The Semidirect Effect of Combined Dust and Sea Salt Aerosols in a Multimodel Analysis

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Abstract To date, very few studies have focused on dust and sea salt cloud interactions, particularly the semidirect effect (SDE) that results from changes in column temperature and moisture. Here, we isolate the SDE using several climate models driven by semiempirical dust and sea salt direct radiative effects. The global annual mean SDE varies from 0.01 to 0.10 W/m², with the bulk of the signal coming from an increase in shortwave radiation. This is consistent with decreases in low cloud over ocean due to cloud burn-off and reductions in middle level cloud due to atmospheric stabilization and decreased convection. Overall, longwave effects weaken the positive SDE but with opposing effects over land and sea. High cloud is reduced over land but enhanced over sea. We conclude that dust and sea salt likely exert a global mean warming effect through cloud rapid adjustments.

1. Introduction

Aerosols represent one of the largest sources of uncertainty in the estimation of past and future climate change. Much of this uncertainty stems from aerosol-cloud interactions (Boucher et al., 2013), which include cloud microphysical effects (changes in cloud albedo and lifetime), as well as aerosol-driven changes to the environment in which clouds form. Such rapid adjustments are particularly important for absorbing aerosols such as black carbon (BC) and dust, which impact clouds by heating the atmosphere and altering the tropospheric temperature and moisture structure (Hansen et al., 1997). Available semidirect effect (SDE) estimates due to BC suggest a large range of $-0.44$ to $0.1$ W/m² (Bond et al., 2013). Thus, the sign of the SDE is not well known, implying a low level of scientific understanding.

Previous studies have described conditions under which absorbing aerosols either increase or decrease cloud cover. This effect depends on several factors including the altitude of the absorbing aerosols relative to the cloud and cloud type (Koch et al., 2009). Modeling studies show that when absorbing aerosols reside above the cloud top, they can stabilize the underlying layer, enhancing stratocumulus clouds (Allen & Sherwood, 2010; Johnson et al., 2004; Koch & Genio, 2010). However, when absorbing aerosols and clouds are colocated, cloud burn-off occurs (Ackerman et al., 2000). Bollasina et al. (2008) showed high May aerosol loading in South Asia leads to reduced cloud amount, increased surface shortwave radiation, and land surface warming, as well as a stronger summer monsoon. Recent modeling studies show that a tenfold increase in BC leads to a significant increase in global annual mean low-level cloud but reductions in midlevel and in particular high-level cloud, leading to a negative aerosol-cloud SDE (Stjern et al., 2017, 2018) dominated by enhanced longwave cooling (Allen et al., 2019). However, Allen et al. (2019) find a robust positive SDE using multiple models driven by semiempirical fine-mode aerosols without dust and sea salt. This adjustment is consistent with decreases in low and midlevel clouds, leading to an increase in shortwave radiation. They suggest that the conflicting SDE and cloud adjustment is due to an aerosol atmospheric heating profile that is too vertically uniform in freely running aerosol simulations.

Studies have showed that dust can exert a SDE. For example, Amiri-Farahani et al. (2017) use observations to quantify the radiative effect of Saharan dust on North Atlantic marine stratocumulus (MSc). They infer that Saharan dust-MSc interactions off the coast of northern Africa are likely dominated by the SDE, particularly during boreal summer. Modeling studies yield conflicting results. For example, Perlwitz and Miller (2010) find a significant increase in low cloud cover with increasing absorptivity of soil dust particles in regions with high dust load (except during Northern Hemisphere winter). Miller et al. (2004) found that low cloud cover and precipitation increase in response to dust radiative forcing over the western Sahara desert. These
increases in low cloud imply a negative SDE. However, Tegen and Heinold (2018) find a positive dust-cloud SDE of 0.12 W/m$^2$. Heating by dust may also impact convection (and the associated clouds), as Stephens et al. (2004) show lofted African dust yields low-level convergent flow toward the dust region and enhanced convection. The adjacent, nondusty regions experienced reduced convection.

