Aerosol Direct Radiative Effect Sensitivity Analysis

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ABSTRACT

Both to reconcile the large range in satellite-based estimates of the aerosol direct radiative effect (DRE) and to optimize the design of future observing systems, this study builds a framework for assessing aerosol DRE uncertainty. Shortwave aerosol DRE radiative kernels (Jacobians) were derived using the MERRA-2 reanalysis data. These radiative kernels give the differential response of the aerosol DRE to perturbations in the aerosol extinction coefficient, aerosol single-scattering albedo, aerosol asymmetry factor, surface albedo, cloud fraction, and cloud optical depth. This comprehensive set of kernels provides a convenient way to consistently and accurately assess the aerosol DRE uncertainties that result from observational or model-based uncertainties. The aerosol DRE kernels were used to test the effect of simplifying the full vertical profile of aerosol scattering properties into column-integrated quantities. This analysis showed that, although the clear-sky aerosol DRE can be had fairly accurately, more significant errors occur for the all-sky DRE. The sensitivity in determining the broadband spectral dependencies of the aerosol scattering properties directly from a limited set of wavelengths was quantified. These spectral dependencies can be reasonably constrained using column-integrated aerosol scattering properties in the midvisible and near-infrared wavelengths. Separating the aerosol DRE and its kernels by scene type shows that accurate aerosol properties in the clear sky are the most crucial component of the global aerosol DRE. In cloudy skies, determining aerosol properties in the presence of optically thin cloud is more radiatively important than doing so when optically thick cloud is present.

1. Introduction

The most elementary understanding of how aerosols influence forcing of Earth’s climate begins with the aerosol direct radiative effect (DRE)—the effect of aerosol scattering and absorption on the shortwave radiation at the top of the atmosphere (TOA). Passive satellite remote sensing estimates of the aerosol DRE (Yu et al. 2006, and references therein) can be highly uncertain outside of cloud-free ocean scenes (Li et al. 2009; Kokhanovsky et al. 2010). Advances have been made in passive retrievals over optically thick cloud (Torres et al. 2007; Waquet et al. 2009, 2013; de Graaf et al. 2012; Jethva et al. 2013; Meyer et al. 2015) and land (Dubovik et al. 2011; Lyapustin et al. 2011a,b), but, even with these advanced techniques, model information is needed to make truly global all-sky estimates (e.g., Lacagnina et al. 2017). Active remote sensing—for example, the lidar on board the Cloud–Aerosol Lidar and Infrared Pathfinder Satellite (CALIPSO; Winker et al. 2009, 2010)—allows for more complete and consistent retrievals in all-sky conditions over both land and ocean (Henderson et al. 2013; Oikawa et al. 2013; Matus et al. 2015; Oikawa et al. 2018). However, CALIPSO’s detection sensitivity limits the accuracy of the aerosol DRE inferred from its standard data products (Thorsen and Fu 2015; Thorsen et al. 2017).

The shortcomings of all current satellite remote sensing observations are apparent when comparing aerosol DRE estimates over clear-sky ocean, which vary by ±1.25 W m\(^{-2}\) (Yu et al. 2006; Henderson et al. 2013; Oikawa et al. 2013; Matus et al. 2015). This is despite uncertainties being minimized over clear-sky ocean for most types of remote sensing retrieval techniques. In addition, ±1.25 W m\(^{-2}\) is considerably larger than the stated uncertainties in each of these studies. The large range in observation-based estimates of aerosol DRE and its disconnect from existing...
uncertainty estimates demonstrate the difficulty in comprehensively probing and quantifying the associated uncertainties. In this study, we focus on improving uncertainty estimates in remote sensing–based calculations of the aerosol DRE by performing detailed sensitivity calculations. These sensitivity calculations are also a pragmatic way to identify how the current observational uncertainties might be reduced by future observing systems.

Our analysis is similar to that of McComiskey et al. (2008), who derived clear-sky aerosol DRE sensitivities over three ground-based sites. Here, a more thorough analysis is made by performing sensitivity calculations using 1 year of 3-hourly Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2), data (Gelaro et al. 2017). By using MERRA-2, a reasonably realistic representation of the covariability of aerosols, clouds, and surface albedo is captured along with fully resolving the sensitivities spatially, temporally, spectrally, and vertically. By fully quantifying the dimensionality of the sensitivities, the impact of simplifications to them can be assessed—an important consideration since no single source of observations completely samples the relevant properties in all dimensions.

A description of computing the aerosol DRE from the MERRA-2 inputs and comparisons of these computed fluxes to those observed by Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996) is given in section 2. Section 3 outlines how systematic perturbations to the inputs are made to determine the sensitivity of the aerosol DRE to each variable: we derive aerosol DRE partial derivatives (Jacobians), also known as radiative kernels (Soden et al. 2008; Shell et al. 2008) or radiative efficiencies (e.g., Henze et al. 2012). These radiative kernels are a linearization of the aerosol DRE calculation, and this assumption of linearity is tested in section 3 by comparing calculations of the speciated aerosol DRE with that approximated by the kernels. Section 4 presents and discusses the aerosol DRE kernels. This comprehensive set of aerosol radiative kernels provides a convenient way to consistently assess the aerosol DRE uncertainties that result from uncertainties in observational or model-based properties.

The aerosol DRE kernels are then used to explore the implications of common vertical (section 5a) and spectral (section 5b) simplifications made in remote sensing retrievals of aerosol scattering properties. The relative importance of the aerosol DRE and the average radiative kernels in various scene types (e.g., ocean, land, clear, cloudy) is quantified in section 5c. Concluding remarks are given in section 6.

2. Computing aerosol direct radiative effect

The aerosol DRE is computed as the difference between two sets of radiative flux calculations:

\[
DRE_{\text{all}} = \left[ F^{\dagger}_t(TOA)_{\text{all}} - F^{\dagger}_t(TOA)_{\text{all}} \right] - \left[ F^{\dagger}_t(TOA)_{\text{clear}} - F^{\dagger}_t(TOA)_{\text{clear}} \right],
\]

that is, the difference between the net (downward minus upward) shortwave flux \( F \) at the TOA with both clouds and aerosols included (all sky) and with aerosols removed (clear sky). The clear-sky aerosol DRE is computed in a similar fashion as the difference between the clear-sky flux (the flux computed with clouds removed) and the clear-clean-sky flux (without clouds and aerosols):

\[
DRE_{\text{clear}} = \left[ F^{\dagger}_t(TOA)_{\text{clear}} - F^{\dagger}_t(TOA)_{\text{clear}} \right] - \left[ F^{\dagger}_t(TOA)_{\text{clear_clean}} - F^{\dagger}_t(TOA)_{\text{clear_clean}} \right].
\]

In this study we focus only on the shortwave aerosol DRE at the TOA, although the methods presented here are general enough for use in investigating the aerosol DRE sensitivities in the longwave, at the surface and for radiative heating within the atmosphere. The calculation of radiative fluxes is performed with the NASA Langley Fu–Liou radiative transfer model (NFLRTM) that is fully described in Rose et al. (2013). Briefly, the NFLRTM is based on Fu (1991) and Fu and Liou (1992, 1993) with updates to the shortwave and longwave gaseous absorption (Kato et al. 1999; Kratz and Rose 1999). The effects of ice cloud crystals are included using Fu (1996) and Fu et al. (1998) and liquid cloud droplets using Hu and Stamnes (1993). The radiative transfer equation is solved using the delta-Eddington two-stream solution (Fu 1991).

