth $\tau$. The LWC = 4$pr_e^3$ is proportional to the height above cloud base $z' = (z - z_{\text{CB}})$. Therefore, $r_e = \Gamma_r(z)^{1/3}$, where $\Gamma_r$ is the constant of proportionality. Thus,

$$\int r_e^2 dz = \Gamma_r^2 \int_0^{\Delta z} (z')^{3/5} dz' = \Gamma_r^2 \left( \frac{3}{5} \right) \Delta z / 5,$$

where the radius at cloud top $r_{e,\text{top}} = \Gamma_r(\Delta z)^{1/3}$ and

$$(r_{e,\text{top}})^2 = \Gamma_r^2 (\Delta z)^{2/3},$$

where the cloud depth $\Delta z = z_{\text{top}} - z_{\text{CB}}$.

Conservation of water is important to accurately represent surface transmission ratios; thus, column LWP must be preserved. This is done via water volume,

$$\int_0^{\Delta z} r_e^2 dz' = \int_0^{\Delta z} r_e^3 dz' = \int_0^{\Delta z} \Gamma_r^2 z' dz',$$

where $r_e$ is the effective radius that varies with height above the cloud base. Integrating this throughout the depth of the cloud gives

$$r_{e,\text{top}}^3 \Delta z = \frac{1}{2} \Gamma_r^2 (\Delta z)^2 = \frac{1}{2} \Gamma_r^2 (z_{\text{top}} - z_{\text{CB}})^2,$$

where $r_{e,\text{top}}$ is the effective radius at the top of the nonuniform cloud. Thus,

$$r_{e,\text{top}} = 2^{1/3} r_e,$$

such that the effective radius at the top of a vertically varying cloud is simply equal to a constant multiplied by the integrated uniform effective radius. Substituting this relationship into Eq. (6) we get the constant of proportionality,

$$\Gamma_r(\Delta z)^{1/3} = 2^{1/3} r_e^2,$$

and can further solve

$$r_e = r_e \left( \frac{2z'}{(\Delta z)} \right)^{1/3},$$

where $r_e$ is the effective radius at any height above cloud base $z'$ in a cloud with a triangular LWC profile.

By incorporating this height-dependent $r_e$, the equations for $\tau$ from D97 and $N_d$ from D98 are likewise transformed for application to a vertically varying cloud and to remove a dependence on (the unknown) droplet number concentration $N_d$. From Eqs. (2) and (4) in D97,

$$\tau = Q_s \pi \exp(2\sigma^2) \frac{N_d}{\text{LWP,uniform}} \exp(10\sigma^2/2) \Delta z,$$

and $N_d$ is estimated, as in D98, resulting in

$$\tau = Q_s \pi \frac{3 \text{LWP}}{4\pi r_w \text{LWP,uniform}}.$$

It follows from Eq. (9) that

$$r_{e,\text{top}} = \frac{Q_s \pi 3 \text{LWP}}{4\pi r_w^{1/3} \tau},$$

where $r_{e,\text{top}}$ is a microphysical parameter needed to accurately solve $r_e$ throughout the cloudy column. Substituting in Eqs. (9) and (11) and assuming the extinction efficiency $Q_s$ equals 2, one obtains

FIG. 2. Schematic of method TL. The LWC profile (thick black line) has a maximum at cloud top ($Z_{\text{CT}}$) and decreases linearly to 0 g m$^{-2}$ at cloud base ($Z_{\text{CB}}$). The integrated LWC for each 10-min profile is equal to the retrieved LWP from 10-min averaged microwave radiometer measurements.
where $r_e$ is the vertically varying effective radius. This $r_e$ that increases with height is computed for each 10-min realization as a function of 10-min average $LWP_{mic}$ and $\tau$ taken from the southeastern tropical Pacific synthesis dataset (de Szoeke et al. 2012). $\Delta z$ is calculated from cloud boundaries as discussed above, and $\zeta'$. The effective radius at the midpoint of each vertical radar range gate is used as the $r_e$ representative of the entire bin.

During night, Eq. (4) is not applicable because of its use of the transmission ratio. Therefore, Eq. (2) utilizing a triangular LWC profile consistent with $LWP_{mic}$ is employed during nighttime to calculate the effective radius. This is of minor consequence, as shortwave radiation is unimportant during night and longwave radiation is largely insensitive to effective radius.