There are two approaches to understanding aerosol impacts on the climate system. Bottom-up approaches use aerosol and precursor emissions to characterize the physical properties of the aerosol. Optical and cloud active properties can then be inferred and used to quantify aerosol radiative forcing using a global model. Top-down approaches involve prescribing aerosol optical properties based on a combination of observations and models. Traditional bottom-up aerosol simulations are subject to many uncertainties, including emissions, transport, vertical distribution, and removal (Allen & Landuyt, 2014; Bond et al., 2013; Textor et al., 2006). In the context of dust and sea salt, absolute aerosol optical depth (AOD) and the fraction of total AOD contributed by dust and sea salt vary by more than a factor of 2 between models (Shindell et al., 2013; Textor et al., 2006). Furthermore, most models underestimate African dust emission, transport, and optical depth (Evan et al., 2014). In an attempt to circumvent some of these uncertainties, we use the top-down approach, involving prescribing aerosol optical properties based on a combination of observations and models. Specifically, we prescribe a monthly varying climatology of semiempirical dust and sea salt direct radiative effects (DREs) (Chung et al., 2016) into multiple atmospheric general circulation models to quantify the SDE. Our simulations only account for aerosol direct effects and SDEs and do not include cloud microphysical (indirect) effects. Although our aerosol forcing includes sea salt, it likely has negligible contribution to the SDE since it is not an absorbing aerosol and therefore does not directly perturb the atmospheric temperature and water vapor profile. Moreover, our experimental design precludes cooling of the sea surface, which is where most of the sea salt resides. Similar to Allen et al. (2019), this study quantifies the cloud rapid adjustment to semiempirical aerosol forcing. As opposed to fine-mode aerosols without dust and salt, the present study focuses on dust and sea salt. A description of our data set and methodology is provided in section 2. Results are presented in section 3, and conclusions are presented in section 4.

2. Materials and Methods
2.1. Semiempirical Dust and Sea Salt DREs

We use semiempirical aerosol DRE data from Chung et al. (2016), which is based on the approach of Chung et al. (2005) and Lee and Chung (2013). Indirect aerosol (cloud microphysical) effects are not included. Satellite total AOD data from 2001–2010 are nudged toward AERosol ROBotic NETwork (AERONET) AOD. Aerosol optical properties (e.g., the single-scattering albedo) for the total aerosol are derived by nudging Goddard Chemistry Aerosol Radiation and Transport (GOCART) simulated values toward the AERONET data. Aerosol vertical profiles are obtained from the space-borne Cloud-Aerosol Lidar with Orthogonal Polarization (Liu et al., 2009; Winker et al., 2013). The 2001–2010 average top-of-the-atmosphere (TOA) aerosol DRE for each month is computed by incorporating the integrated global aerosol data into the Monte Carlo Aerosol Cloud Radiation (MACR) model (Choi & Chung, 2014; Podgorny et al., 2000).

A similar procedure is used to estimate fine and coarse-mode aerosol DRE. Although observations alone cannot be used to globally derive individual aerosol species’ AOD, fine-mode AOD (fAOD) is obtained by using AERONET fAOD and total AOD to derive the fine-mode fraction (FMF). AOD Ångström exponent data are derived by adjusting the satellite data toward AERONET data and then converted into FMF data, which are nudged toward AERONET FMF data to derive reliable FMF and thus fAOD over the globe. FMF over land is difficult to retrieve from satellites, and past semiempirical estimates have only used FMF from satellites over the ocean (Bellouin et al., 2008; Myhre, 2009). AERONET data, however, provide relatively reliable FMF over both land and ocean (mostly land). Nudging the satellite data toward AERONET allows a global FMF and, thus, fine-mode AOD estimate. Coarse-mode AOD (cAOD) is obtained by subtracting fAOD from the total AOD. Fine-mode dust and sea salt AOD is obtained by using observed cAOD multiplied by FMF$_S$/(1 − FMF$_S$), where FMF$_S$ is simulated FMF of dust and sea salt (supporting information).