a. Inputs

The necessary inputs for the radiative transfer calculations are mostly taken from the MERRA-2 reanalysis (Gelaro et al. 2017). Relative to the original MERRA reanalysis (Rienecker et al. 2008, 2011), and of particular relevance to this work, is the addition of simulated aerosols by radiatively coupling the Goddard Earth Observing System, version 5 (GEOS-5), atmospheric model and data assimilation system (Rienecker et al. 2008; Molod et al. 2015) to the Goddard Chemistry, Aerosol, Radiation, and Transport model (GOCART; Chin et al. 2002; Colarco et al. 2010). GOCART simulates 15 externally mixed aerosol tracers: dust (five size bins), sea salt (five size bins), hydrophobic and hydrophilic black and organic
carbon, and sulfate. Aerosol observations from MODIS (Remer et al. 2005; Levy et al. 2007), MISR (Kahn et al. 2005), and AERONET (Holben et al. 1998) are assimilated in MERRA-2 (Randles et al. 2017). The MERRA-2 aerosol properties show reasonable agreement relative to independent space and airborne aerosol observations (Randles et al. 2017; Buchard et al. 2017).

An overview of the MERRA-2 variables and corresponding data collections used as input into the NFLRTM are shown in Table 1. In essence, we are looking to repeat, and subsequently modify, the radiative transfer calculations already done and archived in the MERRA-2 data collections. Comparisons with the nominal MERRA-2 shortwave fluxes and aerosol DRE are shown in section 2b. Calculations are performed using a single year (2007) of data, as is typical for kernel calculations (e.g., Soden et al. 2008; Shell et al. 2008) due to their computational cost. Therefore, the kernels computed in this study will be most accurate when applied to measurements/data from the same year they were created with (i.e., 2007). However, the impact of interannual variability on the kernels is expected to be small since interannual variations in the aerosol DRE itself is typically small (e.g., Loeb and Manalo-Smith 2005; Matus et al. 2019). The 3-hourly MERRA-2 inputs are regridded from their native grid of $\frac{3}{8}^\circ$ by $\frac{1}{8}^\circ$ (longitude by latitude) to $2.5^\circ$ by $2.5^\circ$ to reduce the number of radiative transfer calculations needed for the kernels (section 3) to a manageable amount. Details on the how the MERRA-2-based inputs are assembled for the radiative transfer calculations are given in the appendix.

### Table 1. MERRA-2 data collections and variables used in the radiative transfer calculations.

<table>
<thead>
<tr>
<th>Collection</th>
<th>Variables</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>inst3_3d_asm_Nv</td>
<td>$T$, QV, O3, CLOUD</td>
<td>Temperature, Specific humidity, Ozone, Cloud fraction</td>
</tr>
<tr>
<td>tavg1_2d_rad_Nx</td>
<td>ALBVISDR and ALBVISDF, ALBNIRD, ALBNIRD</td>
<td>Direct and diffuse visible surface albedo, Direct and diffuse near-infrared surface albedo</td>
</tr>
<tr>
<td></td>
<td>CLDLOW, CLDMID, and CLDHGH</td>
<td>Area fraction of the low, middle, and high cloud super layer</td>
</tr>
<tr>
<td></td>
<td>TAULOW, TAUMID, and TAUHGH</td>
<td>Optical depth of the low, middle, and high cloud super layer</td>
</tr>
<tr>
<td>tavg3_3d_cld_Nv</td>
<td>TAUCLW and TAUCLI, INCLUDQL and INCLUDQI</td>
<td>Liquid and ice cloud optical depth, Liquid and ice cloud mass mixing ratio</td>
</tr>
<tr>
<td>inst3_3d_aer_Nv</td>
<td>RH, AIRDENS, DU001, DU002, SS001, SS002, SS003, SS004, SS005, SO4, BCPHOBIC and BCPHILIC, OCPHOBIC and OCPHILIC</td>
<td>Relative humidity, Air density, Dust mass mixing ratio, bins 1–5, Sea salt mass mixing ratio, bins 1–5, Sulfate mass mixing ratio, Hydrophobic and hydrophilic black carbon mass mixing ratio, Hydrophobic and hydrophilic organic carbon mass mixing ratio</td>
</tr>
</tbody>
</table>

### b. Validating fluxes

As explained in the appendix, instantaneous and time-averaged MERRA-2 data collections are mixed together to create the necessary inputs for the radiative transfer calculations. Because of these slightly mismatched times, the regridding to a coarser horizontal resolution, and the different radiative transfer models used, we do not expect to exactly replicate the radiative fluxes provided by MERRA-2. Regardless, comparisons to the nominal MERRA-2 fluxes are a useful evaluation to ensure that the time-shifted inputs do not cause egregious differences. A comparison of the aerosol DRE computed in this study and that from MERRA-2 is given in Fig. 1. Overall, the calculations in this work give a slightly weaker (less negative) global mean aerosol DRE than MERRA-2. The all-sky and clear-sky aerosol DRE both have a global mean bias of 0.3 W m$^{-2}$ relative to the nominal MERRA-2 values. Jones et al. (2017) showed that a similar aerosol DRE bias (0.2–0.3 W m$^{-2}$) can be caused solely by the choice of radiative transfer model. Therefore, we consider the differences in Fig. 1 to be reasonable small and do not expect them to significantly impact our sensitivity and uncertainty estimates detailed in the following sections.

An additional assessment of the computed all-sky and clear-sky shortwave TOA fluxes is made in Table 2. Table 2 compares our computed fluxes and the nominal MERRA-2 ones with Edition 4 of the CERES Energy Balance and Filled (EBAF)-TOA observations (Loeb et al. 2018). EBAF-TOA improves the accuracy of the CERES-observed fluxes by adjusting the net flux using an in situ derived value from Johnson et al. (2016).
and infers clear-sky fluxes from partly cloudy CERES footprints. The fluxes computed in this study give similar global mean root-mean-square (RMS) errors as the nominal MERRA-2 fluxes. However, in terms of global mean bias errors, the fluxes computed in this study agree better (by about a factor of 2) with EBAF-TOA than the nominal MERRA-2 fluxes. When compared with EBAF-TOA, both our computed fluxes and MERRA-2’s show similar regional patterns of differences (not shown). While we cannot completely rule out compensating biases, it is encouraging that our computed fluxes give smaller bias errors than the nominal MERRA-2 fluxes relative to CERES EBAF-TOA.

