Aerosol–Stratocumulus–Radiation Interactions over the Southeast Pacific

GUOXING CHEN AND WEI-CHYUNG WANG
Atmospheric Sciences Research Center, University at Albany, State University of New York, Albany, New York

JEN-PING CHEN
Department of Atmospheric Sciences, National Taiwan University, Taipei, Taiwan

(Manuscript received 28 October 2014, in final form 25 March 2015)

ABSTRACT
Atmosphere–ocean general circulation models tend to underestimate the solar radiative forcing by stratocumulus over the southeast Pacific, contributing to a warm sea surface temperature (SST) bias. The underestimation may be caused by biases in either macro- or micro- (or both) physical properties of clouds. This study used the WRF Model (incorporated with a physics-based two-moment cloud microphysical scheme) together with the 2008 Variability of the American Monsoon Systems Ocean–Cloud–Atmosphere–Land Study (VOCALS) field observations to investigate the effects of anthropogenic aerosols on the stratocumulus properties and their subsequent effects on the surface radiation balance. The effects were studied by comparing two cases: a control case with the anthropogenic aerosols and a sensitivity case without the anthropogenic aerosols. Results show that the control case produced cloud properties comparable with the measurements by aircraft and that aerosol–cloud microphysical interactions play an important role in regulating solar cloud radiative forcing. As expected, the anthropogenic aerosols increase the cloud droplet number and decrease the cloud droplet size, resulting in an enhancement of solar cloud radiative forcing and a reduction in solar radiation reaching the sea surface, up to a maximum of about 30 W m$^{-2}$ near the coast. Results also show that aerosol–cloud microphysics–radiation interactions are sensitive to cloud fraction, thus highlighting the role of cloud diurnal variation in studying the cloud–radiation interactions. Analysis of the high-resolution (3 km) model simulations reveals that there exists an inherent scale dependence of aerosol–cloud–radiation interactions, with coarser horizontal resolution yielding a weaker variability.

1. Introduction
Extensive marine stratocumulus clouds (MSC) lie over the west coasts of subtropical North and South America and Africa. These clouds significantly reflect the incoming shortwave radiation and cause a radiative cooling to the surface. It is known that atmosphere–ocean general circulation models (AOGCMs) simulate larger surface insolation (de Szoek et al. 2010) and warmer SSTs over oceans in the stratocumulus regions (Large and Danabasoglu 2006; Wang et al. 2014), which is consistent with the biases found in the smaller simulated shortwave cloud radiative forcing in these regions (Calisto et al. 2014; Flato et al. 2014).

It is also known that these regions have been affected by aerosols emitted from the continents. For instance, the southeast Pacific (SEP), where the largest and most persistent stratocumulus deck in the world resides, is exposed to anthropogenic aerosols produced by copper smelters in South America (Huneeus et al. 2006). The increased aerosols not only directly reduce the shortwave radiation reaching the surface but also indirectly modify the cloud microphysical properties, resulting in an enhancement in cloud albedo that further reduces the surface insolation. The purpose of the present study is to investigate to what extent the anthropogenic aerosols affect the cloud microphysical properties and their subsequent effects on the surface radiation balance over the SEP within the context of the biases mentioned above. One unique aspect of the present study is that we used the measured aerosol size distribution from the 2008 Variability of the American Monsoon Systems Ocean–Cloud–Atmosphere–Land Study (VOCALS)
field campaign (Wood et al. 2011) as inputs to the WRF Model coupled with a physics-based two-moment cloud microphysical scheme, which responds effectively to the aerosol effects (Cheng et al. 2007, 2010; Hazra et al. 2013a). These are described below.

