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Impacts of vertical distributions of absorbing aerosols and clouds on the direct radiative forcing and radiative heating rates

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Abstract

Using a delta-four stream radiative transfer model (RTM) we assess the impacts of vertical distributions of cloud-embedded exponential aerosol layers (EAL) and single aerosol layer (SAL) above clouds on the direct radiative forcing and vertical heating rates. In the cloud-embedded aerosol layer scene, the input height in the RTM represents the height at which the aerosol loading exponentially reduces to $1/e$ (~36.8%) of that at the surface layer. Changes of the mid-visible aerosol optical thickness (AOT) and cloud optical thickness (COT) on the TOA upward and surface downward radiative fluxes are examined. The critical COT occurs near COT=4, where TOA fluxes remain almost constant with increasing AOT. Aerosols induce scattering effects below the critical COT and absorbing effects above the critical COT. When clouds are fixed at 0.9km in an EAL for solar zenith angle of 35°, COT=10.0, cloud effective radius of 10µm and AOT=1.0, the TOA radiative forcing for aerosols increases from ~11Wm$^{-2}$ to ~60Wm$^{-2}$ when the height of the $1/e$ aerosol loading increases from 0.9km to 5.8km. For the downward radiative forcing, surface radiative fluxes increase from ~28Wm$^{-2}$ to ~86Wm$^{-2}$. For cloud-embedded EAL distributions, vertical heating rates increase by ~0.5Kday$^{-1}$ with a decrease of single scattering albedo (SSA) by 5% at the cloud level. For SAL located at 5km above clouds, vertical heating rates increase from ~23.8Kday$^{-1}$ to ~27.4Kday$^{-1}$ with a decrease of SSA by 5% at the aerosol level. These findings provide further insight into the effect of absorbing aerosols and clouds and how they impact radiative fluxes through the atmospheric column.
1. Introduction and background

The representation of aerosols, clouds, and aerosol-cloud radiative forcing remains highly uncertain, restricting the reconstruction of past climate and the prediction of future climate change (Kahn, 2012). The radiative effect by anthropogenic aerosols due to scattering and absorption of shortwave (SW) radiation is known as the aerosol direct radiative forcing (Satheesh et al., 1999). The term “forcing” emphasizes perturbations stemming from anthropogenic sources. It is defined as the difference between radiative fluxes in the absence and the presence of anthropogenic aerosols (Loeb and Manalo-Smith, 2005). The aerosol direct forcing in cloud-free regions generally leads to cooling (negative forcing) at the top-of-atmosphere (TOA). This forcing reduces surface temperatures since aerosols scatter more solar radiation than that of the ocean surface. The exception occurs over bright surfaces such as deserts (e.g. Patadia et al., 2009) and snow (e.g. Nair et al., 2013). Absorbing aerosols above clouds (AAC) also create warming at TOA since aerosols absorb reflected solar radiation by clouds (Chand et al., 2009; Feng and Christopher, 2015; Jethva et al., 2013; Keil and Haywood, 2003; Meyer et al., 2013, 2015; Peers et al., 2015; Zhang et al., 2014). Thus, the presence of AAC leads to an underestimation of cloud optical thickness (COT) retrieval compared to retrieval of that in a pristine cloud scene. For southern Africa biomass burning aerosols, the COT (at 0.66μm) of 8–12 is the critical value range below (above) which aerosol scattering (absorption) dominates TOA fluxes based on satellite observations (Feng and Christopher, 2015).

Using the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) Aerosol and Cloud Layer Products and the Moderate Resolution Imaging SpectroRadiometer (MODIS) cloud product (MYD06), Meyer et al. (2013) found that the regional mean direct radiative forcing efficiency in an AAC scene increases from 50.9Wm⁻² per unit aerosol optical thickness (AOT⁻¹)
to 65.1 Wm$^{-2}$ AOT$^{-1}$ after correcting for the low COT bias in the standard MODIS COT product at pixel-level computation. Although pixel-level analysis provides detailed radiative forcing magnitudes, it requires high computational time. Subsequently, Zhang et al. (2014) implemented an efficient method to calculate AAC radiative forcing by using the joint histogram of cloud properties (i.e., cloud optical depth and cloud top pressure) in the MODIS Level 3 cloud product and precomputed LUT during ADE calculations. They noted that the change in COT plays a greater role in the direct radiative forcing than the above-cloud AOTs. While AAC generally lead to positive radiative forcing at TOA, forcing magnitudes vary significantly on daily, seasonal, and interannual scales according to measurements from Scanning Imaging Absorption Spectrometer For Atmospheric Chartography (SCIAMACHY) (de Graaf et al., 2014).

Jethva et al. (2013) implemented a technique to simultaneously retrieve AOT and COT using the “color ratio technique” from MODIS. Using the 470nm and the 860nm MODIS channels, they found that the critical COT is between 2 and 5. They also performed sensitivity analyses on AOT and COT retrieval uncertainty associated with single scattering albedo (SSA) and aerosol height biases. They found that an underestimation of SSA leads to an underestimation of both AOT and COT while an underestimation of aerosol height leads to an overestimation of optical thicknesses of both features. The greatest overestimation occurs for AOT=2, which is the thickest AOT examined in their study. Chang and Christopher (submitted to IEEE, 2015) adopted the color ratio technique and textural statistics on the Spinning Enhanced Visible and Infrared Imager (SEVIRI) to detect AAC to extend the detection to absorbing aerosols above closed-cell stratocumulus clouds that can be used to study the daytime variation of AAC optical properties.
Using multi-angular and polarized radiances from the POLarization and Directionality of the Earth’s Reflectances (POLDER) instrument, Peers et al. (2015) implemented a technique to simultaneously retrieve aerosol and cloud properties for AAC scenarios, including AOT, COT, Angstrom exponent, and SSA. Over the southeast Atlantic, the mean instantaneous direct radiative effect during August 2006 is 33.5 Wm$^{-2}$. The maximum instantaneous direct radiative effect of $\sim$125 Wm$^{-2}$ prevails at $8^\circ$S near coastal regions. Using the Clouds and the Earth’s Radiant Energy System (CERES) instrument and a RTM, Oh et al. (2013) found that the global all-sky mean direct radiative effect increases with COT for COT>2 but are weakened by both increasing surface albedo and cloud effective radii. Aerosols with a relatively lower SSA and asymmetry parameter (g) tend to induce stronger radiative effects.