3. Radiative kernels

The sensitivity of the aerosol DRE to the inputs assembled from the MERRA-2 data collections is

<table>
<thead>
<tr>
<th>RMS error</th>
<th>Bias error</th>
</tr>
</thead>
<tbody>
<tr>
<td>All sky</td>
<td>Clear sky</td>
</tr>
<tr>
<td>All sky</td>
<td>Clear sky</td>
</tr>
</tbody>
</table>

This study 15.4 4.3 -3.5 1.0
MERRA-2 15.3 4.3 -6.2 2.0
quantified by deriving aerosol DRE partial derivatives (Jacobians), also known as radiative kernels (Soden et al. 2008; Shell et al. 2008). McComiskey et al. (2008) performed a similar analysis for the clear-sky aerosol properties, the second term in both Eqs. (5) and (6) is zero so the kernels computed for those variables are approximated as

\[
\frac{\partial \text{DRE}_{\text{all}}}{\partial x} \approx \frac{\delta \text{F}_{\text{net, all}}}{\delta x} - \frac{\delta \text{F}_{\text{net, clean}}}{\delta x} = \frac{\text{F}_{\text{net, all}}(x + \delta x, y_1, \ldots, y_n) - \text{F}_{\text{net, all}}(x, y_1, \ldots, y_n)}{\delta x}
\]

where \( y_1, \ldots, y_n \) are the other variables beside the variable of interest \( x \) needed to calculate the radiative flux \( F \). Similarly, the clear-sky radiative kernel is approximated as

\[
\frac{\partial \text{DRE}_{\text{clear}}}{\partial x} \approx \frac{\delta \text{F}_{\text{net, clear}}}{\delta x} - \frac{\delta \text{F}_{\text{net, clear, clean}}}{\delta x} = \frac{\text{F}_{\text{net, clear}}(x + \delta x, y_1, \ldots, y_n) - \text{F}_{\text{net, clear}}(x, y_1, \ldots, y_n)}{\delta x}
\]

Although not denoted in the above equations, all fluxes \( F \) and kernels \( \partial \text{DRE}/\partial x \) are a function of the NRLRTM wavelength bands (Table 3), latitude, longitude, and time. The kernels are also a function of model layer for quantities that vary in the vertical. Kernels are created for the variables: aerosol extinction coefficient \( \alpha \), aerosol single-scattering albedo \( \omega_s \), aerosol asymmetry factor \( g \), surface albedo \( a \), cloud fraction, cloud optical depth (COD), and cloud particle size (liquid effective radius and ice generalized effective diameter).

Repeated evaluations of Eqs. (5) and (6) are needed to avoid nonlinearity in the ratios. For the aerosol scattering properties (\( \alpha \), \( \omega_s \), and \( g \)), this means determining the kernels separately for model layers by performing radiative transfer calculations to isolate the impact of a single perturbed model layer at a time. Likewise, for the cloud properties (fraction, COD, and particle size), the overlapping cloud layers must be perturbed individually. Note that for the aerosol scattering properties, the second term in both Eqs. (5) and (6) is zero so the kernels computed for those variables are

**Table 3. Shortwave wavelength bands in the NASA Fu–Liou radiative transfer model.**

<table>
<thead>
<tr>
<th>Band no.</th>
<th>Wavelength range (( \mu m ))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.1754–0.2247</td>
</tr>
<tr>
<td>2</td>
<td>0.2247–0.2439</td>
</tr>
<tr>
<td>3</td>
<td>0.2439–0.2857</td>
</tr>
<tr>
<td>4</td>
<td>0.2857–0.2985</td>
</tr>
<tr>
<td>5</td>
<td>0.2985–0.3225</td>
</tr>
<tr>
<td>6</td>
<td>0.3225–0.3575</td>
</tr>
<tr>
<td>7</td>
<td>0.3575–0.4375</td>
</tr>
<tr>
<td>8</td>
<td>0.4375–0.4975</td>
</tr>
<tr>
<td>9</td>
<td>0.4975–0.595</td>
</tr>
<tr>
<td>10</td>
<td>0.595–0.6896</td>
</tr>
<tr>
<td>11</td>
<td>0.690–0.794</td>
</tr>
<tr>
<td>12</td>
<td>0.794–0.889</td>
</tr>
<tr>
<td>13</td>
<td>0.889–1.042</td>
</tr>
<tr>
<td>14</td>
<td>1.042–1.410</td>
</tr>
<tr>
<td>15</td>
<td>1.410–1.9048</td>
</tr>
<tr>
<td>16</td>
<td>1.9048–2.5</td>
</tr>
<tr>
<td>17</td>
<td>2.5–3.5088</td>
</tr>
<tr>
<td>18</td>
<td>3.5088–4</td>
</tr>
</tbody>
</table>

Our analysis here is more detailed as the aerosol DRE kernels are derived globally for all-sky conditions. We follow a similar approach as Thorsen et al. (2018), who, using observational datasets, decomposed the individual contributions to the variability of radiation budget and derived radiative kernels. The aerosol properties used in Thorsen et al. (2018) came from the MATCH MODIS aerosol assimilation product (Collins et al. 2001). In this study, we instead opt to use the more detailed and complete set of aerosol properties provided by MERRA-2.

From Eqs. (1) and (2), the partial derivatives of the all-sky and clear-sky aerosol DRE with respect to a variable \( x \) are

\[
\frac{\partial \text{DRE}_{\text{all}}}{\partial x} = \frac{\partial \text{F}_{\text{net, all}}}{\partial x} - \frac{\partial \text{F}_{\text{net, clean}}}{\partial x}
\]

and

\[
\frac{\partial \text{DRE}_{\text{clear}}}{\partial x} = \frac{\partial \text{F}_{\text{net, clear}}}{\partial x} - \frac{\partial \text{F}_{\text{net, clear, clean}}}{\partial x}
\]

respectively. These partial derivatives are approximated by imposing a small perturbation \( \delta x \) to the base-state value of \( x \); that is,
are both aerosol DRE kernels and kernels for the all-sky/clear-sky fluxes.

The size of the perturbation is taken as a 1% relative increase over the base value (i.e., $\delta x = 0.01x$). The choice of 1% is somewhat arbitrary because we desire a perturbation that is small enough to induce a linear impact on the radiative flux but large enough to avoid noise from numerical truncation errors. We found that perturbations of smaller than 1% (e.g., 0.01%) produced near-identical kernels apart from outliers that result from truncation errors (not shown).

**Linearity**

Although determining the aerosol DRE kernels requires a significant amount of radiative transfer calculations, once produced they provide a convenient and quick way to approximate the radiative impact of a perturbation or uncertainty in a variable. The accuracy of using the kernels in this way is dependent on the assumption that the linearization of the aerosol DRE calculation is valid for the magnitude of the perturbation/uncertainty they are applied to. This makes a general quantification of the nonlinear errors difficult as they depend on the specific application of the kernels. Therefore, in this section, the errors in using the kernels to approximate the speciated aerosol DRE are quantified and those results are used to discuss potential errors in more broad applications. This analysis is performed for the all-sky DRE only.

The aerosol DRE of an aerosol species $i$ is determined by taking the difference between the flux $F$ computed with all aerosol species and that computed excluding species $i$, that is,

$$DRE_i = F_{\text{net}}(\alpha^0, \omega^0, g^0, \ldots) - F_{\text{net}}(\alpha', \omega', g', \ldots),$$  

(7)

where $\alpha^0$, $\omega^0$, and $g^0$ are the nominal aerosol scattering properties and $\alpha'$, $\omega'$, and $g'$ are the aerosol properties with species $i$ removed. Since the aerosols are externally mixed (see the appendix), the aerosol layer properties with species $i$ removed are

$$\alpha' = \alpha^0 - \alpha',$$  

(8)

$$\omega' = \frac{\alpha^0 \omega^0 - \alpha' \omega'}{\alpha'},$$  

(9)

$$g' = \frac{\alpha^0 \omega^0 g^0 - \alpha' \omega' g'}{\alpha' \omega' g'},$$  

(10)

The ellipses in Eq. (7) are used to denote that other inputs beside the aerosol scattering properties are the same in each calculation. The speciated aerosol DRE can also be approximated without additional radiative transfer calculations using the aerosol DRE kernels

$$DRE_{i}^k = \sum_{x,z} \frac{\partial DRE_{all}}{\partial \alpha} \Delta \alpha + \frac{\partial DRE_{all}}{\partial \omega} \Delta \omega + \frac{\partial DRE_{all}}{\partial g} \Delta g,$$

(11)

where

$$\Delta \alpha = \alpha^0 - \alpha',$$  

(12)

$$\Delta \omega = \omega^0 - \omega',$$  

(13)

$$\Delta g = g^0 - g'.$$  

(14)

The summation in Eq. (11) sums the contributions to the TOA DRE from the perturbed scattering properties in each wavelength band $\lambda$ and model layer $z$.