### 3. Evaluation of radiative transfer simulations

RRTM simulated shortwave (SW) and longwave (LW) radiative fluxes through the depth of the atmosphere. There are no continuous observations of radiation throughout the atmospheric column. Simulated SW fluxes are validated only at the surface, against SW fluxes observed by pyranometers on a ship (filled gray area, Figs. 3 and 4) for the entire time series. Simulated clear-sky conditions are shown by the thick black line. Simulated full-sky surface shortwave downwelling (SWD) fluxes values are weighted averages of the clear-sky and cloudy-sky simulation, using the overhead cloud fraction as in de Szoeke et al. (2012). For reference, the W-band Doppler radar observed reflectivity is plotted in Figs. 3a and 4a, with cloud boundaries in gray.

Compared to surface observations, Method Z (Figs. 3b and 4b) results in lower simulated SWD fluxes during instances of high radar reflectivity (associated with precipitation; e.g., on days 319, 326, 331, 332). This method results in simulated SWD higher than observed when there are thin clouds (e.g., days 320, 323, 324). The instances of undersimulation and oversimulation of SWD are especially evident when simulated values are compared against observed values (Fig. 5). Figure 5 (top left) highlights the lack of tracking of simulated values with observed values and spread in simulated values at any observed SWD value. The inconsistencies within observations may be due to a combination of the assumed water content relation to radar reflectivity employed.
[Eq. (1)] and fixed $N_d$. In these instances of SW under-simulation, the amount of water is being distributed over a constant but too high $N_d$, resulting in too much solar radiation being scattered. To further examine the behavior of the simulation and to remove the sensitivity to time of day, simulated versus observed transmission ratio $\gamma$ is examined (Fig. 6). Method Z undersimulates the transmission ratio at the lowest ratio while consistently oversimulating for $\gamma \approx 0.6$. In the middling transmission ratios, method Z has a wide spread. This wide spread is indicative of the large variation (and disagreement) in simulated all-sky SWD when there are observed midrange transmission ratios. Times when $\gamma_{\text{simulated}}$ is exactly 1 are when there were no clouds modeled. Instances of no cloud ($\gamma = 1$) may not match up perfectly with the observed transmission ratio as a result of differing view angles—in this case, there is a very narrow overhead view of field for the ceilometer versus the hemispheric view of the pyranometer—and also because of the detection of cloud boundaries. If there is no radar reflectivity, it is considered a clear-sky retrieval (i.e., no cloud observed by the radar), regardless of what the wider field of view cloud fraction may be. Throughout the dataset, method Z has a mean offset of 6 W m$^{-2}$, a root-mean-squared error (RMSE) of 124 W m$^{-2}$, and a fraction of variance explained ($r^2$) of 0.87 with respect to the surface observations made by the microwave radiometer (Table 1).

Method ZL (Figs. 3c and 4c) shows some improvement over method Z, but it still results in a model that underestimates surface SWD at times during high reflectivity (days 331, 332) and simulates SWD fluxes greater than those observed during thin clouds (day 320, 323, 334). Overall, simulated SWD using method ZL in comparison to observed SWD yields a mean offset of 7 W m$^{-2}$, an RMSE of 86 W m$^{-2}$, and an $r^2$ of 0.93 (Table 1). From Fig. 5 (top right) it is clear that there are a few strong outliers with large individual errors, despite the low mean offset and RMSE. Simulated transmission ratios (Fig. 6, top right) show no clear bias but rather are generally symmetric about the 1:1 line.

The first two methods calculate LWC from an empirical relationship with $Z$ [Eq. (1)]. Since $Z$ is proportional to the sixth power of droplet radius $r^6$, LWC is disproportionately affected by large precipitation-sized particles, in this case virga and drizzle. These large particles have low solar scattering cross sections despite their high radar reflectivity. However, because of how inputs are specified, the RRTM does not always simulate precipitation-dominated layers with these same properties. For example, as a result of the prescribed droplet number concentration, the calculated LWC in the precipitation-dominated layers may be spread over more droplets than would have been observed (i.e., the model simulating many small drops rather than a few large precipitation drops), resulting in atmospheric
layers having greater total scattering. The additional scattering within the atmospheric column will decrease simulated surface SW radiative fluxes below observed values. On account of the reliance on the reflectivity profile, the vertical structure of radiative fluxes can be inaccurately impacted as a result of the model simulating radiatively important layers far below cloud base.