a. The southeast Pacific climate system

The Andes cordillera in South America, with its high altitude and extensive length, creates a setting favoring the formation and maintenance of stratocumulus over the SEP (Richter and Mechoso 2006). Meanwhile, aerosols (composed of mostly sulfates) in the boundary layer have higher concentrations near the coast than over the remote ocean because of emissions associated with the industrial and urban activities in the South American coastal region (Kleinman et al. 2012). Complex interactions between the meteorology and aerosols are involved in modifying cloud macro and micro properties, one striking feature of which is the significant east–west gradients as shown by the VOCALS campaign. For cloud macro properties, both cloud-top and -base heights increase from the coast to the remote ocean, with weakening coupling between cloud and subcloud layers (Bretherton et al. 2010; Jones et al. 2011), which is associated with gradients in SST, large-scale subsidence above the inversion layer, and other large-scale factors. For cloud micro properties, cloud droplet number concentration (effective radius) is larger (smaller) near the coast than over the remote ocean (Twohy et al. 2013), implying a connection to aerosols originated from the continent. Thus, it is necessary to resolve both the large- and microscale processes to accurately simulate the aerosol–stratocumulus interactions over the SEP region.

b. The physics-based two-moment microphysical scheme

The cloud microphysical scheme used here was documented in Cheng et al. (2007, 2010) and Hazra et al. (2013a). It predicts both number and mass mixing ratios of cloud droplet, rain droplet, cloud ice, snow, and graupel. Aerosols are assumed to be composed of ammonium sulfate with a trimodal lognormal size distribution,

\[
\hat{n}_i(r) = \frac{3}{\sqrt{2\pi}\sigma_i^2} \exp \left( -\frac{(\ln r - \mu_i)^2}{2\sigma_i^2} \right),
\]

where \(\hat{n}_i\), \(\mu_i\), and \(\sigma_i\) represent total number concentration, geometric mean radius, and geometric standard deviation of each mode, respectively. Parameters \(\mu_i\) and \(\sigma_i\) are fixed from the initialization, so the total number concentration directly converts to aerosol mass mixing ratio, which is a prognostic variable in the model. Note that the warm-phase part of this scheme is not derived by assuming specific droplet size distributions as other bulk microphysical schemes; rather, it is based on statistical analysis of results from bin microphysical simulations (Chen and Liu 2004), which makes it more accurate and computationally efficient. The scheme also has three distinct features. First, aerosol mass concentrations in air and hydrometeors are tracked explicitly, which allows aerosols to be removed by precipitation and recycled upon the complete evaporation of cloud droplets. Second, activation of aerosol particles into cloud droplets are calculated in detail according to the Köhler theory with cloud supersaturation resolved using an embedded Lagrangian parcel method [see Cheng et al. (2007) for details], and the initial size of activated droplets is calculated according to the CCN size and updraft speed. Third, the scheme considers the activation of giant CCN directly into rain embryos to allow more realistic drizzle initiation in the clean environment. The mixed-phase part is from the cloud parameterization of Reisner et al. (1998).

The cloud microphysical scheme has been used to simulate midlatitude warm stratus (Cheng et al. 2007), frontal systems (Cheng et al. 2010), and tropical cyclones (Hazra et al. 2013b). Nevertheless, these studies are mainly sensitivity simulations while direct comparisons with observations have not been conducted. The available observations of cloud microphysical parameters in VOCALS thus provide a good reference for such a comparison. Below, section 2 describes the model configurations, major findings are given in section 3, and summary and discussion of the uncertainties are given in section 4.

2. Model configurations

a. WRF simulations

Figure 1a presents the domain setup used in this study. Three two-way nested domains, with resolutions of 27 (d01), 9 (d02), and 3 (d03) km, cover the southeast Pacific. Domain d03 matches the 20 surveys of the VOCALS campaign (Wood et al. 2011). As stratocumulus clouds are usually thin and sensitive to the vertical resolution in treating boundary layer processes (cf. Zhu et al. 2010), two setups of vertical layers are used in this study. In one setup (L), there are 37 vertical levels, where the resolution is about 40 m near the surface, about 110 m in the cloud layer, and stretched to about 4000 m around the model top (30 hPa); in the other setup (H), there are 63 vertical levels, where the resolution keeps at 40–50 m in
the boundary layer (including subcloud and cloud layers) and is stretched similarly above as in the L setup.