The speciated aerosol DRE computed with Eq. (11) is compared to the reference value [Eq. (7)] for each MERRA-2 species $i$ (Table 1) in Fig. 2. The box plots in Fig. 2a summarize the relative error found across all grid boxes and all times over the full year of calculations. For context, the global mean, annual mean aerosol DRE for each species is given in Fig. 2b. Errors are largest for the sulfate (SO$_4$) DRE as expected since sulfate has the largest DRE of any species. When the kernels are used to estimate the sulfate DRE, medians errors are around 11% but the interquartile range (IQR) and 95% confidence interval (CI) indicate there are occasions with larger errors with the IQR showing up to 42% and 95% CI up to 82% error. For other than SO$_4$, the errors in Fig. 2a are narrowly distributed with median errors less than 7%. Overall, the errors in Fig. 2a are acceptably small, although the longer tails of the distributions indicate that larger errors could be encountered when using the kernels for more refined spatial or temporal applications. While larger errors do exist for the sulfate DRE, completely ignoring all sulfate globally represents a large perturbation to the aerosol scattering properties. It is expected that these perturbations are significantly larger than the intended application of the kernels for making aerosol DRE uncertainty estimates using observational or model based errors.

4. Aerosol DRE kernels

Using the method discussed in the previous section, a year (2007) of 3-hourly kernels were computed and, in this section, the resulting kernels are examined. Since the aerosol scattering properties (i.e., extinction, single-scattering albedo, and asymmetry factor) and surface albedo vary as a function of wavelength, their kernels...
vary spectrally as well. A broadband (BB) kernel can be obtained by summing over the wavelength bands; that is,

$$\frac{\partial \text{DRE}}{\partial x} \bigg|_{\text{BB}} = \sum \frac{\partial \text{DRE}}{\partial x} \bigg|_{\lambda}$$ \hspace{1cm} (15)

where $\lambda$ are the wavelength bands in Table 3 and $x$ represents $\alpha$, $\omega$, $g$, or $a$. Instead of presenting these BB kernels for the aerosol scattering properties, we instead express the kernels relative to a perturbation in the midvisible (vis):

$$\frac{\partial \text{DRE}}{\partial x} \bigg|_{\text{vis}} = \sum \frac{\partial \text{DRE}((\lambda)}}{\partial x((\lambda))} \frac{x((\lambda))}{x(550 \text{ nm})}$$ \hspace{1cm} (16)

where the midvisible is taken to be 550 nm. Equation (16) is equivalent to assuming perfect knowledge of the spectral dependence in each variable. The impact of incomplete knowledge of these spectral shapes is explored further in section 5b. The kernels expressed at the midvisible provides the sensitivity relevant to the wavelength at which aerosol scattering properties are commonly expressed and/or retrieved. Since the sun’s emission peaks in the midvisible, the broadband kernels are often dominated by values in the midvisible (e.g., Redemann et al. 2006).

Figure 3 shows the vertically integrated, annual mean, global mean aerosol DRE kernels. Kernels are expressed in units of watts per meter squared per unit change in the respective variables; for example, Fig. 3a shows that if the aerosol optical depth (AOD) were to be increased by 1 globally then the all-sky aerosol DRE would be reduced by 10.2 W m$^{-2}$. The magnitudes of the DRE kernels are generally larger in the clear sky (Fig. 3b) than in the all sky (Fig. 3a), with the exception of single-scattering albedo (SSA) and, of course, the cloud properties. While the sensitivity to the COD is applicable to any all-sky DRE calculation, the cloud fraction kernel would only be relevant when performing radiative transfer calculations over a nonhomogeneous pixel or grid box. Also computed, but not shown here, were the aerosol DRE kernels for the liquid and ice cloud particle sizes. The sensitivities of the aerosol DRE to the cloud particle size was found to be very weak: the liquid-effective-radius and ice-generalized-effective-diameter global mean aerosol DRE kernels are $-0.0016$ and $-0.00012$ W m$^{-2} \mu$m$^{-1}$, respectively.

The spatial structures of the vertically integrated annual mean kernels for the aerosol scattering properties are shown in Fig. 4. The AOD kernels (Figs. 4a,d) are typically negative (i.e., an increase in AOD results in an increased amount of aerosol cooling) but can also be
positive in regions with a high surface albedo that increases the relative importance of aerosol absorption (e.g., Chylek and Coakley 1974). In the clear sky (Fig. 4d), the AOD kernel is fairly spatially uniform while the all-sky kernel is heavily modulated by the presence of clouds. Clouds mainly mask the sensitivity to AOD perturbations as the difference between all-sky and clear-sky AOD kernels (Fig. 4g) is significantly correlated with the spatial pattern of cloud fraction ($r = 0.58$; not shown).

The SSA aerosol DRE kernel is always negative (Figs. 4b,e) since any increase in the relative amount of scattering yields a cooling at the TOA. The SSA kernel is most sensitive in areas with larger amounts of aerosol absorption: regions with significant anthropogenic black carbon and biomass burning (Asia, Africa, and South America) as well as dust from northern Africa and central Asia (e.g., Samset et al. 2018). This sensitivity is further enhanced in regions with the largest surface albedos: the Sahara and the Arabian peninsula. The difference between all-sky and clear-sky SSA kernels (Fig. 4h) is more nuanced than a simple masking effect. The largest differences between all sky and clear sky are found in regions where aerosol occurs above cloud frequently: the southeastern Atlantic Ocean, tropical eastern Atlantic, and northwest Pacific Ocean (Zhang et al. 2016). In these regions, the all-sky SSA kernels are more negative due to the contrast between the large cloud albedo and the dark ocean surface.

Figure 5 gives the annual mean, zonal mean aerosol DRE kernels for the aerosol scattering properties. The clear-sky extinction kernel (Fig. 5d) is rather vertically uniform while the all sky (Fig. 5a) is less so, with the kernel increasing as pressure decreases. Both the SSA and asymmetry factor kernels (Figs. 5b,c,e,f) are larger in magnitude near the surface corresponding to the larger near-surface aerosol extinction. Relative to the clear-sky kernel, the all-sky asymmetry factor kernel shows a more muted sensitivity but with a similar vertical structure (Figs. 5c,f,i). The difference between the all-sky and clear-sky SSA kernels contains both positive and negative values (Fig. 5h). Near the surface (below about 850 hPa), the all-sky SSA kernel (Fig. 5b) is less sensitive (less negative) than in the clear sky (Fig. 5e); that is, clouds mostly serve to mask perturbations to the SSA. Above that level, the all-sky SSA kernel is slightly more negative (more sensitive) as a result of the increased importance of absorption in aerosol that resides above bright low cloud.