Using spaceborne lidar observations Devasthale and Thomas (2011) found that, globally, the highest frequency of smoke above clouds occurs during months of June – August. They also found that the seasonal variability of cloud and aerosol separation are the greatest in the $0^\circ$–$30^\circ$S latitude. The largest vertical separation occurs within this latitudinal belt during September – November. Major contributions of aerosol in this region include biomass burning aerosols advected from central Africa to the southeast Atlantic and the advection of biomass burning aerosols from South America to the eastern Pacific Ocean. Alfaro-Contreras et al. (2015) also conducted AAC frequency on a global scale using measurements from both CALIOP and Ozone Monitoring Instrument (OMI). They also found a high AAC frequency over the southeast Atlantic from both sensors.

Min and Zhang (2014) examined the influence of cloud fraction on the diurnal cycle of the direct radiative forcing using the rapid radiative transfer model, shortwave (RRTM_SW) model (Clough et al., 2005; Iacono et al., 2000). They found that using instantaneous cloud fraction
from MODIS as a representation of daily mean cloud fraction led to an underestimation of cloud
fraction and direct radiative forcing relative to a sinusoidal model. The sinusoidal model
included diurnal cycle of cloud fraction from SEVIRI (15-min repeat cycle) and both MODIS
instruments, providing a more realistic perspective of diurnal cloud fractions. Using the daily
mean cloud fraction from Terra, in contrast, led to an overestimation of these two variables.

Differences in cloud fractions (for low-level clouds) occurred because the peak cloud fraction
typically took place at around 07:00 am local time and dropped to a minimum at around 05:00
pm. Thus, using a constant cloud fraction to assess its impact on the direct radiative forcing
would only include uncertainty associated the change in solar zenith angle ($\theta_0$). Interestingly,
they found that the maximum direct radiative forcing took place at around 09:00 am when both
the $\theta_0$ and cloud fractions are considered. Peak direct radiative forcing would intuitively take
place at local noon if merely $\theta_0$ is considered since the highest SW TOA flux occurs during
which the solar incident angle is the highest relative to Earth’s surface. Such findings
demonstrate the complexity of direct radiative forcing calculations. Moreover, they found that
using grid mean COT to assess direct radiative forcing in lieu of a full histogram technique
developed by Zhang et al. (2014) produced high biases since the former assumes plane-parallel
cloud albedo below the aerosol. They also noted that an increase (decrease) in SSA led to a
decrease (increase) in direct radiative forcing, confirming the importance of accounting for
variations in aerosol properties when assessing radiative forcing of AAC.

During the Southern African Fire-Atmosphere Research Initiative (SAFARI 2000), biomass
burning aerosols over the south Atlantic are found to be located at altitudes between 1.8–3.7km,
whereas clouds are located below 1km (Keil and Haywood, 2003). However, Costantiono and
Bréon (2013) noted that it is not uncommon that absorbing aerosols and clouds coexist in a
similar altitude in the southeast Atlantic according to CALIOP observations. The frequency of this scenario increases westward away from the fire burning source. CALIOP can misidentify aerosols below liquid clouds since clouds lead to a complete attenuation of the backscattered radiation to the CALIOP sensor, resulting in an undetected aerosols below liquid clouds (Winker et al., 2009; Young and Vaughan, 2009). Liao and Seinfeld (1998) noted that the direct radiative forcing of cloud-embedded aerosol layer cannot be neglected since the radiative forcing values can be a factor of 3 greater than cloud-free cases.

Uncertainties of AI retrieval above clouds depend on the aerosol layer height. Using the Vector Linearized Discrete Ordinate Radiative Transfer code (VLIDORT), Torres et al. (2012) found that AI changes by ~0.56km\(^{-1}\) and ~0.51km\(^{-1}\) above cloud altitude of 1.5km with COT of 10 and 20, respectively. The greatest dependence of AI retrieval on the aerosol layer height occurs over clear-sky conditions. A 2.5km shift of the aerosol layer above COT of 30 leads to ~25% change in the AI, but the change is over 60% over clear sky.

Johnson et al. (2004) simulated the direct and the semi-direct radiative forcing of absorbing aerosols within and above clouds using the MET Office large eddy model. They found that absorbing aerosols above clouds enhances the buoyancy of free-tropospheric air, thereby deteriorating the entrainment rate. Hence, the boundary layer becomes shallower and more humid, increasing the cloud liquid water path. Moreover, the presence of above-cloud absorbing aerosols causes a reduction of downwelling flux reaching the cloud top, which is another contributing factor that enhances the cloud liquid water path. Both factors give rise to negative semi-direct forcing. In contrast, the cloud liquid water path subsides when aerosols reside in the boundary layer because of elevated solar heating rate in the cloud layer. Aerosols aid to enhance
the existing stability in the boundary layer during daytime, reducing the moisture flux from the
surface to the cloud layer.

Several studies have investigated the impact of the direct radiative forcing of cloud-
embedded aerosols and aerosols above clouds. Previous studies have also examined heating rates
of dust aerosols (e.g., Naeger et al., 2013; Quijano et al., 2000) and aerosols in the Southern
Great Plains of the United States (e.g., Guan et al., 2010). Su et al. (2008) have examined the
impact of dust aerosols on the cloud radiative forcing. However, a comprehensive study on the
dependence of direct radiative forcing and SW heating rates as a function of the vertical
distributions of absorbing aerosols and clouds have not been documented. Acquiring knowledge
of these relationships would advance the understanding of the uncertainty of absorbing aerosol-
cloud direct radiative forcing and heating rates and their implications of the tropospheric stability.

2. Data and Methodology

2.1. Radiative transfer calculations

To assess the sensitivity of radiative fluxes and forcing on AAC, we use a delta-four stream
plane-parallel broadband RTM (Charlock and Alberta, 1996; Fu and Liou, 1993). The model
initially calculates fluxes only in clear-sky and cloudy conditions. Subsequently, it has provided
the capability for calculating radiative fluxes for aerosols (Christopher et al., 2002, 2000). In the
SW spectrum, an uncertainty of up to 5% is documented compared to adding/doubling
calculations (Liou et al., 1988). Model calculations also account for gaseous absorption, water
vapor absorption, and Rayleigh scattering. This model calculates SW fluxes between 0.2–4.0µm
and LW fluxes between 2200 to 0 cm⁻¹. The SW spectrum partitions into six bands (i.e., 0.2–0.7,
0.7–1.3, 1.3–1.9, 1.9–2.5, 2.5–3.5, and 3.5–4.0µm) with ten finer sub-bands between 0.2–0.7µm.
We mainly focus on the SW bands in this study since biomass burning smoke aerosol particles are primarily sensitive to the SW radiation (Haywood et al., 2003). The atmospheric conditions for the entire study are based on the default tropical atmospheric conditions of water vapor, temperature, and other atmospheric constituents (McClatchey et al., 1972).