Except to locate the cloud top, method TL abandons the radar reflectivity, instead imposing a triangular LWC profile. The time series of simulated fluxes employing method TL versus observed surface SWD fluxes (Figs. 3d and 4d) show that many of the instances of fluxes simulated by method Z and method ZL that underestimated SWD are resolved in this method. However, simulations in which SWD is overestimated persist (e.g., days 320, 324, 334) and new overestimates occur (e.g., day 325 near local noon). The systematic overestimation is evident in the comparison of simulated and observed surface SWD fluxes and the best-fit line with a nearly constant offset from the 1:1 line (Fig. 5, bottom left). Likewise, this overestimation is apparent in the comparison of simulated and observed transmission ratios (Fig. 6, bottom left) with the majority of the simulated transmission ratios located above the 1:1 line. The surface SWD simulated from this method has a mean offset of 65 W m$^{-2}$, an RMSE of 129 W m$^{-2}$, and an $r^2$ value of 0.89, relative to that observed.

Method TR has a mean offset of 47 W m$^{-2}$, an RMSE of 112 W m$^{-2}$, and an $r^2$ value of 0.91 (Table 1). While there is some spread in the simulated versus observed surface SWD fluxes at all values, method TR tends to overestimate the observations particularly at higher SWD (Fig. 5, bottom right). Method TR simulates low transmission ratios well (Fig. 6, bottom right), with spread increasing and tending toward oversimulation as the transmission ratios increase. Method TR has the least spread and the best constrained simulated versus observed transmission ratios. While it is clear that

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**Fig. 5.** Comparison of surface downward SW radiation observed by the ship-based pyranometer and simulated by each of the four methods: Z, ZL, TL, and TR. Best-fit line (black) and the 1:1 line (gray) are also plotted.
method TR does not produce a perfect simulation of the surface SWD fluxes, its physics and vertical structure are sound and this method performs well with regard to reproducing SW observations.

To probe the differences between these methods, sensitivity tests have been run to determine the model’s response to changes in cloud thickness and vertical location. While tests show that simulated surface SW radiation is insensitive to vertical translations of atmospheric liquid water in the marine boundary layer, it is sensitive to the distribution of water within a cloud layer (i.e., cloud thickness). This sensitivity is the modeled cloud albedo effect to droplet number (Twomey 1974); that is, when the same volume of water is distributed across a greater number of cloud droplets, the scattering cross section increases. Following this reasoning, when a method underestimates surface SW fluxes, it is likely caused by the cloud water being distributed over too many droplets and thus having a greater cross-sectional area and extinguishing more solar radiation by scattering. In the simulations presented here, since the number concentration is prescribed, a larger number of smaller droplet effective radii are used than would have been observed, thereby resulting in a simulated cloud with higher reflectivity than observed. Conversely, times when simulated surface downward SW fluxes are

![Fig. 6. As in Fig. 5, but for γ. The 1:1 line is plotted (light gray line) across the major diagonal.](image)

<table>
<thead>
<tr>
<th>Metric</th>
<th>Z</th>
<th>ZL</th>
<th>TL</th>
<th>TR</th>
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<tr>
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<td>112</td>
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<tr>
<td>r²</td>
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</tr>
<tr>
<td>Mean offset (W m⁻²)</td>
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<td>7</td>
<td>65</td>
<td>47</td>
</tr>
</tbody>
</table>

Table 1. Validation statistics for each of the four methods investigated. Statistics are calculated between simulated surface downwelling SW radiative fluxes and surface downwelling SW radiative fluxes observed by a pyranometer aboard the RHB.
greater than that observed occur generally when thin clouds are present and the effective radius of the cloud droplets may be overestimated, thereby scattering less and transmitting more SW radiation to the surface. Method TR is consistent both optically and theoretically structurally (i.e., a triangular LWC profile, an increasing \( r_e \) with height, a uniform \( N_d \) with height) with well-mixed marine stratocumulus clouds. Therefore, method TR is assumed to simulate the most accurate atmospheric profiles of radiative fluxes, and these resultant simulated fields will be discussed throughout the rest of this paper.

To examine the uncertainty in simulated fluxes as related to observations utilized within method TR, simulated transmission ratios resulting from a range of retrieved LWP and calculated uniform effective radii \((r_e,\text{uniform})\) are shown in Fig. 7. Median meteorological conditions, cloud depth, and cloud height were input parameters for the RRTM to simulate the transmission ratios. The uncertainty in 10-min LWP observations is 35 g m\(^{-2}\) (Zuidema et al. 2005), which is used to calculate the uncertainty in surface downwelling shortwave radiation via the transmission ratio. At low LWP (<100 g m\(^{-2}\)) the transmission ratio measurement uncertainty is ±0.07, while in the middle range of LWP (150–230 g m\(^{-2}\)) it is ±0.03, and at the high LWP (>300 g m\(^{-2}\)) the uncertainty is ±0.01.