5. Sampling effects on aerosol DRE

The aerosol DRE kernels presented in the previous section are applicable to assessments that assume that both the spectral and vertical dependencies of the aerosol scattering properties are fully resolved. However, this is often not the case, particularly with remote sensing observations. Spectrally, most remote sensing techniques only obtain aerosol scattering properties at a few, at most, discrete wavelengths, necessitating some spectral interpolation and extrapolation to provide the full broadband spectrum required for the

![Fig. 3.](image_url)
radiative transfer modeling. Additionally, the vertical dependence of these scattering properties is often only coarsely resolved, particularly for passive remote sensing observations that only retrieve column-effective values. The following subsections quantify the impact of these simplifications on the aerosol DRE using the radiative kernels derived above. Additionally, this section also explores the relative importance of the aerosol DRE and the average kernels in various scene types since retrieval techniques and their accuracies vary greatly depending on the scene type being observed.

While each of these vertical, spectral, and scene type components have a spatial and temporal structure, the effects on the global mean DRE are mostly highlighted in this section. This is done both for the sake of brevity and because the global mean DRE is directly relevant to the bulk climate system in the linear forcing-feedback framework (e.g., Gregory 2004).

For the aerosol DRE kernels themselves, the impact of any inaccuracies in the MERRA-2 properties is lessened since they are used to linearize the behavior of the aerosol DRE near the nominal value of each variable. However, the analyses in this section rely on taking the MERRA-2 spectral and vertical dependencies as truth and are therefore strongly dependent on the accuracy of MERRA-2. Despite this, we expect the subsequent analysis to have some merit as the MERRA-2 aerosol properties show reasonable agreement relative to independent space and airborne aerosol observations (Randles et al. 2017; Buchard et al. 2017) although errors do exist (e.g., Christian et al. 2019).

a. Vertical simplifications

The impact on the aerosol DRE from the typical vertical simplifications made in a passive remote sensing retrieval (e.g., Dubovik et al. 2011) are quantified by applying midvisible kernels discussed in section 4. By using these midvisible kernels, we are assuming perfect knowledge of the spectral
dependence and quantify errors solely due to representing the 550-nm (vis) scattering properties as an AOD:

$$\text{AOD}_{\text{vis}} = \int_{\text{TOA}}^{\text{SFC}} \alpha_{\text{vis}}(z) \, dz,$$

where \( z \) is the layer height and SFC is the surface, a column-integrated SSA

$$\omega_{\text{vis}} = 1 - \frac{\text{AAOD}_{\text{vis}}}{\text{AOD}_{\text{vis}}} = \int_{\text{TOA}}^{\text{SFC}} \frac{\omega_{\text{vis}}(z) \alpha_{\text{vis}}(z) \, dz}{\int_{\text{SFC}}^{\text{TOA}} \alpha_{\text{vis}}(z) \, dz},$$

where AAOD is the absorption aerosol optical depth, and a column-integrated asymmetry factor

$$g_{\text{vis}} = \frac{\int_{\text{TOA}}^{\text{SFC}} g_{\text{vis}}(z) \omega_{\text{vis}}(z) \alpha_{\text{vis}}(z) \, dz}{\int_{\text{SFC}}^{\text{TOA}} \omega_{\text{vis}}(z) \alpha_{\text{vis}}(z) \, dz},$$

$$= \frac{\int_{\text{TOA}}^{\text{SFC}} g_{\text{vis}}(z) \omega_{\text{vis}}(z) \alpha_{\text{vis}}(z) \, dz}{\text{AOD}_{\text{vis}} - \text{AAOD}_{\text{vis}}}. \quad (19)$$

In addition, some passive retrieval techniques (e.g., Waquet et al. 2009; Dubovik et al. 2011; Ding et al. 2016; Wu et al. 2016; Xu et al. 2017; Hasekamp et al. 2019; Xu et al. 2019) also obtain a characteristic height of the aerosol extinction profile. Therefore, we represent the aerosol extinction profile as
\[ a_{\text{vis}}^c(z) = \text{AOD}_{\text{vis}} \frac{f(z)}{dz}, \quad (20) \]

where \( dz \) is the thickness of layer \( z \) and \( f(z) \) is the fraction of the AOD that resides in layer \( z \) which is defined using a scale height \( H \):

\[ f(z) = \frac{\exp(-z/H)}{\sum_z \exp(-z/H)} \quad (21) \]

The scale height \( H \) is obtained by fitting each MERRA-2 profile of layer AOD.

The difference between the aerosol scattering properties given by Eqs. (18)–(20) and their fully vertically resolved counterparts [i.e., \( \alpha_{\text{vis}}(z) \), \( \omega_{\text{vis}}(z) \), and \( g_{\text{vis}}(z) \)] gives the error intrinsic to a perfect set of observations of AOD, scale height, and the column-integrated SSA and asymmetry factor. To assess the impact of these differences on the aerosol DRE, the kernels are used:

\[
\Delta \text{DRE}^c(z) = \sum_z \left\{ \left( \frac{\partial \text{DRE}}{\partial \alpha} \right)_{\text{vis}} \left[ \alpha_{\text{vis}}^c(z) - \alpha_{\text{vis}}(z) \right] + \left( \frac{\partial \text{DRE}}{\partial \omega} \right)_{\text{vis}} \left[ \omega_{\text{vis}}^c(z) - \omega_{\text{vis}}(z) \right] + \left( \frac{\partial \text{DRE}}{\partial g} \right)_{\text{vis}} \left[ g_{\text{vis}}^c(z) - g_{\text{vis}}(z) \right] \right\}. \quad (22)
\]

Figure 6 shows the annual mean aerosol DRE errors from the aforementioned vertical simplifications along with the contribution from each term in Eq. (22). The clear-sky aerosol DRE can be accurately determined when using column-integrated values with a total global mean bias of only \(-0.06 \text{ W m}^{-2}\) (Fig. 6h). The relative insensitivity of the clear-sky aerosol DRE to vertical structure has been noted by numerous other studies (e.g., Haywood and Shine 1997; Liao and Seinfeld 1998; Yu et al. 2006; McComiskey et al. 2008; Guan et al. 2010). The impact of vertical structure on the all-sky aerosol DRE (Figs. 6a–d) is larger with larger errors particularly from the use of a column-integrated SSA (Fig. 6b). This leads to a global mean all-sky aerosol DRE bias of \(-0.22 \text{ W m}^{-2}\) (Fig. 6d) intrinsic to a perfect set of AOD, scale height, column-integrated SSA, and column-integrated asymmetry factor observations.