The southeast Atlantic is dominated by high biomass burning aerosol loadings and semi-permanent stratocumulus clouds during austral winter (June – September) (Yu et al., 2012). This region, therefore, serves as a natural laboratory for studying above-cloud aerosol radiative forcing (Alfaro-Contreras et al., 2015; Devasthale and Thomas, 2011; Wilcox, 2012; Yu et al., 2012). Aerosol properties such as SSA and the asymmetry parameter that will be used for the RTM calculations are obtained from Level 2.0 AErosol Robotic NETwork (AERONET) data (Holben et al., 1998). One of the most frequently available AERONET data near the south Atlantic is the Mongu, Zambia station (Figure 1). Aerosol properties from Mongu provide measurements at 0.441, 0.674, 0.870, and 1.020µm, so the interpolation and extrapolation have been applied to estimate values of SSA and g at particular wavelengths in the RTM (Figure 2). It can be seen that both SSA and g decrease with wavelength, consistent with SAFARI 2000 data (Bergstrom et al., 2007). However, SSA of aerosols increases as they are transported over the south Atlantic as a result of organic carbon condensation (Abel et al., 2003). The differences in wavelength-dependent SSA of aerosols between the south Atlantic and the Mongu station undoubtedly induce uncertainties in assessing the radiative forcing and heating rate of aerosols and clouds. All AOT values throughout this study are reported at 0.55µm. The Aerosol above cloud radiative forcing (AACRF) is obtained by:

\[
\text{AACRF} = F_{\text{cloud}} - F_{\text{cloud+aerosol}}
\]

Eq. 1
where $F_{\text{cloud}}$ is the average SW fluxes at TOA for pristine clouds and $F_{\text{cloud+aerosol}}$ is the SW fluxes at TOA for simultaneous presence of aerosols and clouds. The input values for cosine of $\theta_0$ and the viewing zenith angle, $\theta$, in the model are 0.82 and 0.9, respectively. Based on these geometries for a clear sky condition, the downward SW flux at TOA is $\sim 1122 \text{ W m}^{-2}$. The surface downward flux and the TOA upward flux are $\sim 852 \text{ W m}^{-2}$ and $\sim 66 \text{ W m}^{-2}$, respectively.

2.2. Instantaneous SW heating rates

The vertical profile of the instantaneous radiative heating rate provides information on the magnitude of warming in a vertical column. The equation for the instantaneous SW heating rates (e.g., Guan et al., 2010; Quijano et al., 2000) is given by:

$$\frac{\partial T}{\partial t} = \frac{a \Delta F}{c \Delta p}$$

Eq. 2

where $a$ is acceleration due to gravity, $c$ the specific heat capacity of dry air at constant pressure, $\Delta P$ the atmospheric pressure differences between the top and the bottom of each atmospheric layer, and $\Delta F$ the net flux divergence between two pressure levels. The unit for heating rate is expressed in Kelvin per day (K day$^{-1}$). For brevity, we will refer the instantaneous SW heating rate as “heating rate” for the remainder of this study. We mainly focus on the tropospheric heating rates up to 9 km above mean sea level.

Vertical distributions of aerosols can significantly impact vertical heating rates, since they warm the atmosphere at which they reside by absorbing solar radiation (Guan et al., 2010; Liao and Seinfeld, 1998). In the southeast Atlantic, aerosol vertical distributions can vary in both cloudy and cloud-free conditions. Figure 3 shows vertical distributions of aerosols in two
different scenarios from the CALIOP vertical feature mask feature layers and the aerosol subtype layers. The first scenario pertains to clouds embedded within aerosol layers on 19 August 2009 (Figure 3a). Between 6.3°S to 6.5°S, aerosols vertically extend from the surface to ~4km. The aerosol subtype layer information in Figure 3b indicates that smoke aerosols dominate this domain. Between 5.8°S to 6.2°S, clouds occur at ~1.2km above the surface with attenuation underneath this altitude. Whether aerosols are present below the cloud layer are unknown due to cloud attenuation. Since a narrow smoke layer is present from the surface up to ~0.8km at 5.7°S, one should be skeptical that aerosols may be present below the cloud layer that is bounded by the adjacent smoke layers. The second scenario highlights an aerosol above cloud case from 13 August 2006 (Figure 3c). Aerosols from 19°S to 20°S conspicuously reside above clouds, but various aerosol types are present between these two latitudes as shown in the aerosol subtype layer (Figure 3d).

Given the complexity of the vertical distributions of aerosols, we examine heating rates for both clouds within aerosols and aerosols above clouds. In the first scenario, we simulate heating rates for aerosol loadings that decrease exponentially from the lower troposphere. The equation for layer aerosol loading, \( L \), is given by:

\[
L = \exp\left(-\frac{z}{H}\right)
\]

Eq. 3

where \( H \) is the input scale height in the RTM and \( z \) is the height of focus given by:

\[
z = 8\ln\left(\frac{1013.25}{P}\right)
\]

Eq. 4

where \( P \) is the pressure level of focus. The input height in the RTM represents the scale height at which the stepwise (or layer-by-layer) aerosol loading exponentially reduces to approximately
1/e (~36.8%) of the assigned aerosol loading (Figure 4). For a scale height of 1km, the aerosol loading is ~0.40 of the assigned AOT in the RTM from 0−0.9km. The aerosol loading reduces to ~0.16 of the assigned AOT from 0.9−1.8km. For a scale height of 4km, the aerosol loading is ~0.39 of the assigned AOT from 2.8−3.8km. The brown dashed line in Figure 4 joins the 1/e aerosol loading of each of the five scale heights to be used to assess the heating rates and radiative fluxes in this study. For brevity, we will refer to this mode of layer constraint as the “exponential aerosol layer (EAL)” distribution. The magnitude of columnar aerosol loading increases with the specified scale height since the aerosol loading at each modelled layer increases with the specified scale height. Note that the assigned scale height in the RTM would not necessarily yield exactly 1/e (i.e., 0.368) of the assigned aerosol loading at the layer in which the scale height was assigned since the loading calculations are stepwise rather than continuous. It is also important to note that the actual AOT in the bottom layer (i.e., 0−0.9km) of the EAL begins less than 1. For instance, the aerosol loading in the bottom layer for an assigned scale height of 1km is ~0.40 of the assigned AOT. For an assigned scale height of 4km, the bottom layer aerosol loading is ~0.80 of the assigned AOT.