The global annual mean 2001–2010 semiempirical dust and salt AOD is 0.097. Two global aerosol reanalyses, the Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA2; Randles et al., 2017) and the Monitoring Atmospheric Composition and Climate (MACC) Inness et al., 2013), yield corresponding values of 0.071 and 0.100, respectively (supporting information Figure S1). Both reanalyses
use an aerosol module (e.g., GOCART in the case of MERRA2) and assimilate AOD from various ground and satellite AOD observations. The average of 12 AeroCom Phase II models for the year 2006 yield 0.071 ± 0.033 (one-standard deviation uncertainty). Over Africa only, the semiempirical dust and sea salt AOD is 0.217. MERRA2 and MACC yield corresponding values of 0.143 and 0.135. Thus, our semiempirical dust and sea salt AOD is a bit larger than these other estimates, particularly over Africa.

Figure 1 shows the annual mean atmospheric heating ($F_{ATM}$) and reduction in surface radiation ($F_{SFC}$) for semiempirical dust and sea salt aerosols as estimated by Chung et al. (2016) using the MACR radiation model. This forcing is subsequently inserted into our atmospheric general circulation models (section 2.3) to estimate the cloud rapid adjustment. The global average atmospheric heating ($F_{ATM}$) and reduction in surface radiation ($F_{SFC}$) is 1.12 and $-3.3$ W/m$^2$, respectively ($-2.18$ W/m$^2$ TOA DRE). Based on AeroCom Phase II simulations for 2006, the TOA dust radiative effect in the three models that archived the relevant diagnostic is $-0.81$ W/m$^2$ in HadGEM2-ES, $-0.51$ W/m$^2$ in INCA, and $+0.02$ W/m$^2$ in CAM5.1 MAM3 PNNL. In seven models that archived the TOA radiative effect of sea salt, values range from $-0.18$ to $-1.23$ W/m$^2$ in SPRINTARS-v384 and CAM5.1 MAM3 PNNL, respectively. Thus, there is considerable uncertainty in simulated dust and sea salt TOA radiative effects.

Three AeroCom models archived both dust and sea salt TOA radiative effects, with INCA yielding $-0.91$ W/m$^2$, CAM5.1 MAM3 PNNL yielding $-1.13$ W/m$^2$, and HadGEM2-ES yielding $-1.63$ W/m$^2$, for a multimodel mean of $-1.22$ W/m$^2$. The corresponding estimate from Chung et al. (2016) of $-2.18$ W/m$^2$ is larger than each model and about 80% larger than the multimodel mean. Two AeroCom models allow quantification of dust and sea salt radiative effects at the surface and in the atmosphere. $F_{SFC}$ ranges from $-2.48$ to
−2.07 W/m², and \( F_{\text{ATM}} \) ranges from 0.85 to 1.16 W/m² in HadGEM2-ES and INCA, respectively. \( F_{\text{SFC}} \) from Chung et al. (2016; −3.3 W/m²) is about 45% larger than these model estimates. \( F_{\text{ATM}} \) from Chung et al. (2016; 1.12 W/m²) is similar to that simulated by INCA.

Figure 1a shows maximum atmospheric heating occurs over North Africa (the world’s largest dust source, the Sahara desert), the Middle East (due to Arabian Desert), and over parts of China (due to the Gobi desert). Within the Gobi desert, there are two maxima—the western maximum is coincident with the Taklamakan desert; the eastern maximum is downwind of the Gobi. AeroCom models also tend to show this eastern maximum in dust AOD (not shown). Also included is the vertical profile of the global mean atmospheric heating rate, which shows most of the heating occurs over land (consistent with the dust source), peaking near 800 hPa but with significant heating throughout the midtroposphere. Over the sea, however, most of the heating occurs near the surface (~925 hPa) and then rapidly decays to 0 near 500 hPa. Similar heating profiles exist over Africa (Figure 1d) but with larger magnitude.