To fully assess the errors in a passive-based estimate of the aerosol DRE, the errors in Fig. 6 (i.e., those intrinsic to a simplified vertical representation of the aerosol scattering properties) need to be combined with errors in the retrieved quantities themselves. To compute the latter, the errors in AOD, column-integrated SSA, and column-integrated asymmetry factor can be combined with vertically integrated aerosol DRE kernels (e.g., Fig. 3 or Fig. 4). In addition, an assessment of the impact of a scale height retrieval error is needed. However, since a perturbation to the scale height changes the extinction in multiple vertical layers, one cannot create a scale height aerosol DRE kernel. Instead, to understand the impact of scale height error, a fixed perturbation is applied to the nominal values of the scale height. The resulting differences in the extinction profiles are combined with the aerosol DRE extinction kernels to compute the global mean DRE bias for each fixed perturbation to the scale height. Figure 7 gives the result of this analysis showing the global mean all-sky
and clear-sky aerosol DRE bias that results from imposing various scale height perturbations. The effect of scale height errors on aerosol DRE is not straightforward with some larger-scale height errors giving less DRE biases than smaller ones due to some fortunate cancellation of errors. In general, scale height biases yield larger aerosol DRE biases in the all-sky than in the clear sky. Loeb and Su (2010) also performed a similar analysis by perturbing the scale height by $1.8 \text{ km}$, which caused an $0.15 (0.05) \text{ W m}^{-2}$ all-sky (clear sky) aerosol forcing bias. Here, our analysis shows a DRE bias of about one-half of the magnitude of that in Loeb and Su (2010)—from Fig. 7, a $+0.8$-km scale height bias causes an aerosol DRE bias of $0.07 (-0.02) \text{ W m}^{-2}$ in the all sky (clear sky).

Rather than representing the vertical dependence of the AOD with a scale height, a Gaussian function is often used instead (e.g., Waquet et al. 2009; Dubovik et al. 2011; Wu et al. 2016; Hasekamp et al. 2019). Using Eq. (16) from Dubovik et al. (2011), we found that a Gaussian provided better fits to the layer AOD profiles with smaller chi-square statistics on average (not shown). However, after convolving with the aerosol DRE extinction kernel, the fits computed using a scale height gave smaller mean aerosol DRE errors. Therefore, we chose to just perform our analysis using the scale height in this study.

This section quantified the expected minimum errors in a purely passive remote sensing–based calculation of the aerosol DRE. However, there are several limitations to our analysis. First, we have assumed that we can perfectly correct for any nonaerosol signal (i.e., surface reflection, three-dimensional effects, and the influence of clouds for the all-sky DRE). This assumes the signal from clouds, whether above or below aerosols, can be corrected for. While passive techniques for aerosol above optically thick cloud have been demonstrated (Torres et al. 2007; Waquet et al. 2009, 2013; de Graaf et al. 2012; Jethva et al. 2013; Meyer et al. 2015), attempts at retrievals below cloud have been limited in scope (Roskovensky and Liou 2006; Pierce et al. 2010; Lee et al. 2013). Additionally, the reflected radiance received by a passive sensor is proportional to a vertical average, weighted by the attenuated backscatter specific to the scene and viewing geometry. This is not accounted for in the analysis above. However, by applying the aerosol DRE kernels this impact is somewhat taken into account since the kernels quantify the relative importance of each vertical layer’s contribution to the TOA flux.

b. Spectral simplifications

The calculation of aerosol DRE requires specifying the aerosol scattering properties for the full broadband spectrum: from about 175 nm to 4 $\mu$m (Table 3). Typically, remote sensing observations can only obtain scattering properties at a handful, at most, of discrete wavelengths. In this section, we test the impact of limited knowledge of the spectral shapes on the calculation of aerosol DRE. Determining these spectral dependencies is often achieved or complemented by the identification of the aerosol type (dust, sea salt, etc.) which then determines the spectral shapes from a predetermined aerosol optical model (e.g., Loeb and Su 2010; Henderson et al. 2013; Oikawa et al. 2013; Matus et al. 2015; Kato et al. 2018). Generalizing the uncertainty associated with this approach is difficult since it would involved determining the accuracy of each specific method of aerosol typing and the aerosol optical models. Instead, we assess errors in a more naive approach that would interpolate/ extrapolate the broadband scattering properties from those at a discrete set of wavelengths (channels). In other words, we seek to determine how many channels are needed to constrain the aerosol scattering property spectral shapes directly.

The aerosol DRE biases associated with a limited spectral knowledge is quantified for various combinations of channels: 340 nm [near ultraviolet (NUV)], 550 nm [midvisible (VIS)], 1020 nm [near infrared (NIR)], and 1640 nm [shortwave infrared (SWIR)]. Four different combinations of channels are considered: 1) NUV + VIS, 2) VIS + NIR, 3) NUV + VIS + NIR, and 4) NUV + VIS + NIR + SWIR. For each of these channel
combinations and in each vertical layer, an interpolant $q_x$ is determined using a cubic spline fit between $x$, where $x$ is an aerosol scattering property [$\alpha(z)$, $\omega(z)$, or $g(z)$], and $c$, where $c$ are the channel wavelengths (NUV + VIS, VIS + NIR, etc.). These interpolants are then used to determine the spectral shapes relative to the nominal mid-visible wavelength, that is, $$s_x(\lambda) = \frac{q_x(\lambda)}{q_x(550 \text{ nm})}. \quad (23)$$

where the interpolants $q_x$ are used to interpolate/extrapolate to the wavelength bands $\lambda$ needed for the radiative transfer (Table 3). The spectral shapes are then applied to the mid-visible aerosol scattering properties to obtain the broadband properties:

$$x'(\lambda) = s_x(\lambda)x(550 \text{ nm}). \quad (24)$$

Last, the associated aerosol DRE error is computed by applying the aerosol DRE kernels to the difference between the approximate spectral scattering properties ($x'$) and the nominal values:

$$\Delta \text{DRE}_{\lambda}' = \sum_{\lambda} \frac{\partial \text{DRE}}{\partial x}(\lambda)[x'(\lambda) - x(\lambda)]. \quad (25)$$

This process is repeated for each of the four channel combinations listed above and for each aerosol scattering property $x$: aerosol extinction, single-scattering albedo, and asymmetry factor. The total aerosol DRE error in each channel combination is determined by summing the contributions from each aerosol scattering property. A similar analysis is also performed but with column-integrated spectral shapes; that is, Eq. (23) is evaluated with $x$ representing the AOD, column-integrated SSA ($\omega_c$), and column-integrated $g$ ($g_c$). These column-integrated spectral shapes $x'_x$ are then used in lieu of $s_x$ in Eqs. (24) and (25) to determine the aerosol DRE errors.

Figure 8 summarizes the aerosol DRE annual mean, global mean bias that results from interpolating/extrapolating the aerosol scattering properties using the four channel combinations. Biases in the clear-sky aerosol DRE are generally slightly larger than the all sky, but both show a similar dependence on the chosen channels. Figure 8
shows that, for the combinations with just two channels, NUV + VIS gives larger biases than VIS + NIR mostly due to the biases caused by the SSA. By examining the spatial distribution of these biases (not shown), it is clear that the increased SSA bias for the NUV + VIS is due to a poorer fit to the SSA in regions dominated by dust. Dust absorption is stronger in the NUV than at longer wavelengths, which causes a fit using NUV + VIS to overestimate the dust SSA at longer wavelengths. Vertically resolving the spectral shapes (stippled bars in Fig. 8) typically reduces the DRE bias compared to using column-integrated values (hatched bars in Fig. 8), but this reduction is only noteworthy when four channels (NUV + VIS + NIR + SWIR) are used to fit the spectral shapes. Having four channels (NUV + VIS + NIR + SWIR) of vertically resolved aerosol scattering properties reduces the global mean aerosol DRE bias to near zero. With the approach adopted here, the aerosol DRE uncertainties in Fig. 8 should be considered upper limits of the true uncertainty since we are willfully ignoring the additional information that could be provided by aerosol typing. However, directly determining the spectral dependencies is a more straightforward approach because it sidesteps the need for an aerosol typing and optical models, both of which add extra layers of uncertainty into the calculation of aerosol DRE. Figure 8 shows that simply interpolating/extrapolating from column-integrated values in a mere two channels (VIS + NIR) gives an all-sky global mean aerosol DRE bias less than 0.14 W m⁻². Such an approach reduces the reliance on the additional complication of aerosol typing and developing aerosol optical models.