In the second scenario, we simulate heating rates by constraining all aerosol loadings in a single layer (SAL) for both aerosols in cloud-free and above cloud cases. Note that this study merely focuses on the direct radiative forcing of aerosols and clouds and that both the semi-direct and the indirect radiative forcing (Costantino and Bréon, 2013; Twomey, 1974, 1977) are not considered when aerosols and clouds co-exist in the same altitude in the RTM.

3. Results

3.1. Radiative forcing as a function of AOT and COT
Figure 5a shows the TOA upward SW flux as a function of both AOT and COT for a cloud-embedded EAL where the $1/e$ of the assigned aerosol loading is located at ~4.8km above the surface. The influence of aerosol particles on TOA upward SW fluxes depends on the magnitude of AOT and COT. For COT=2, TOA fluxes increase from ~163Wm$^{-2}$ to ~211Wm$^{-2}$ when the overlying AOT increases from 0 to 2. However, for COT=12, the upward fluxes decrease from ~450Wm$^{-2}$ to ~305Wm$^{-2}$ for the same increase in AOT. These results suggest that the magnitude of COT affect the scattering effect of the incoming and outgoing solar radiation. Aerosols enhance scattering effects of incoming solar radiation with embedded thin clouds, but they scatter less incoming solar radiation than embedded thick clouds. Since the SSA of absorbing aerosols is less than one, these aerosols would absorb the upward fluxes that have been reflected by underlying clouds to reduce the TOA fluxes rather than enhancing the TOA fluxes as in the case for thin clouds. In contrast, thin clouds (COT=2) only reflect a lower fraction of solar radiation compared to thick clouds. The reflected flux by thin clouds is so weak to the magnitude that aerosols primarily scatter and absorb little or none of the reflected radiation from underlying thin clouds. In this case, aerosols actually enhances scattering of the solar radiation, resulting in a negative radiative forcing at TOA. For COT=4, aerosols play a negligible role in perturbing the TOA fluxes. Above (Below) this COT, an increase in AOT results in a decrease (increase) in the TOA fluxes. The “critical COT” is the COT where aerosols transition from largely scattering to absorbing modes. In this case, the critical COT lies at ~4. Similarly, the critical COT for which aerosols reduce or increase the TOA fluxes can be observed for dust above Saharan Desert. Dust aerosols are also found to enhance (reduce) TOA fluxes above dark (bright) surfaces (Patadia et al., 2009) in the same analogy as clouds in EAL. These findings suggest that the sign (positive or negative) of aerosol radiative forcing depends on the magnitude of the background albedo, where
aerosol scattering exceeds surface scattering to enhance TOA fluxes. Previous studies have also

demonstrated the dependence of surface albedo on the direct radiative forcing from satellite
measurements and RTM (e.g., Feng and Christopher, 2015; Quijano et al., 2000).

In clear-sky conditions, downward fluxes reduce from TOA at 1122 Wm$^{-2}$ to the surface at
852 Wm$^{-2}$ due to molecular scattering and absorption. Using a radiative transfer model, Guan
et al. (2010) noted a decrease in the downwelling flux of $\sim$300 Wm$^{-2}$ from TOA to the surface
over the Southern Great Plains during the May 2003 Atmospheric Radiation Measurement
Aerosol Intensive Observation Period (AIOP). An additional aerosol layer in the atmosphere
results in a lower surface downward flux due to scattering and absorption of aerosols. The
differences in the surface downward flux between this study and other studies is expected since
due to different input parameters, types of assumptions, and the nature of model calculations. A
decrease in the downward surface fluxes with an increase in AOT has been documented by
Russell et al. (1999) using sunphotometer measurements on an aircraft in the Atlantic Ocean
near northeast US during the Tropospheric Aerosol Radiative Forcing Observational Experiment
(TARFOX). The reduction of surface downward fluxes with an increased COT has been
documented by Su et al. (2008) who found that surface downward fluxes with an increased COD.

When clouds are present in an EAL for surface downward fluxes (Figure 5b), there is a
decrease in surface fluxes with both increasing COT and AOT. However, the higher the COT,
the smaller the decrease of downward flux with increasing AOT. As an example, for COT=2, the
downward flux decreases from $\sim$731 Wm$^{-2}$ to $\sim$398 Wm$^{-2}$ as AOT increases from 0 to 2. However,
for COT=20, downward fluxes only decreases from 277 Wm$^{-2}$ to 155 Wm$^{-2}$ for the same increase
of AOT. This trend occurs since the reflection of incoming (downward) solar radiation increases
with COT, so there would be less fluxes reaching the surface. Increasing AOT also enhances
solar reflection, which reduces surface fluxes. Furthermore, as the COT increases, the change in
the surface flux for the same amount of change in AOT becomes less sensitive. During TARFOX,
Russell et al. (1999) found that the increase of AOT results in a decrease of surface downward
fluxes for aerosols over the ocean. Moreover, reducing the SSA results in a greater decrease of
the downward fluxes.

For the SAL distribution, TOA upward fluxes are lower than those in the EAL distribution
for the same COT-AOT pair by 10-20Wm$^{-2}$ since the entire aerosol loading has been positioned
above clouds (not shown). In contrast, the highest magnitude of aerosol loading occurs at the
lower troposphere for EAL, which is below the prescribed location of liquid clouds. Hence, the
remaining aerosol loadings above clouds are lower than those in the single layer distribution. The
download surface fluxes for the SAL distribution is higher by less than 10Wm$^{-2}$ than those in the
EAL distribution. The slightly lower surface downward fluxes in the EAL due to its higher
aerosol loading. While the TOA and downward fluxes provide some meaningful comparison
between the SAL and EAL distributions, the vertical heating rate profile would provide more
details on differences between the two distributions. The remainder of the results in this study
will be based on EAL. We will resume comparisons between the EAL and the SAL cases in
detail in the heating rate section.