The RRTM longwave scheme (Mlawer et al. 1997) and the Dudhia shortwave scheme (Dudhia 1989) are called every 27 min for all domains. The Monin–Obukhov surface-layer scheme (Monin and Obukhov 1954), the thermal diffusion land surface scheme (Dudhia 1996), and the Yonsei University (YSU) boundary layer scheme (Hong et al. 2006) are used over all domains. The Grell–Devenyi ensemble cumulus parameterization (Grell and Dévényi 2002) is applied over d01 and d02, but not in d03. SST is fixed to its initially prescribed values during the simulation period. Results in d03 are stored once per hour.

Aerosols in the model are initialized with specific size and spatial distributions and are replenished from the surface and the lateral boundaries. For the aerosol size distribution, the initial aerosol numbers, the geometric mean radii, and the geometric standard deviations are determined through fitting the mean spectrum of subcloud (below 500 m) aerosols observed by C130-RF04 during the VOCALS field campaign (the fitted spectrum is shown in Fig. 1b, with the modal parameters given in the caption). This flight took place during the local nighttime of 23 October 2008 and relatively uniformly sampled the air from the coast to about 85°W (flight track shown by the gray line in Fig. 1a, with subcloud legs indicated by thick black dashes). The subcloud aerosols can be considered as representing the mean background size distribution. For the aerosol spatial distribution, aerosol mass is assumed to be horizontally homogeneous, vertically well mixed below PBL top.
(850 hPa), and decreasing exponentially with a scale height of 800 m above. Aerosols concentrations are fixed at the lateral boundaries of the outermost domain (d01), where aerosols can flow in and out based on local aerosol concentrations and winds. The aerosol replenishment rate from the surface keeps constant throughout the simulation for each grid, as explained below.

Two cases were simulated: a control case (C for short), which represents the current polluted situation, and a sensitivity case (S for short), which excludes the anthropogenic aerosols. The influence of natural aerosols is assumed to be uniform over the whole region and identical for the two cases. Natural aerosols are represented by the initial aerosol loading combined with the replenishment from lateral boundaries. Both cases were run for 9 days, from 0000:00 UTC 20 October to 0000:00 UTC 29 October 2008, driven by NCEP Final Operational Global Analysis $1^\circ \times 1^\circ$ data (National Centers for Environmental Prediction/National Weather Service/NOAA/U. S. Department of Commerce 2000). During this period, the SEP was disturbed by a midlatitude synoptic system around 22–26 October, which can be seen in the 500-hPa height fields. In case C, aerosols are replenished from the surface with rates based on the estimated monthly mean anthropogenic SO2 emission rates within the d03 region, focusing on the transitions from the coast to the remote ocean. The domain-mean total column aerosol mass (including both interstitial aerosols and aerosols inside cloud/rain droplets) is 8–12 mg m$^{-2}$ in the control case, larger than that in the sensitivity case, which has reached a steady state of about 5 mg m$^{-2}$ after the model spinup (2 days). The main difference between the two cases lies in the boundary layer near the coast. For example, as shown in Fig. 1c, the difference in the dry aerosol mixing ratio below clouds between CL and SL decreases westward: about 4.5 times at the coast and about 3.5 times at 75°W. To the west of 80°W, the difference is ignorable. In the following subsections, we discuss the effects of aerosols on cloud macro and micro properties and the subsequent impact on cloud radiative forcing.

3. Results

We present mainly the simulations and the comparisons with observations of the cloud characteristics within the d03 region, focusing on the transitions from the coast to the remote ocean. The domain-mean total column aerosol mass (including both interstitial aerosols and aerosols inside cloud/rain droplets) is 8–12 mg m$^{-2}$ in the control case, larger than that in the sensitivity case, which has reached a steady state of about 5 mg m$^{-2}$ after the model spinup (2 days). The main difference between the two cases lies in the boundary layer near the coast. For example, as shown in Fig. 1c, the difference in the dry aerosol mixing ratio below clouds between CL and SL decreases westward: about 4.5 times at the coast and about 3.5 times at 75°W. To the west of 80°W, the difference is ignorable. In the following subsections, we discuss the effects of aerosols on cloud macro and micro properties and the subsequent impact on cloud radiative forcing.