c. Scene type

Retrieval techniques and their accuracies vary greatly depending on the scene type being observed. Different approaches are warranted for passive retrievals of aerosol over dark (Kaufman et al. 1997; Levy et al. 2013) and bright surfaces (Hsu et al. 2004; Lyapustin et al. 2011a,b). Passive retrievals over optical-thick cloud require different specialized algorithms (Torres et al. 2007; Waquet et al. 2009, 2013; de Graaf et al. 2012; Jethva et al. 2013; Meyer et al. 2015). From an active remote sensing perspective, while the same general retrieval using the CALIPSO lidar can obtain aerosol optical properties in all scene types (Young and Vaughan 2009; Young et al. 2018), techniques with improved accuracies are applicable over ocean surfaces (Josset et al. 2008; Venkata and Reagan 2016) and for aerosols over optically thick clouds (Hu 2007; Liu et al. 2015). Given all this variety, in the pursuit of making an observation-based global all-sky estimates of aerosol DRE, it is critical to understand the relative importance of the aerosol DRE in various scene types as well as the sensitivities to retrieval uncertainties in these scenes.

Figure 9 shows the annual mean, global mean midvisible aerosol DRE kernels in the clear sky (Fig. 9a), in profiles with thin cloud (COD < 2; Fig. 9b), and in profiles with thick cloud (COD ≥ 2; Fig. 9c). The aerosol DRE kernels are further separated into the means over ocean (solid bars) and over land or sea ice (hatched bars). Also given in Fig. 9, in brackets, is the annual mean, global mean aerosol DRE themselves in each scene type. Both the kernels and the aerosol DRE averages in Fig. 9 have been weighted by area and their respective frequencies so that summing the values across Figs. 9a–c yields the all-sky global mean value (i.e., Figs. 1a and 3a). Note that mean values given for clear columns in Fig. 9a differ from the clear-sky values referenced earlier (e.g., Fig. 3b) since the former are weighted by the clear-sky fraction.

For both clear and cloudy columns, the aerosol DRE over ocean is larger by about a factor of 2–3 than over land. The all-sky aerosol DRE is dominated by the clear sky (−1.31 W m⁻²), but there is still a significant cooling from aerosols in profiles with thin clouds (−0.41 W m⁻²). The smallest contribution to the all-sky aerosol DRE comes from profiles with thicker clouds where the presence of aerosol above these clouds contributes 0.26 W m⁻² of warming. Despite the considerable attenuation that aerosols above optically thick clouds have received (Chand et al. 2009; Wilcox 2012; Peters et al. 2011; de Graaf et al. 2012, 2014; Meyer et al. 2013, 2015; Zhang et al. 2014, 2016; Peers et al. 2015; Feng and Christopher 2015; Kacenelenbogen et al. 2019), Fig. 9 reveals that, radiatively, aerosols in presence of thin cloud are a more important component of the global all-sky aerosol DRE with an aerosol DRE that is 60% larger in magnitude. This conclusion is even more strongly supported by studies that have made CALIPSO-based DRE estimates. Oikawa et al. (2018) showed that the magnitude of the aerosol DRE is about a factor of 7 larger in instances of aerosol below cloud rather than when aerosol is above cloud (their Table 2). The aerosol DRE from Matus et al. (2015) was larger by a factor of 12 in magnitude in columns with thin cloud (COD < 1) than those with thick cloud (COD > 1). While both Oikawa et al. (2018) and Matus et al. (2015) based their calculations on the standard CALIPSO data products, Kacenelenbogen et al. (2019) performed their own, more accurate, CALIPSO retrievals of AOD above cloud. The Kacenelenbogen et al. (2019) method gives a larger (more positive) above-cloud aerosol DRE than that inferred from the standard CALIPSO data products. Comparing the above-cloud aerosol DRE from Kacenelenbogen et al. (2019) (their Table 4) with
below-cloud aerosol DRE values from Matus et al. (2015) and Oikawa et al. (2018) shows that the DRE magnitude is ~3–4 times as large in columns with aerosol below cloud. So, regardless of which CALIPSO-based estimate is examined, the MERRA-2-based aerosol DRE values in Fig. 9 likely underestimate the relative importance of the aerosol DRE in columns with thin clouds.

Figure 9 also reveals the relative sensitivities of the aerosol DRE to retrieval uncertainties (i.e., the aerosol DRE kernels) across different scenes. The aerosol DRE kernels show that, overall, it is most important to have accurate inputs in clear-sky scenes. For cloudy scenes, the aerosol DRE kernels are more sensitive in the case of thin cloud (COD < 2) rather than thick (COD ≥ 2) with the exception of SSA and cloud fraction. The aerosol DRE kernels are typically larger over ocean, except in the clear sky where the land SSA kernel is larger. As described above, the values in Fig. 9 are a convolution of mean DRE sensitivity in a particular scene type and the frequency with which that scene type occurs globally. While these frequency-weighted kernels give the sensitivity relevant to the global mean DRE, the unweighted values are also insightful for different applications. Figure S1 in the online supplemental material gives the mean DRE kernels that are not weighted by the scene type’s frequency of occurrence.

### 6. Summary and conclusions

Aerosol direct radiative effect (DRE) radiative kernels (Jacobians) were derived using 1 year of 3-hourly MERRA-2 data (Gelaro et al. 2017) by imposing small systematic perturbations to the base-state values. Aerosol DRE kernels where derived for the aerosol extinction coefficient, aerosol single-scattering albedo (SSA) aerosol asymmetry factor, surface albedo, cloud fraction, and cloud optical depth. This comprehensive set of aerosol radiative kernels provides a convenient way to consistently assess the aerosol DRE uncertainties that result from observational or model-based uncertainties. While the focus here was only on the shortwave aerosol DRE at the TOA, the methodologies presented are general enough for use in investigating the aerosol DRE sensitivities in the longwave, at the surface, and for radiative heating within the atmosphere.

The presumption that the aerosol DRE kernels represent a linearization of the aerosol DRE was confirmed by the overall good agreement between calculations of the true speciated aerosol DRE and that approximated by the kernels. Additionally, the computed shortwave fluxes were shown to agree reasonably well to those from Edition 4 of the CERES EBAF-TOA observations (Loeb et al. 2018) and the nominal MERRA-2 fluxes.

The aerosol DRE kernels resolve sensitivities spatially, temporally, spectrally, and vertically, enabling the
ability to explore simplifications to these dimensions. Aerosol DRE errors caused by the simplified vertical representation of aerosol scattering properties inherent to passive remote sensor retrievals were quantified using the aerosol DRE kernels. It was shown that errors in the clear-sky aerosol DRE were typically small. However, the impact of limited knowledge on the aerosol vertical structure was more significant for the all-sky aerosol DRE with a total aerosol DRE bias of $-0.22 \text{ W m}^{-2}$. The majority of this bias was due to simplifying the SSA profiles into column-integrated SSAs.