The simulated fields resulting from method TR are used to examine the average diurnal cycle of radiation (section 4a) and profiles of the radiative flux components [section 4b(1)], which are useful for exploring such topics as radiative divergence through the cloud layer. Radiative divergence through the cloud layer has been applied to turbulence and entrainment calculations (e.g., Lilly 1968; Lock and Macvean 1999; Caldwell et al. 2005; Ghate et al. 2015). Additionally, the effect of clouds on radiation in the atmosphere is examined through calculating heating rates and cloud radiative forcing [sections 4b(2) and 4c, respectively].

4. Simulated fields

a. Average diurnal cycles

Adopting method TR, average diurnal cycles of the simulated flux components from the surface to 2000 m are displayed in Fig. 8. SW fluxes (Fig. 8, top row) start at sunrise, reach a maximum at local noon, and then taper until sunset. Upwelling (Fig. 8b) and downwelling (Fig. 8a) SW fluxes decrease with depth as solar radiation penetrates the atmospheric column. The SW fluxes end much more abruptly throughout the entire atmospheric column at sunset than when they begin following sunrise, when they slowly penetrate the atmospheric column. Upward SW fluxes (Fig. 8b) are strongest during the first half of the day (~(0700–1100) LT]. This is due to the greater cloud cover during the morning. The daily cycle of cloudiness can be seen in the plot of reflectivity (Figs. 3a and 4a) and has been noted previously (de Szoèke et al. 2010; Burleyson et al. 2013). While upward SW fluxes do affect net SW (Fig. 8c), upward SW fluxes are only, on average, approximately 3% of the downward SW fluxes at cloud top.

LW radiative fluxes (Fig. 8, middle row) are fairly consistent at all heights throughout the day. Similar to upward SW fluxes, the composite upwelling LW fluxes (Fig. 8e) increase at higher altitudes in local afternoon and evening when cloud cover decreases. Likewise, there is a decrease in downwelling LW radiation during decreased cloud fraction during afternoon and evening (Fig. 8d). The diurnal composite of LW net radiation (Fig. 8f) shows that upward LW fluxes dominate in the upper altitudes (~1500–2000 m). In contrast, the upward and downward components nearly balance each other below cloud as a result of the similarity in surface and cloud-base temperatures. However, the below-cloud LW fluxes do not exactly balance each other and this imbalance is important when considering boundary layer turbulence and mixing.

Total radiative fluxes (Fig. 8, bottom row) are dominated by the shortwave components during daylight hours when the sun is high enough (solar zenith angle < 60°). There is a slight modulation of the contribution by the net SW radiation, which is a strong downward flux, in the net total radiation (Fig. 8i) during the local day. This is due to the contributions by the net LW fluxes being upward, particularly above ~1500 m. The diurnal
cycle of net total radiation shows a clear cycle of upwelling LW radiation during the nighttime hours above clouds (≥1500 m) with nearly zero radiation below the cloud layer. This is followed by a transition to downward SW radiation dominating during daylight hours and strengthening until local noon before tapering off and returning to the upward LW regime. The height of stronger upwelling LW increases between nightfall and daybreak as a result of increased cloudiness and the diurnal cycle of cloud-top height (de Szoek et al. 2012).

The simulated radiative fluxes presented here agree well with longwave radiative flux aircraft observations during VOCALS-REx along 20°S (Bretherton et al. 2010). In particular the upwelling LW in the subcloud layer reported by Bretherton et al. (2010) was consistently −380 W m⁻² across their longitude bins from 70° (nearshore) to 85°W (remote), while the simulated values in this study range from approximately −398 nearshore to −394 W m⁻² at the Bretherton-defined remote bin (80°–86°W; not shown)—a <5% difference.
Simulated downwelling LW is similarly close at the near-shore and remote longitude bins. The largest discrepancies occur in the simulation of upwelling LW above the cloud layer, with the simulated fluxes being approximately 30 W m\(^{-2}\) greater than those reported in Bretherton et al. (2010). However, this is still a difference of approximately 10% and may be attributed to sampling differences.