2.2. Global Climate Models

This study uses the National Center for Atmospheric Research (NCAR) Community Atmosphere Model version 4 (CAM4; Neale, Richter, et al., 2010) and version 5 (CAM5; Neale, Gettelman, et al., 2010). Both models have a horizontal resolution of 1.9° × 2.5°; CAM4 has 25 vertical layers and CAM5 has 30. They share the same Zhang-McFarlane deep convection scheme (bulk mass flux with CAPE closure; Zhang & McFarlane, 1995). CAM4 uses a shallow convection scheme that involves three-level adjustment of moist static energy (Hack, 1994) and a prognostic single-moment microphysics scheme, including diagnostic cloud fraction (Rasch & Kristjánsson, 1998). Cloud fraction depends on several factors, including relative humidity (RH), lower-tropospheric stability (S), water vapor, and convective mass fluxes. Three types of cloud are diagnosed: low-level marine stratocumulus, convective cloud, and stratus (layered cloud). Layered clouds form when RH exceeds a pressure dependent threshold. Marine stratocumulus clouds are diagnosed using an empirical relationship based on lower-tropospheric stability (S). Convective cloud fraction is related to updraft mass flux in the deep and shallow cumulus schemes. The remaining cloud types are diagnosed on the basis of relative humidity (Neale, Richter, et al., 2010). The calculation of shortwave radiation in CAM4 is based on a \( \delta \)-Eddington approximation (Briegleb, 1992; Coakley et al., 1983; Joseph et al., 1976).

CAM5 uses a mass flux scheme with convective inhibition closure for shallow convection (Park & Bretherton, 2009) and a prognostic double moment microphysics scheme (Morrison & Gettelman, 2008) with ice supersaturation (Gettelman et al., 2010) and a diagnostic cloud fraction scheme for cloud microphysics and macrophysics. Although deep cumulus cloud fraction is diagnosed as in CAM4, shallow cumulus fraction in CAM5 is directly computed using the definition of convective updraft mass flux from the new shallow convection scheme (Park & Bretherton, 2009). Liquid stratus fraction is derived from a triangular distribution of total relative humidity. The rapid radiative transfer model (RRTMG) provides the radiative transfer calculations in CAM5, which is an accelerated and modified version of the correlated k-distribution model, RRTM (Clough et al., 2005; Iacono et al., 2008; Mlawer et al., 1997). Although CAM4 and CAM5 share several similar features, they do have several differences (Neale, Richter, et al., 2010; Neale, Gettelman, et al., 2010) that warrant inclusion of both models in our analysis.

We also use the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model version 2.1 (AM2.1) (Anderson et al., 2004). It has a horizontal resolution of 2° × 2.5° and has 24 vertical layers. Moist convection is represented by the Relaxed Arakawa-Schubert formulation (Moorthi & Suarez, 1992). Large-scale clouds are parameterized with separate prognostic variables for specific humidity of cloud liquid and ice, with an updated treatment of mixed phase clouds (Rotstayn, 1997; Rotstayn et al., 2000). Stratocumulus cloud cover is based on large-scale subsidence, diabatic cooling by radiation, and turbulent entrainment of warm and dry air from above the inversion (Tiedtke, 1993). The \( \delta \)-Eddington method is used to solve for the layer reflection and transmission, while the thick-averaging method is used to combine layers.