a. Aerosol–cloud interactions

Figure 2 presents the simulated longitudinal distributions of cloud macro properties—cloud fraction (defined by liquid water path greater than 1 g m$^{-2}$), gridded liquid water path (LWP; only low clouds), cloud-mean LWP, and cloud-top and cloud-base heights averaged over 22–28 October—together with the observed LWP and cloud-top and cloud-base heights. The model generally captures the observed gradients in four parameters from the coast to the remote ocean, but large biases exist. All runs show decreasing cloud fraction, decreasing cloud LWP, and increasing cloud-top and cloud-base heights toward the remote ocean. Despite the large differences in the aerosol concentration, this model can directly use cloud droplet effective radius $r_e$ as an input, making it helpful in assessing the effects of cloud droplet number or size on cloud radiative properties. In addition, cirrus clouds, occasionally observed near the coast (Bretherton et al. 2010), may have a great impact on the longwave radiation transfer so they were removed in the offline calculation in order to focus on the radiative forcing by stratuscumulus clouds only. The WRF-simulated profiles of water vapor mixing ratio, temperature, cloud liquid water content (only low clouds), and $r_e$ [diagnosed with a formula in Chen and Liu (2004) that considers the dependence of $r_e$ on droplets’ spectral dispersion] of each column in d03 were inputted through the user-defined entry to drive the SBDART model. Both shortwave (0.25–4 μm) and longwave (4–50 μm) radiative transfers were integrated over six wavelengths.

b. Offline evaluation

The Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) model (Ricchiazzi et al. 1998) was used to do the offline calculation of the radiation fields.
between the control and sensitivity cases (especially near the coast), all four parameters show minor differences between the two cases (thick black lines versus thin black lines and thick gray lines versus thin gray lines). This implies that the aerosol concentration has little effect on the macro features of clouds, which are most likely dominated by the large-scale meteorology. This finding is consistent with the results of Cheng et al. (2007) for midlatitude shallow stratus clouds.

The simulated spatial distribution and temporal variation of cloud fraction are comparable to observations but have relatively large biases. The cloud fraction in CL and SL is about 90% near the coast and gradually decreases westward to about 45% over the remote ocean (Fig. 2a). It is much smaller than the low-cloud fraction derived from GOES-10 channel-4 radiance (not shown in the figure; Abel et al. 2010), especially over the remote ocean. This underestimation exists for both daytime and nighttime cloud fractions. Lowering the threshold LWP for cloud definition from 1 to 0.1 g m\(^{-2}\) does not appreciably improve the result. The cloud fraction in CH and SH tends to be slightly smaller (larger) near the coast (over the remote ocean) than that in CL and SL, but it is also smaller than the observed values. Meanwhile, daytime and nighttime cloud fractions show opposite responses to the vertical resolution. As shown in Fig. 3, the cloud fraction in CH and SH is larger during the nighttime (dominating over the remote ocean) but smaller during the daytime (dominating near the coast) than that in CL and SL. This has an impact on the cloud radiative forcing, as discussed below. Besides the apparent diurnal cycle, the cloud fraction decreases around 24 October and recovers after 25 October owing to the disturbance by the midlatitude synoptic system around 22–26 October. The simulated cloud fraction is more consistent with the observed cloud fraction during the disturbed period than during the steady period after that. This is in line with the finding by Rahn and Garreaud (2010) that the model noise dominates when it lacks strong synoptic forcing.