b. Profiles

1) Radiative flux average profiles

Profiles of radiative fluxes for local noon, midnight, and a daily average are displayed in Fig. 9. Considering the 24-h composite profile (Fig. 9, top), the net SW flux (positive downward) increases slightly from the surface to the cloud base and increases from the loud base to the cloud top before eventually becoming uniform in the free troposphere. The decrease of SW fluxes downward with depth is due to the absorption of SW radiation by water vapor and the scattering by clouds and aerosols. Net LW fluxes are nearly constant in the subcloud layer with a strong decrease through the upper portion of the cloud layer and a continued decrease to the top of the MBL. Unsurprisingly, the 24-h composite profiles of total net and net SW radiation are positive (downward) throughout the MBL. In contrast, the 24-h composite net LW radiation is negative (upward) throughout the MBL.
Similar patterns are observed in the flux profiles for local noon (Fig. 9, middle), although with greater values for SW and, consequently, net fluxes. At midnight, solar radiation is absent from the system, leaving just LW fluxes in the total. The patterns noted in the average profile are amplified for the LW fluxes, with a steeper decrease within the cloud layer in the net flux, coming from a stronger decrease in the downward LW flux between the cloud base and the cloud top. The profiles plotted here are smoother than those presented in previous studies (e.g., Stevens et al. 1999; Ghate et al. 2014) likely as a result of averaging many days of data together. Slight wiggles still exist in the profiles as a result of averaging clouds of differing heights and depths together. The surface radiative flux values presented here compare well with surface values presented in de Szoeke et al. (2012). Averaged simulated surface downwelling LW radiative fluxes calculated in this study have a reported value of 341 W m\(^{-2}\), approximately 9% lower than the 373 W m\(^{-2}\) measured in 2008 in the southeastern Pacific stratus synthesis dataset (de Szoeke et al. 2012). Average surface downwelling shortwave simulated fluxes are 233 W m\(^{-2}\), which is nearly identical to that presented in de Szoeke et al. (2012).

2) Cloud-Top Height Relative Average Radiative Flux and Heating Rate Profiles

The average profiles of the shortwave (red), longwave (blue), and total (black) radiative flux components as simulated by RRTM are shown in Fig. 10 (left), along with the associated heating rates (Fig. 10, right). Profiles have been plotted relative to cloud-top height and are shown from 200 m above cloud top to 500 m below cloud top. Only times when cloud boundaries were identified are included. Average profiles over the entire time series are displayed in the top panels, with average noon profiles displayed in the middle, and average midnight profiles at the bottom.

SW net radiative flux profiles in total and at noon decrease slightly from cloud top to 500 m below cloud top, with a steep decrease in the upper portion of the cloud where the net solar radiation is absorbed. Above cloud top net SW radiation has a nearly constant profile. At cloud top SW net radiative fluxes are approximately 222 and 805 W m\(^{-2}\) for total (daily mean) and noon, respectively. SW heating rates increase sharply at cloud top during daytime, reaching a maximum of approximately 2.9 K h\(^{-1}\).
LW net radiative flux profiles show a slight increase from 500 m below cloud top as the height approaches cloud top until there is a dramatic decrease near cloud top. A slight decrease in LW net radiative flux continues above cloud top. Throughout the entire profile shown, the LW net radiative flux is negative. The slight increase in the LW heating rate (Fig. 10, right, blue lines) is produced by the weak increase in LW radiative fluxes. Considering all times, the maximum LW heating rate magnitude is approximately 6 times the SW heating rate; thus, the net heating rates echo the LW profiles but with slightly diminished magnitudes. In contrast, during local noon, while LW cooling dominates the net heating rate at cloud top and above, SW radiation penetrating the cloud layer dominates the net heating rate profile below, resulting in a positive net heating rate extending from the cloud top down approximately 200 m. The heating rate profiles are similar to previous observational studies (Duda et al. 1991; Ghate et al. 2014).