Low-, middle-, and high-level clouds are directly calculated in each model. CAM4/5 assumes a maximum-random overlap, with low clouds based on pressure levels from the surface pressure to 700 hPa; midlevel clouds are based on 700 to 400 hPa; and high-level clouds are based on 400 hPa to the model top. For GFDL AM2.1, clouds are assumed to randomly overlap. Low clouds are calculated over 1,000–680 hPa; midlevel clouds from 680–440 hPa and high clouds from 440 to 10 hPa.
Table 1
Global Annual Mean Cloud Adjustments to Semiempirical Dust and Sea Salt Aerosol Direct Radiative Effects

| Model | $C_{\text{BOT}}$ | CLow | CMED | CHI |
|-------|-----------------|------|------|-----|
|       | Total | Land | Sea | Total | Land | Sea | Total | Land | Sea | Total | Land | Sea |
| CAM4  | 0.06 | 0.20 | −0.001 | −0.03 | 0.17 | −0.11 | −0.05 | −0.24 | 0.03 | −0.06 | −0.22 | 0.01 |
| CAM5  | 0.04 | 0.07 | 0.03 | −0.04 | −0.01 | −0.05 | −0.03 | −0.34 | 0.10 | 0.04 | −0.25 | 0.15 |
| GFDL  | 0.07 | 0.14 | 0.04 | −0.05 | −0.02 | −0.06 | −0.04 | −0.33 | 0.08 | −0.01 | −0.48 | 0.18 |

Note. Cloud changes are shown for low-level (CLOW), midlevel (CMED), and high-level (CHI) cloud, as well as at the model’s bottom level ($C_{\text{BOT}}$). Also included are land and ocean adjustments. Changes significant at the 90% confidence level are denoted by bold font based on a t test for the difference of means using the pooled variance. Unit is percent.

2.3. Experimental Design

We conduct simulations with fixed sea surface temperature, which is based on a repeating cycle of monthly climatological sea surface temperatures. Significant global mean temperature change is prohibited by using this setup, and only rapid adjustments like the SDE are considered. Simulations are run for 40 years, and the last 35 years of each simulation are analyzed in this study. The first 5 years are therefore discarded as model spin-up, although this is probably overly conservative. Monthly semiempirical dust and sea salt (atmospheric heating rate and surface solar radiation reduction, Figure 1) are interpolated to each model’s horizontal resolution and incorporated into their radiation modules. The atmospheric heating rate is vertically interpolated to each model’s hybrid pressure levels. Although aerosol forcing is almost independent of solar zenith angle ($\theta$) when the angle is small, aerosol forcing approaches 0 as $\theta$ approaches 90°. Thus, the added aerosol DRE is multiplied by a scaling factor that depends on zenith angle (Allen & Sherwood, 2010; Chung, 2006). The rapid adjustment is estimated as the difference between the simulation with semiempirical dust and sea salt (the “experiment”), and a corresponding simulation without semiempirical dust and sea salt aerosols (the “control”).

In CAM5, cloud microphysical processes depend upon aerosols, so species in addition to dust and sea salt must be included (e.g., sulfates and carbonaceous aerosols). Thus, CAM5 simulations include prescribed modal aerosols. Although this represents a double counting of dust and sea salt effects in the semiempirical CAM5 simulation (the experiment), we have verified that subtracting the experiment and control eliminates this double counting of the default prescribed modal aerosols (which will be similar in semiempirical and control simulations). For example, the percent change in the burden of BC is 0.02%, 0.009% for dust, 0.009% for primary organic matter, 0.04% for sea salt, and 0.003% for sulfate and secondary organic aerosol. The reason why these changes are not identical to 0 is because the prescribed CAM5 model aerosol implementation does not use mixing ratio values that have been time interpolated from monthly mean values. Instead, the mixing ratio values are obtained by random sampling of the time interpolated log normal distribution of each prescribed species.

3. Results

3.1. Cloud Adjustment

Table 1 shows global annual mean changes in low-level (CLOW), midlevel (CMED), and high-level (CHI) cloud, as well as cloud changes at the model’s bottom level ($C_{\text{BOT}}$). $C_{\text{BOT}}$ increases in all models, particularly over land. In contrast to $C_{\text{BOT}}$, all models yield global annual mean reductions in both CLOW and CMED, ranging from $-0.05\%$ to $-0.03\%$. Most of the CLOW decrease occurs over ocean, where all models show significant low-cloud reductions of $-0.05\%$ to $-0.11\%$. A land-sea contrast exists for CMED, with an increase in midlevel clouds over ocean, but with larger reductions over land ($-0.24\%$ to $-0.34\%$). A similar land-sea contrast exists for high-level clouds, with significant reductions over land ($-0.22\%$ to $-0.48\%$). The CMED and CHI reductions are largest over Africa (supporting information Figure S2).