The simulated mean gridded LWP agrees well with the observed value over the remote ocean but shows an overestimation near the coast, where CH and SH have relatively better results than CL and SL (Fig. 2b). However, the LWP averaged over cloudy columns (cloud LWP) for all four runs is much larger than the observed LWP at all longitudes (Fig. 2c). CH and SH tend to have smaller cloud LWP than CL and SL (by around 20 g m\(^{-2}\)) and are closer to observations.
The simulated cloud-top and cloud-base heights generally increase from the coast to the remote ocean, consistent with the observed gradient (Fig. 2d). CL and SL tend to have cloud top and cloud base lower than the observed values, which is a common phenomenon in modeling studies of VOCALS (e.g., Andrejczuk et al. 2012; Rahn and Garreaud 2010). In CH and SH, the underestimation in cloud-top and cloud-base heights is reduced significantly, especially near the coast, where the simulated dynamics is sensitive to resolutions because of the complicated topography (Wang et al. 2011). Overall, despite the difference in the definition of cloud between the simulation and various observation platforms, it is believed that the model underestimates cloud fraction but overestimates the cloud thickness (or LWP). The same biases also exist in the CMIP5 results (Nam et al. 2012) and in the simulation of stratocumulus over the southeast Atlantic with multiscale modeling framework (MMF) by Painemal et al. (2015). Note that there exist sampling biases in both time and space. For instance, the observation covered a longer time period with more meteorological variability than the simulations, whereas the aircraft sampled much less space than the model calculation.

Figure 4 presents the simulated cloud microphysical properties: mean cloud droplet number concentration \( N_c \) and effective radius over 22–28 October. It shows that the control case (CL and CH; thick lines) is able to capture the observed gradient (black dots) of decreasing \( N_c \) and increasing \( r_e \) from the coastal region to the remote ocean and that an increase in aerosols can substantially increase cloud droplet numbers but decrease cloud droplet sizes. In the CL run, near the coast, \( N_c \) is more than 200 mg\(^{-1}\) (or cm\(^{-3}\)), while \( r_e \) is about 8 \( \mu \)m; at around 80°W, \( N_c \) is about 80 mg\(^{-1}\), while \( r_e \) is about 11 \( \mu \)m; to the west of 80°W, both \( N_c \) and \( r_e \) show minor variations. The simulated \( r_e \) is slightly larger than the observed \( r_e \). This is related to the overestimation of cloud LWP. In the CH run, \( r_e \) is reduced somewhat as a result of the reduced cloud LWP. CH has relatively large \( N_c \) near the coast because the higher vertical resolution allows the model to better resolve the peaks in supersaturation around the cloud base, which leads to stronger activation. In contrast, the sensitivity case (SL and SH; thin lines) does not show significant changes in cloud droplet number and size from the coast to the remote ocean. To the east of 80°W, their \( N_c \) (\( r_e \)) is smaller (larger) than that in the control case; to the west of 80°W, both cloud droplet number and size are the same as those in the control case. The gradual decrease of differences in cloud droplet number and size from the coast to the remote ocean is consistent with the gradual decrease of difference in the dry aerosol concentration below cloud base shown in Fig. 1c, which reveals the effects of continent-emitted aerosols on microphysical properties of stratocumulus clouds over the SEP. Meanwhile, it can be concluded that effects of anthropogenic aerosols emitted from the continent in this case are largely limited to the east of 80°W, which confirms the conclusion based on trajectory analysis by Bretherton et al. (2010).

b. Aerosol–cloud–radiation interactions

Figure 5 presents the meridian-mean cloud radiative forcing at the surface by stratocumulus clouds. The shortwave cloud forcing (SWCF) is shown to be enhanced by the increased aerosols and sensitive to the variation in cloud fraction (Fig. 5a). For all cases, the SWCF decreases from the coast to the remote ocean. This is consistent with the westward decrease in cloud fraction. The control case (CL and CH) shows stronger SWCF than the sensitivity case (SL and SH) to the east of 80°W. The difference is the largest near the coast (about 30 W m\(^{-2}\) between CL and SL and about 20 W m\(^{-2}\) between CH and SH), where stratocumulus is more influenced by aerosols from the continent. As the two cases have similar cloud fraction and thickness, the difference between SWCFs in the two cases is a result of the smaller droplet sizes in the control case, which enhances the cloud albedo by more than 10%. Meanwhile, CH and SH tend to have weaker SWCFs than CL and SL and show weaker cloud albedo enhancement because of the smaller daytime cloud fraction. This highlights the importance of diurnal variation of stratocumulus (especially cloud fraction). Daily mean cloud properties that are closer to observations do not ensure more accurate SWCF. The diurnal variation is important to both
the mean value and the diurnal variability of the surface radiation budget. It also highlights the dominance of cloud macro properties (cloud fraction and cloud thickness) in the gross SWCF and the surface shortwave fluxes.