Variations of COT induce a greater influence on both TOA and surface fluxes than variations
of AOT, consistent with annual mean direct radiative forcing found in Zhang et al. (2014). These
findings confirm that the important role that surface albedos play in determining the radiative
forcing at both the surface and TOA. The direct radiative effect for dust aerosols over the
Saharan desert exhibits a similar characteristics as that for absorbing aerosols above clouds.
Patadia et al. (2009) showed that mineral dust acts as a scattering agent when residing above a
surface albedo of less than 35% but acts as an absorbing agent above this albedo, which is consistent with the radiative processes outlined in Figure 5.

3.2. Aerosol vertical distributions and radiative forcing

The upward TOA and downward surface fluxes for pristine clouds at 0.9km are ~415Wm$^{-2}$ and ~431Wm$^{-2}$, respectively. The change in the direct radiative forcing as a function of EAL distributions are depicted in Figure 6. An AOT of 1 is used for all six aerosol height distributions. Clouds are fixed 1.8km with COT and cloud effective radius of 10 and 10µm, respectively. It can be seen that the upward radiative forcing at TOA increases with increasing height of 1/e aerosol loading. This increase occurs simply due to the increase of aerosol loading with an increase of scale height as shown in Figure 4. For example, when 1/e aerosol loading occurs near 0.9km, the forcing is ~11Wm$^{-2}$. The forcing value increases up to ~60Wm$^{-2}$ when 1/e aerosol loadings are located near 4.8m above the surface.

The surface downward radiative forcing also increases with an increase in 1/e aerosol scale height. For the downward forcing, fluxes decrease with increasing height. The surface downward forcing increases from ~28Wm$^{-2}$ for 1/e at 1km to ~86Wm$^{-2}$ for 1/e at 4.8km. Therefore, changes in the surface downward forcing exceed the TOA upward forcing with the same change in the aerosol scale height in an EAL distribution. This discrepancy can be explained by the greater change of AOT in the lower troposphere than that in the upper troposphere so that greater solar absorption occurs in the lower troposphere for a higher assigned aerosol scale height. In cloud-free EAL conditions, the TOA fluxes are ~140Wm$^{-2}$ while the surface downward flux is ~633Wm$^{-2}$. Fluxes vary by less than 1Wm$^{-2}$ over the 5km aerosol scale height change for aerosols over cloud-free conditions, indicating that changes in aerosol height in the lower
troposphere does not perturb significant radiative fluxes. Additionally, the surface downward radiative fluxes do not vary significantly with the change of the single aerosol layer height in both cloud-free and above cloud cases. Thus, changes in cloud heights play a greater role in the surface radiative forcing than changes in the aerosol scale heights. These findings are not surprising given that a significant upward and downward fluxes take place at the cloud height, altering both the TOA and surface fluxes (Quijano et al., 2000).

3.3. Cloud height variations and radiative forcing

The relationship between radiative fluxes and cloud heights with a fixed EAL distribution (1/e at 5km) is shown in Figure 7. At the TOA, fluxes increase from ~342 Wm$^{-2}$ to ~434 Wm$^{-2}$ from a cloud height of 0.9 km to that at 4.8 km, suggesting that the reflection of solar radiation increases as cloud height increases. As aforementioned, downward fluxes decrease from TOA to the surface due to molecular scattering and absorption of the atmosphere. Since a higher flux reaches the cloud top of higher clouds, those clouds would reflect a higher flux and subject to less molecular scattering and absorption towards TOA. After removing the aerosol layer, fluxes increase from ~415 Wm$^{-2}$ to ~459 Wm$^{-2}$ over the same range of cloud heights. These differences confirm that the reduction of fluxes is due to the absorption of solar radiation by smoke aerosols at both TOA and the surface. The surface downward flux for clouds at 0.9 km is ~319.5 Wm$^{-2}$ and increases to ~325 Wm$^{-2}$ at 4.8 km. Above this height, the flux increases by another 1 Wm$^{-2}$ before reducing its value. Removing the overlaying aerosol layer leads to an increase of surface fluxes by ~100 Wm$^{-2}$ at each corresponding height level, attributable to the attenuation of SW radiation by aerosols.
The net radiative forcing for the upward and downward fluxes with variation of cloud heights are inferred from Eq. (1). At TOA, the upward radiative forcing is about +73Wm\(^{-2}\) when the cloud is located at 0.9km. The positive forcing reduces to about +25Wm\(^{-2}\) for clouds at 4.8km since the upward flux increases with cloud height in the presence of overlaying aerosols. For the surface downward fluxes, the positive forcing for clouds at 0.9km is about +112.2Wm\(^{-2}\). The forcing drops to about +109.5Wm\(^{-2}\) for clouds located at 1.8km but increases by ~1Wm\(^{-2}\) at the highest cloud layer. Hence, the surface downward radiative forcing responds weakly to the changes in cloud height for the same COT than changes in COT for the same cloud height. An explanation for the insensitive surface downward flux is that the scattering by the cloud bottom exceeds molecular scattering and absorption between the surface and the cloud bottom.

3.4. SW Heating rates

Figure 8 shows the EAL of SW heating rates for clear sky, cloud only, aerosol only, and AAC for \(\theta_0=35^\circ\) and \(\theta=25^\circ\). In the clear-sky scenario, heating rates are about 2Kday\(^{-1}\) above 0.9km in the troposphere due to gaseous absorption, which is consistent with the clear-sky scenario in Liao and Seinfeld (1998). However, the 2Kday\(^{-1}\) heating rates in their work occupies the entire troposphere rather than from 0.9km upward. This discrepancy is likely due to the differences in reference level of calculations. The inclusion of \(1/e\) EAL at 5km increases the heating rate to nearly 4Kday\(^{-1}\) in the entire troposphere as a result of elevated solar absorption by aerosols. In a pristine cloud case, heating rates at the cloud level increases to ~5.7Kday\(^{-1}\) owing to the increased downward and upward fluxes. Hence, the cloud level also experiences the peak heating rate in the entire troposphere, identical to the pristine cloud case in Liao and Seinfeld (1998). Heating rates decrease to ~2.8Kday\(^{-1}\) at 1.8km and steadily decreases upward. When
clouds lie at 0.9km with the 1/e aerosol loading occurring at 0.9km, heating rates increase to
~12.4Kday\(^{-1}\) at the cloud layer. Similarly, introducing soot aerosols in clouds have found to
elevate the heating rate of the troposphere (e.g., Liao and Seinfeld, 1998). At the cloud level,
heating rates decrease to ~6.7Kday\(^{-1}\) for 1/e aerosol layer at 5.8km. This reduction of heating rate
at the cloud level is anticipated since a higher altitude of 1/e means that more aerosols have
attenuated the downwelling fluxes before reaching the cloud top. Furthermore, the higher the 1/e
height, the smaller the drop in heating rates with height. Since the aerosol loading is greater and
more evenly distributed for a higher assigned aerosol scale height, the net heating rate
throughout the troposphere would be more evenly distributed as well. In a less evenly distributed
EAL such as 1/e at 0.9km aerosol scale height, the strongest heating would be expected to take
place at lower layers due to drastic aerosol loading decrease. When both the cloud and the
maximum aerosol loading occurs at the same altitude, the heating rates are significantly
enhanced as shown by the in the purple profile (1/e at 0.9km) in Figure 8.