Since remote sensing techniques typically only obtain aerosol scattering properties at a few discrete wavelengths, the effect of using limited sets of wavelengths to interpolate/extrapolate the aerosol scattering properties to the full broadband spectrum required for radiative transfer modeling was examined. Having four channels (near-ultraviolet, midvisible, near-infrared, and shortwave infrared) of vertically resolved aerosol scattering properties nearly eliminates the aerosol DRE uncertainty (in terms of global mean bias) due to determining the broadband spectral dependencies. A more modest combination of column-integrated aerosol scattering properties in two channels (midvisible and near-infrared) was shown to cause an all-sky global mean aerosol DRE bias of $0.14 \text{ W m}^{-2}$. Since determining spectral dependencies is often achieved or complemented by the identification of the aerosol type and a predetermined aerosol optical model, a pathway not explored in our analysis, these aerosol DRE errors are likely upper bounds.

While the aerosol DRE kernels were derived on a 3-hourly, $2.5^\circ$ grid, lacking in this study is a quantification of the impact of spatial and temporal sampling. Such an undertaking warrants its own dedicated study and may be better served by determining errors from modeling at higher spatial resolutions than MERRA-2 combined with a more advance representation of the subgrid variability (e.g., Wind et al. 2013, 2016; Norris and da Silva 2016). Previous work has shown that sampling aerosol optical depth (AOD) using a single-pixel along-track satellite instrument causes only small biases in global monthly means (Geogdzhayev et al. 2013, 2014), although more substantial errors can be found in regional means (Colarco et al. 2014). Several studies have shown that the diurnal sampling of polar-orbiting satellites (i.e., a single daytime and a single nighttime observation) do not significantly impact the computation of diurnally averaged aerosol DRE or AOD (Kaufman et al. 2000; Arola et al. 2013; Kassianov et al. 2013).

The relative importance of the aerosol DRE and the average kernels in various scene types was examined since retrieval techniques and their accuracies vary greatly depending on the scene type being observed. This parsing by scene type showed that accurate clear-sky retrievals are the most crucial component of the aerosol DRE. Aerosols in presence of thin cloud are shown to be a more radiatively important component of the all-sky aerosol DRE than those with optically thick cloud. While the latter situation has inspired numerous specialized remote sensing techniques (Hu 2007; Torres et al. 2007; Waquet et al. 2009, 2013; de Graaf et al. 2012; Jethva et al. 2013; Liu et al. 2015; Meyer et al. 2015), retrieving aerosol properties in the presence of thin cloud is considerably more difficult despite their relevance to the aerosol DRE.

This study builds a framework for assessing aerosol DRE uncertainty. The aerosol DRE kernels in section 4 can be multiplied by uncertainties in midvisible (550 nm) aerosol scattering properties, surface albedo, and cloud properties to assess the impact on the aerosol DRE. The impact of error covariances on the aerosol DRE could also be quantified using the kernels. To this, an assessment of systematic uncertainties would then add the bias contributions from incomplete knowledge of spectral shape (section 5b) and, for a passive retrieval, the biases from vertical simplifications (section 5a). This general formulation can help in reconciling the large range in satellite-based estimates of the aerosol DRE (Yu et al. 2006; Henderson et al. 2013; Oikawa et al. 2013; Matus et al. 2015) and better design future observing systems to reduce these current observational uncertainties. Both these applications are being actively pursued. Understanding and quantifying the aerosol DRE globally is a prerequisite for placing observational constraints on aerosol radiative forcing (i.e., the radiative effect of anthropogenic aerosols) and the more complex influences of aerosol–cloud interactions on the climate system.

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APPENDIX

Radiative Transfer Model Inputs

An overview of the MERRA-2 variables and corresponding data collections used as input into the NFLRTM is given in Table 1. All inputs are regressed
from the native MERRA-2 grid of $\frac{5}{6}^\circ$ by $\frac{1}{2}^\circ$ (longitude by latitude) to 2.5° by 2.5°. The MERRA-2 data collections in Table 1 are provided as either instantaneous or time-averaged values that are combined to create 3-hourly inputs. To accomplish this, the time-averaged collections (“tavg1” and “tavg3”) are linearly interpolated to center the averages at the 3-hourly instantaneous times (“inst3”). Therefore, the inputs used are slightly mismatched in time. The MERRA-2 visible surface albedo (ALBVISDR and ALBVISDF) is used for the first 10 NFLRTM bands (Table 3; from 175 to 690 nm) and the near-infrared surface albedo (ALBNIRDR and ALBNIRDF) for bands 11–18 (Table 3; 690 nm–4 μm). The direct (diffuse) surface albedos are used in the clear-sky (cloudy sky) calculations. The TOA solar irradiance is nominally set to 1360.0 W m$^{-2}$, adjusted for the daily-mean Earth–sun distance, and scaled by the day fraction—the fraction of time during which the sun is above the horizon at each grid point and in each 3-hourly window. The solar zenith angle is set to the mean value that occurs in each grid box and each 3-hourly window. Trace gases (carbon dioxide and methane) are set to 2007 global mean values from the NOAA Earth System Research Laboratory (ESRL) Global Monitoring Division data (Conway et al. 1994; Dlugokencky et al. 2009).

Radiative fluxes are computed over seven cloudy subcolumns in each grid box by randomly overlapping the MERRA-2 cloud “super layers” of high (HGH; above about 400 hPa), middle (MID; from about 400–700 hPa), and low (LOW; below about 700 hPa) cloud (Chou et al. 1996). The cloud area fraction and total optical depth in these super layers are provided by MERRA-2 (Table 1). The cloud phase of each super layer is determined by integrating the profiles of cloud liquid and ice layer optical depth profiles (TAUCLIW and TAUCLI, respectively). These liquid/ice optical depth profiles are combined with their respective mass mixing ratio profiles (INCLOUDQL and INCLOUDQI) to determine the mean liquid effective radius (Minnis et al. 1998) and ice generalized effective diameter (Fu 1996) in each super layer. The cloud-top and cloud-base pressure in each super layer is computed from the cloud fraction profile (CLOUD) using the weighted average of the cloud tops (bases) exposed to space (the surface) (e.g., Stubenrauch et al. 1997). The all-sky flux is determined from the weighted sum of the clear and cloudy sub-column fluxes.

For aerosols, the MERRA-2 aerosol optical property lookup tables (Randles et al. 2017; Meng et al. 2010; Chin et al. 2002; Gerber 1985; Hess et al. 1998; Colarco et al. 2014) are used along with the mass mixing ratios of each aerosol species (Table 1) to obtain profiles of $\alpha$, $\omega$, and $g$. The scattering properties are interpolated to the center wavelengths of each NFLRTM band (Table 3). The MERRA-2 relative humidity (RH) is used for the interpolation of hydrophilic species. Aerosol are assumed to be externally mixed, so the aerosol scattering properties in a layer are computed as

$$\alpha = \sum_i \alpha_i,$$ \hspace{1cm} (A1)

$$\omega = \frac{\sum_i \omega_i}{\sum_i \alpha_i}, \quad \text{and} \quad (A2)$$

$$g = \frac{\sum_i \omega_i g_i}{\sum_i \omega_i},$$ \hspace{1cm} (A3)

where $i$ are scattering properties of the individual species (i.e., dust, sea salt, hydrophobic and hydrophilic black and organic carbon, and sulfate).

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