The difference in the longwave radiative forcing (LWCF) between control and sensitivity cases is minor (Fig. 5b), indicating that the aerosol concentration does not affect cloud-base height (temperature) and longwave emissivity much. LWCF generally decreases from the coast to the remote ocean, in line with the westward-increasing cloud-base height (decreasing cloud-base temperature). CH and SH have weaker LWCF than CL and SL because of the higher cloud-base height in CH and SH, which has a lower temperature and emits less longwave downward.

In total, the continent-emitted aerosols significantly reduce radiation fluxes reaching the surface via aerosol–cloud–radiation interactions (shown in Fig. 5c). The increased aerosols induce stronger SWCF by stratocumulus but have little effect on LWCF, and SWCF dominates the total CF (cloud forcing).

In more detail, Fig. 6 presents the cloud albedo and the enhancement due to the increased aerosols for stratocumulus to the east of 75°W (the region indicated by the black dashed rectangle in Fig. 1a), where anthropogenic aerosols influence stratocumulus clouds significantly. The “cloud albedo” here is defined as the ratio of the attenuation of downward shortwave radiation across the cloud layer to the radiation flux around the cloud top, so, in fact, it includes the attenuation by the absorption of the cloud layer (which is usually a minimum compared with the attenuation by reflection) and the influence of the incidence angle.

The cloud albedo enhancement is shown to be dependent on the cloud LWP and the solar diurnal cycle. As expected, the cloud albedo increases nonlinearly with the cloud LWP (vertical depth) and is modulated by cloud droplet sizes. Clouds tend to show less reflectance around noon than at other moments because of the shorter optical path associated with the smaller incidence angle. However, when the cloud is either too thin or too thick, the albedo enhancement caused by aerosols (control minus sensitivity) is minor. This suggests that in both scenarios the cloud LWP (vertical depth) dominates the cloud droplet size and thus the cloud optical depth. Meanwhile, the albedo enhancement is not identical throughout the daytime. It is weaker around dawn and dusk, but the strongest around noon (local noon time is about 1700 UTC). This further highlights the importance of the diurnal variation of stratocumulus. The albedo enhancement is the most significant (more than 10%) over cloudy columns with LWP between 50 and 100 g m⁻² around noon.

c. Scale dependence of stratocumulus radiative forcing

Since AOGCMs use much coarser horizontal resolution than the 3-km resolution used in the present study, we conducted a scale-dependence analysis of the surface radiation fluxes associated with the aerosol–cloud–radiation interactions. The analysis focuses on the shortwave fluxes because they exhibit much stronger variability than the longwave fluxes. To illustrate the scale dependence, we simply used the 3-km-resolution net shortwave fluxes at the surface of each hour to calculate the horizontal-mean values at resolutions of 6, 12, 24, 48, 96, 192, and 384 km, which cover the resolutions of most AOGCMs. With the lowest resolution of 384 km, d03 is divided into four grids extending from the coastal area to the remote ocean. Then the horizontal variability in daily mean and diurnal variation of surface SW fluxes was analyzed.

Figure 7 presents the frequency distribution of the daily mean surface SW fluxes in CL at different horizontal resolutions (black bar plots). At the finest resolution (3 km), the surface shortwave fluxes are far from a normal distribution but rather have a long tail over the small values, meaning that stratocumulus clouds are skewed toward large optical depths. When the resolution becomes coarser, the long tail is shortened owing to
the removal of optically thick clouds through averaging, and the skewness is reduced. Consequently, the horizontal variability (measured by standard deviation; thick black line) is decreased from 26.5 W m$^{-2}$ (3 km) to 21.2 W m$^{-2}$ (384 km), although the domain-mean SW fluxes remain identical for different resolutions. Meanwhile, the horizontal variability in the diurnal strength of surface SW fluxes (figure not shown; similar to that in the daily mean) also decreases with the coarser resolution. This implies that even if AOGCMs can accurately predict the mean surface radiation fluxes, then there is still an issue related to the underestimation in the horizontal variability from about 10% to 20%.