Heating rates of various cloud heights are presented in Figure 9. SW heating rates for the
change of cloud height in an aerosol-free scene is shown in Figure 9a. In this case, a peak of
heating rate due to high downward and upward fluxes is located at the cloud level regardless of
the height of the cloud (Liao and Seinfeld, 1998; Quijano et al., 2000). Heating rates are
~0.7Kday\(^{-1}\) one layer below the cloud level but increases from ~5.7Kday\(^{-1}\) to ~8.5Kday\(^{-1}\) at the
cloud height when elevating from a cloud height of 0.9km to 4.8km. Since downward fluxes
decrease from TOA to the surface due to molecular attenuation, aerosols that are located at a
higher altitude would receive a higher downward fluxes than those located at a lower altitude.
Above the cloud layer, heating rates decrease rapidly and converge to ~2Kday\(^{-1}\) at the upper
troposphere regardless of the cloud height. When the 1/e EAL is located at 5km (Figure 9b),
heating rates shift to higher magnitudes at all cloud heights. The presence of absorbing aerosols
enhances solar absorption, which increases heating rates at the cloud level. As an example, for
clouds at 4.8km with no smoke, heating rates are ~8.8K/day. However, heating rates increase to
~10.2K/day when smoke 1/e of AOT=1 is located at 5km. For clouds located at 0.9km, heating
rates increase from ~5.6K/day with aerosol-free to ~7K/day after including EAL. A
convergence of heating rates above the cloud layer is also evident in the aerosol case, but it takes
place near ~4K/day rather than ~2K/day. The magnitudes of converged heating rates are
consistent with the cloud-free EAL and clear sky heating rate profiles in Figure 8.

Figure 10 illustrates heating rates for SAL at various aerosol altitudes. In the cloud-free
aerosol scene (Figure 10a), heating rates peaks at the altitude in which aerosols exist. The peak
heating rates tend to increase with the altitude of the aerosol. For a SAL at 1km, the peak heating
rate reaches ~12.7K/day. The SAL at 5km results in a peak heating rate of ~19.7K/day. The
higher peak heating rate for the 5km SAL than that for the 1km SAL is similar to the increased
cloud height where molecular attenuation causes a decrease of downward flux from TOA to
surface. When adding clouds at 0.9km in the model, heating rates also peak at the aerosol layer.
However, the magnitude of the maximum heating rate for each profile increases (Figure 10b)
with respect to cloud-free scenes. For SAL at 1km above clouds, the peak heating rate rises to
~17.9K/day. The SAL at 5km above clouds increases to ~26.3K/day. The elevated heating rates
at the aerosol layer due to the additional underlying cloud occurs because of elevated upward
fluxes from the reflection of solar radiation by cloud top that would otherwise reach the ocean.
The reflection of the cloud top contributes additional upward fluxes to the bottom of the aerosol
layer. Using a two-stream RTM, Quijano et al. (2000) found that solar heating rates increase
with an increase in dust loading. Warming at the TOA also occurs when dust resides above
clouds since clouds reflect solar radiation and heat the dust layer. Hence, the same analogy can be applied to the heating of the smoke layer by underlying clouds in this study.

Elevated heating rates in both EAL and SAL distributions provide further implications on the stabilization of the troposphere and changes in cloud properties. The elevated heating rate throughout most of the troposphere in an EAL distribution suggests that the majority of the ambient temperature experiences a warm anomaly, which stabilizes the troposphere. In an SAL distribution, the elevated heating rate mainly takes place at a specific layer, so that the ambient temperature experiences a drastic temperature increase approaching the aerosol layer either from below or the above the aerosol layer. While this study focuses on the direct radiative forcing of clouds within aerosols and aerosols above clouds, it is imperative to note that cloud liquid water path may experience a decrease for the case of clouds within aerosols due to an enhanced stability in the boundary layer and a reduction of moisture flux from the surface to the cloud layer. Such a decrease can be explained by solar absorption of absorbing aerosols in the vicinity of clouds. For absorbing aerosols above clouds, elevated aerosols would enhance the buoyancy of the free-tropospheric air, which inhibits cloud top entrainment (Johnson et al., 2004).

4. Uncertainty analysis

4.1. Effects of aerosol properties and solar geometry on fluxes

The distribution of aerosol properties in the real atmosphere is far from being homogenous (Eck et al., 2003). The SSA of absorbing aerosols change as a result of changes in fuel types and locations of fire burning. During the beginning of a fire burning season in southern Africa, aerosols are typically characterized with a high black carbon content from flaming combustion of grass. The SSA of aerosols in the southern Africa tends to increase as the dry season progresses.
possibly due to an increasing amount of smoldering combustion of woody fuels. Likewise, the
imaginary refractive index tends to decrease during the progression of a burning season (Eck et
al., 2013). The input aerosol properties for the RTM in this study relies on the Mongu
AERONET station, which is more than 1000km away from the southeast Atlantic. Laboratory
measurement of aerosols collected during SAFARI 2000 indicated that SSA of aerosols increases
during their lifetime due to condensation of scattering material (Abel et al., 2003). Hence,
aerosol properties would certainly differ between these two locations. Moreover, the change in θ₀
can also contribute to uncertainties of retrieved fluxes. Given these complications, one should
question how the variation of θ₀ and aerosol properties affect TOA upward fluxes, surface
downward fluxes, and vertical heating rates?