c. Cloud radiative forcing

Cloud radiative forcing is defined as the difference between the overcast and clear-sky radiative fluxes. Only times when both a cloud top and a cloud base were identified are included in this analysis. The clear-sky fluxes are simulated by simply removing clouds from the input to the RRTM. The simulated surface clear-sky downward SW fluxes are seen as the black sinusoidal curves in Figs. 3 and 4. The average diurnal cycles of cloud radiative forcing components from the surface to 2000 m are displayed in Fig. 11. Cloud radiative forcing attributable to SW fluxes (Figs. 11a–c) results in additional SW radiation above cloud top during daylight hours and decreased SW radiation from the surface to below cloud top. Downward LW cloud radiative forcing (Fig. 11d) results in additional radiation below cloud top at all times (~90 W m\(^{-2}\) in the subcloud layer). Total cloud radiative forcing (Fig. 11i) is dominated during daylight hours by SW cloud radiative forcing, with reduced radiation below cloud top to the surface. The late afternoon and evening hours have a decreased magnitude of cloud radiative forcing compared to earlier in the day as a result of a decrease in cloud thickness as clouds dissipate during these hours. During the night, clouds result in increased radiation in the layer of the MBL below cloud top.

The results presented in de Szoeke et al. (2012), taking into account seven years of cruise data in this region, report a net surface cloud radiative forcing of \(-131\) W m\(^{-2}\). The results considering data just from the second leg of VOCALS-REx presented here report a value of \(-111\) W m\(^{-2}\). The differences presented here are attributed to both different subsets of the data examined and minor differences in clear-sky models employed for each study despite covering roughly the same study location.

5. Summary

This study tested four separate methods of prescribing cloud properties in a radiative transfer model of 10-min simulated radiative fluxes in the tropical southeast Pacific. The trials represent the state of the cloud layer and demonstrate the sensitivity of the model to prescribed microphysical properties. The primary application of this study is to simulate physically consistent radiative profiles throughout the atmosphere, especially within the cloud, where they are difficult to observe precisely. We adopt a method that physically incorporates radar reflectivity and passive microwave cloud remote sensing measurements: cloud depth, retrieved liquid water path, and optical depth. Of the four methods examined, method TR most closely reproduces the structure of observed atmospheric and in-cloud radiative fluxes. The radiative fluxes, both above and below the cloud layer, simulated in this study are comparable (within 5%–10%) to observations presented in previous literature (i.e., Bretherton et al. 2010; de Szoeke et al. 2012).

To investigate the effect of clouds on the marine boundary layer, the model was also run in a clear-sky scenario. Utilizing both sets of simulated radiative fluxes, a diurnal cycle of cloud radiative forcing was presented. Simulated values of net surface cloud radiative forcing are approximately 15% lower than previously published values (de Szoeke et al. 2012) for this study region primarily as a result of using a different data subset.

Fields within this dataset allow for an approach to estimate flux divergence and turbulence with implications for studying entrainment and the redistribution of energy within the marine boundary layer. Flux divergence is a key component in estimating the turbulent heat flux, which is proportional to the largest part of the buoyancy flux in the boundary layer turbulent kinetic energy budget. Understanding the development, maintenance, and destruction of turbulence in the layers near cloud top will lead to further understanding of entrainment at the top of the boundary layer. The ability to estimate cloud-top entrainment rates is key to modeling the marine boundary layer depth and moisture budget. In ongoing research we utilize the simulated radiative fluxes presented here to calculate buoyancy flux that is then used to estimate the turbulent kinetic energy budget and cloud-top entrainment rates for stratocumulus over the southeastern tropical Pacific Ocean.
Applications of this dataset via turbulent mixing schemes extend to commerce via the forecasting of cloud ceilings and surface visibility for the aviation community. The modeling community requires structurally accurate profiles of radiation for forcing large-eddy simulations and initializing numerical weather prediction models. Furthermore, physically consistent simulated radiative fields that are constrained by observations for subgrid-scale clouds, such as those presented here, are important to improving the representation of marine stratus-cumulus clouds and the tropical marine boundary layer in global climate models and future climate projections.

Time series of the input fields, simulated radiative fluxes, cloud radiative forcing, and heating rates can be accessed at the Tropical Eastern Pacific Synthesis page on the Oregon State University website (http://people.oregonstate.edu/~deszoecks/synthesis.html).

FIG. 11. As in Fig. 8, but for radiative flux components of cloud radiative forcing from the surface to 2000 m. Cloud radiative forcing is the effect of clouds on radiation in the atmosphere. (left to right) . Total cloud radiative forcing is shown in (i).
Acknowledgments. This work was supported by NSF Award 1619903. This research would not have been possible without access to the Atmospheric and Environmental Research, Inc., Rapid Radiative Transfer Model. The authors thank Dr. Paquita Zuidema for providing the microwave radiometer dataset and Dr. Virenda Ghate for his invaluable guidance and advice for running the RRTM.

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