Figure 7 also presents the standard deviation of the daily mean surface shortwave fluxes in SL at different resolutions (thin black line). It shows a similar dependence on resolution as in CL but is much smaller because SL has much fewer optically thick cloudy columns than CL. Thus, the scale dependence of aerosol–cloud–radiation interaction is inherently present, and further study of its characteristics in other types of clouds is warranted.

4. Summary and discussions

This study used the WRF Model together with the VOCALS observations to examine aerosol–stratocumulus–radiation interactions over the southeast Pacific, where AOGCMs generally underestimate the solar cloud radiative forcing. Three points are worth highlighting. First, we demonstrated that the aerosol–cloud microphysical interactions alone can enhance the cloud albedo to a level that the solar radiation reaching the surface can be decreased by as much as 30 W m$^{-2}$ near the coast, thus potentially lessening the warm SST biases.

FIG. 6. Cloud albedo and its enhancement (control minus sensitivity) caused by the increased aerosols for stratocumuli to the east of 75$^\circ$W, as a function of cloud LWP and UTC time, averaged over 22–28 Oct. The cloud LWP bin edge starts from 10 g m$^{-2}$ and increases at the power of 1.3.
simulated by AOGCMs. Given the anticipation that decreasing SST tends to increase both the lower-troposphere stability and the cloud fraction, the anthropogenic aerosols could induce (positive) dynamic feedbacks. Further research is warranted to examine air–sea coupling aiming at quantifying the effects on SST. Second, the strength of the aerosol–cloud–radiation interactions depends on the diurnal variation of cloud fraction, which is found to be sensitive to the vertical resolution in the boundary layer; for example, higher resolution simulates smaller daytime fraction and larger nighttime fraction, thus yielding a weaker shortwave cloud forcing. Third, the scale-dependence analysis implies that the coarse resolution used in AOGCMs underestimates the daily variability of aerosol–cloud–radiation interactions due to the inadequate inclusion of the subgrid-scale cloud properties and their effects on solar radiation.

We have also conducted additional simulations for both the control and sensitivity cases to study the uncertainties associated with the initial meteorological condition and the aerosol size distribution. The sensitivity to the meteorology was tested in two ways: two more days were added for model spinup, and the ERA-Interim 0.7° × 0.7° data (European Centre for Medium-Range Weather Forecasts 2009) were used. The sensitivity to the aerosol size distribution was tested with another subcloud aerosol spectrum measured on 6 November 2008 (RF10) instead of that on 23 October 2008 (RF04). Nevertheless, all contrasting cases show similar findings reported above; that is, the anthropogenic aerosols cause significant enhancement in the shortwave cloud radiative forcing.

Finally, because of the significance of aerosol–cloud microphysics–radiation interactions, continued studies are warranted in several topic areas: more realistic representations of temporal and spatial aerosol distributions; improvement in simulating cloud macrophysical properties (e.g., cloud fraction and cloud-top and cloud-base heights), which have been recognized as major model biases in both regional (Painemal et al. 2015) and global models (Nam et al. 2012); and AOGCM simulations, which allow more realistic air–sea interactions to study the SST responses.

Acknowledgments. The authors acknowledge NCAR/EOL under sponsorship of the National Science Foundation (http://data.eol.ucar.edu/) for providing data of CI30-measured aerosol size distributions and gridded LWP during the VOCALS campaign, ECCAD for providing the MACCity dataset, and Dr. Steven Abel for supplying cloud fraction retrieved from GOES-10 channel-4 radiances. The authors thank the editor and three anonymous reviewers for comments on improving this manuscript. The research is supported by a grant from the Office of Science (BER), U.S. DOE. WCW acknowledges the support from Key National Basic Research Program on Global Change (2013CB955803) to facilitate the visits to Peking University discussing with Professor Huiwen Xue.

REFERENCES


European Centre for Medium-Range Weather Forecasts, 2009: ERA-Interim project. CISL Research Data Archive, National Center for Atmospheric Research, Computational and Information Systems Laboratory, accessed 1 January 2015. [Available online at http://rda.ucar.edu/datasets/ds627.0/]