Figure 11 presents a sensitivity analysis of aerosol properties and θ₀ for both the upward and
downward fluxes, where both SSA and g vary by ±5% (Figure 11a). Both the upward and
downward fluxes increase with SSA, confirming the increase of fluxes with aerosol scattering
efficiency. Both the upward and downward fluxes increase with SSA. Specifically, an increase of
5% in SSA results in an increase of upward flux by ~25Wm⁻² and downward flux by ~15Wm⁻².
An increase in flux links to a decrease in forcing, which is consistent with the findings in Min
and Zhang (2014). Upward (downward) fluxes decrease (increase) with increasing asymmetry
parameter but by ~4Wm⁻² from −5% to +5%. Since the forward scattering increases with the
value of the asymmetry parameter, downward fluxes would be expected to increase. Thus,
varying SSA exerts a greater influence than varying the asymmetry parameter on fluxes because
changing aerosol SSA affects its scattering and absorbing efficiency of solar radiation. In
contrast, altering the asymmetry parameter mainly affects the ratio of forward to backward
scattering.
Increasing $\theta_0$ reduces the upward and downward fluxes (Figure 11b). At TOA, the upward fluxes decrease from $\sim377\text{Wm}^{-2}$ to $\sim293\text{Wm}^{-2}$ from $\theta_0=10^\circ$ to $\theta_0=50^\circ$. At the surface, the downward fluxes decreases from $\sim440\text{Wm}^{-2}$ to $\sim205\text{Wm}^{-2}$ between these two solar zenith angles. Hence, solar attenuation is more sensitive to the change in solar zenith angles in the case of downward fluxes than that of upward fluxes. These trends are consistent with the change of fluxes with $\theta_0$ for cloud-embedded exponential dust layer (Su et al., 2008). Since the greatest change of AOT with height takes place in the lower troposphere, surface fluxes experiences a larger change in magnitude with $\theta_0$ than that for TOA fluxes in both studies.

### 4.2. Effects of aerosol properties on heating rates

Figure 12 shows the dependence of vertical heating rates on the uncertainty of SSA. For cloud-free EAL distributions, an increase in SSA results in a decrease of heating rate throughout the profile (a). For the original SSA under cloudy EAL distributions, heating rates at the cloud layer (0.9km) is $\sim7.6\text{Kday}^{-1}$ (Figure 12b). A 5% increase in SSA reduces the heating rates to $\sim7.1\text{Kday}^{-1}$, but a 5% decrease in SSA raises the heating rates to $\sim8\text{Kday}^{-1}$. Heating rates are higher for aerosols with lower SSA throughout the profile. In contrast, an SAL at 3.8km in cloud-free and above cloud cases only shows heating rate differences at the aerosol layer. In the cloud-free SAL scenario (Figure 12c), the heating rate increases (decreases) from $\sim17.7\text{Kday}^{-1}$ to $\sim20.6 (\sim14.4)\text{Kday}^{-1}$ with a 5% decrease (increase) of SSA at the aerosol layer. For SAL above clouds (Figure 12d), the heating rate increases (decreases) from $\sim23.8\text{Kday}^{-1}$ to $\sim27.4 (\sim19.5)\text{Kday}^{-1}$ with a 5% decrease (increase) of SSA at the aerosol layer. The change of $g$ by $\pm5\%$ results in a change in the heating rate only by an order of $10^{-2}\text{K/day}$, which is not surprising consider the fact that changes in fluxes with changing $g$ by $\pm5\%$ is less than $5\text{Wm}^{-2}$ (Figure 11a).
5. Summary and conclusions

The impacts of vertical distributions of absorbing aerosols and clouds on the direct radiative forcing and radiative heating rates are examined using a delta-four stream RTM. Aerosol properties are acquired from the Mongu, Zambia AERONET station on 22 August 2006. For this study, the incoming SW flux at TOA is \(\sim 1122 \text{ Wm}^{-2}\) for \(\theta_0=35^\circ\) and \(\theta=25^\circ\). In clear sky conditions, the surface downward flux is \(\sim 852 \text{ Wm}^{-2}\) while TOA upward flux is \(\sim 66 \text{ Wm}^{-2}\). Two types of aerosol and cloud distributions are examined namely the EAL and the SAL distributions. Both of these scenarios can be found during the austral winter in southeast Atlantic according to CALIOP observations.

In an EAL distribution, the influence of AOT above clouds on TOA fluxes depends on the magnitude of COT. Below (Above) COT of 4, fluxes increase (decrease) with AOT of 1. Thus, aerosol scattering (absorbing) effect prevails below (above) this critical COT. This relationship resembles mineral dust above the Saharan desert where magnitude of surface albedo dictates the sign of radiative forcing at TOA (Patadia et al., 2009). At the surface, the downward fluxes decrease with both increasing AOT and COT. The TOA fluxes decrease as the exponential aerosol layer height increases above a fixed liquid cloud height with COT=10. Hence, the TOA radiative forcing increases with an increase of the 1/e aerosol layer height above a fixed cloud height at a constant COT. For an EAL with 1/e of AOT at 0.9km and a cloud layer at 0.9km, the upward TOA radiative forcing is \(\sim 11 \text{ Wm}^{-2}\). The radiative forcing reaches \(\sim 60 \text{ Wm}^{-2}\) when the 1/e of AOT increases to 6km while fixing the cloud layer at 0.9km. An increase in the cloud height with a fixed 1/e of EAL at 5km leads to an increased upward TOA fluxes as a result of higher reflection of downward fluxes for higher cloud tops. The surface downward radiative forcing increases from \(\sim 28 \text{ Wm}^{-2}\) to \(\sim 86 \text{ Wm}^{-2}\) with an increase in 1/e of EAL from 0.9km to 4.8km. The
higher changes in the downward forcing than the upward forcing with the same increase in the aerosol scale height occurs as a result of greater changes of AOT in the lower troposphere than that in the upper troposphere.

For AOT=1 and COT=10, heating rates peak at the cloud level in EAL distributions. SW heating rates reach \( \sim 12.4 \text{K day}^{-1} \) when both the \( \frac{1}{e} \) aerosol layer and clouds are located at \( \sim 0.9 \text{km} \). When the \( \frac{1}{e} \) aerosol layer raises to 6km at the same cloud height, heating rates decrease to \( \sim 6.7 \text{K day}^{-1} \) at the cloud height. Under pristine cloud conditions, heating rates for clouds at 0.9km and 4.8km are \( \sim 5.7 \text{K day}^{-1} \) and \( \sim 8.5 \text{K day}^{-1} \), respectively. However, when the \( \frac{1}{e} \) aerosol layers are included at 5km, heating rates are enhanced by \( \sim 1.5 \text{K day}^{-1} \) at the cloud level regardless of the cloud height. For a single aerosol layer and clouds located at 0.9km, the peak heating rate is \( \sim 17.9 \text{K day}^{-1} \). When the single aerosol layer shifts to 5km with clouds remaining at 0.9km, heating rates reach \( \sim 26.3 \text{K day}^{-1} \).

Figure 13 illustrates a consolidated summary of radiative fluxes and heating rates of six primary scenarios of aerosols and clouds that have been examined in this study. Heating rates are presented for the lowest 5km above the surface. It can be seen that the vertical distributions of aerosols and clouds modify atmospheric conditions and accurately quantifying these features are imperative to deduce their implications on climate systems.

The transportation of aerosols results in variations of aerosol properties obtained from the AERONET site to the southeast Atlantic depending on different stages of smoke. The uncertainty analysis reveals that both upward and downward fluxes increase with SSA. For cloud-embedded EAL distributions, vertical heating rates increase by \( \sim 0.5 \text{K day}^{-1} \) with a decrease of SSA by 5% at the cloud level. The entire heating profile in the rest of the troposphere also increases. For a SAL at 5km with underlying clouds, vertical heating rates can increase from
~23.8Kday\(^{-1}\) to ~27.4Kday\(^{-1}\) with a decrease of SSA by 5% at the aerosol level. Unlike cloud-embedded EAL distributions, heating rates outside the aerosol layer for the SAL cases change by less than a factor of 0.1Kday\(^{-1}\). A change of 5% in the asymmetry parameter only alters the heating rate by an order of 10\(^{-2}\) at the cloud level.

The direct radiative forcing of cloud-embedded absorbing aerosols and absorbing aerosols above clouds have been studied extensively using satellite observations and RTMs. Field campaigns would significantly improve these calculations since airborne sunphotometer could measure aerosol properties at actual scenes. Future studies will benefit from several upcoming field experiments such as Cloud-Aerosol-Radiation Interactions and Forcing (CLARIFY).

Additionally, quality-controlled data from ObseRvations of Aerosols above CLouds and their intEractionS (ORACLES) and Layered Atlantic Smoke Interactions with Clouds (LASIC) will also provide invaluable information for improving estimates of aerosol-cloud radiative effects.

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References


Figure 1. (a) The Terra MODIS true color composite (Red=0.67 um, Green=0.56 um and Blue=0.47 um) over southern Africa on 22 August 2006. The AERONET station from Mongu, Zambia is denoted by the orange dot at which aerosol properties are obtained.
Figure 2. Wavelength dependence of (a) SSA and (b) g in the SW interpolated and extrapolated from the AERONET station in Mongu, Zambia (15.25°S 23.15°E) on 22 August 2006 at 1408Z. The higher (lower) grey line represent the +5% (−5%) from the original SSA and g. Dots represent the wavelengths that are used in the RTM.
Figure 3. (a) The CALIOP vertical feature mask feature layers for 19 August 2009 at 1339Z. (b) The aerosol subtype layers for the identical time and location as (a). (c) As in (a), but for 13 August 2006 at 1323Z. (d) As in (b), but for the identical time and location as (c).
Figure 4. Vertical distributions of aerosol loading in an exponential aerosol distribution with scale heights from 1km to 5km. The vertical dashed line joins the stepwise $1/e$ aerosol loading of various scale heights assigned in this study.
Figure 5. (a) TOA upward and (b) the surface downward SW flux as a function of AOT for various COTs. The cloud effective radius in this model is 10µm at ~904mb (~0.9km) and the EAL reaches 1/e at 5km above sea-level under a tropical atmosphere.
Figure 6. The (a) upward and (b) downward radiative forcing for clouds in EAL. The cloud layer is fixed at 1.8km above the surface. The aerosol scale height indicate the height at which AOT decreases to $\sim 1/e$. 
Figure 7. TOA upward radiative fluxes as a function of cloud height for upward TOA (black) and downward (gray) fluxes for smoke-free (dashed lines) and exponential smoke AOT with $1/e$ at 5km (solid lines).
Figure 8. Vertical heating rates for clear sky, aerosol only, cloud only, and aerosol (smoke) above cloud of varied smoke layer height. The cloud is located at 0.9km with COT=10 and AOT=1 in the RTM.
Figure 9. SW heating rates for clouds of various heights (a) without and (b) with EAL distributions based on the same aerosol and cloud properties described in Figure 8.
Figure 10. SW heating rates for varied single aerosol layer (denoted by “SL”) (a) excluding clouds and (b) above liquid clouds based on the same aerosol and cloud properties described in Figure 8.
Figure 11. (a) Upward and downward SW flux as a function of changes in SSA and g by ±5% in the SW bands with AOT=1 at 5km, COT=10 and cloud effective radius of 10 µm at 904 mb. $\theta_0$ and $\theta$ are 30° and 20°, respectively. Upward flux is denoted by “up” with solid lines and downward flux is denoted by “dn” with dashed lines. (b) TOA (black) and surface (gray) fluxes as a function of $\theta_0$. 
Figure 12. Sensitivity analysis of vertical heating rates with varied SSA for (a) cloud-free exponential aerosol distributions, (b) cloud-embedded exponential aerosol distributions, (c) cloud-free single aerosol layer, and (d) single aerosol layer above clouds. The $1/e$ of the exponential aerosol distributions is located at 5km for (a) and (b). The single aerosol layers are located at 3.8km for (c) and (d). Aerosol and cloud properties follow those described in Figure 8.
Figure 13. A schematic showing a consolidated summary of radiative fluxes (in Wm\(^{-2}\)) and radiative heating rates (in Kelvin per day) for six scenarios including clear sky, single aerosol layer (SAL), exponential aerosol layer (EAL), cloud-embedded EAL, SAL above clouds, and pristine clouds using the RTM. The TOA downward fluxes is ~1122Wm\(^{-2}\). The input cosine of solar zenith angle and the cosine of viewing zenith angle are 0.82 and 0.90, respectively. Brown boxes illustrate vertical distributions of aerosols where the magnitude of the AOT in each layer is delineated by color intensities. AOT for the SAL cases are 1.0 and COT for all clouds are 10. The \(1/e\) for the EAL cases are located at 4km. All values are rounded off to the nearest whole unit. The top of the EAL, the top of the SAL, and the cloud top shown in the schematic are approximately 5km, 4km, and 2km, respectively. Note that the schematic is not presented in scale.