Figure 2 shows the global annual mean vertical profiles of the total (CLD), convective (CONCLD), and stratus (STRCLD) cloud adjustment over land and ocean, along with relative humidity (RH) and two measures of vertical motion—convective mass flux (CMF) and pressure vertical velocity ($\Omega$). At all pressure levels, CONCLD decreases over land, with weak increases over ocean. These adjustments are consistent with the corresponding changes in CMF. The total cloud adjustment is very similar to the stratus cloud adjustment, both of which are consistent with the change in RH. Over land, the total cloud adjustment includes...
lower-tropospheric increases in RH and cloud, and decreases above \( \sim 800 \) hPa. The increase in cloud near the surface (e.g., \( C_{\text{bot}} \)) over land is related to increases in lower-tropospheric stability (S; supporting information Figure S3) due to larger surface cooling relative to 700 hPa. This traps moisture near the surface, leading to corresponding increases in \( C_{\text{bot}} \) and relative humidity (RH\text{\textsubscript{bot}}), particularly over land (supporting information Figure 3). The middle- and upper-tropospheric RH decreases over land are consistent with reduced ascent (Figures 2c and 2d). Here, the stabilizing effect of dust is very large, which inhibits vertical motion (supporting information Figure S3) and the formation of middle- and high-level cloud. The midtropospheric RH decrease is also consistent with the location of the heating over land (Figure 1c).

Over the sea, both RH and cloud decrease in the lower troposphere, and increase in the middle and upper troposphere. The lower-tropospheric decreases are consistent with the location of the heating (Figure 1c), most of which is located near 900 hPa; the middle- and upper-tropospheric increases are consistent with enhanced ascent. A more detailed regression analysis (supporting information Table S1 and Figure S4) shows that RH is generally the best predictor of the cloud adjustment, with S also important for low cloud, and vertical motion (\( \Omega \)) also important for middle and high cloud.

Although models agree on the global annual mean reduction in low and midlevel cloud, limited regional agreement exists (supporting information Figure S2). For middle- and high-level clouds, there are more regions of cloud adjustment model agreement, particularly over Africa where CMED and CHI reductions occur.

### 3.2. SDE

For this study we use semiempirical dust and sea salt DRE for surface solar reduction and atmospheric solar heating rate. Thus, there are no indirect effects and it is possible to estimate the aerosol-cloud SDE from the cloud radiative flux, calculated as the difference between top-of-atmosphere (TOA) net all-sky and clear sky.
Table 2
Global Annual Mean Total Rapid Adjustments, Including Cloud (SDE) and Noncloud Adjustments to Semiempirical Dust and Sea Salt Aerosol Direct Radiative Effects

| Model  | Total  | Land  | Sea  | Total  | Land  | Sea  | Total  | Land  | Sea  |
|--------|--------|-------|------|--------|-------|------|--------|-------|------|
|        | RAP_ADJ |       |      | RAP_ADJ |       |      | RAP_ADJ |       |      |
| CAM4   | 0.11    | 0.09  | 0.12 | 0.06    | 0.03  | 0.07 | 0.05    | 0.06  | 0.05 |
| CAM5   | 0.12    | 0.14  | 0.12 | 0.13    | 0.35  | 0.04 | −0.01   | −0.22 | 0.08 |
| GFDL   | 0.13    | 0.21  | 0.11 | −0.04   | 0.12  | −0.11| 0.18    | 0.09  | 0.21 |
|        | SDE     |       |      | SDE_SW |       |      | SDE_LW |       |      |
| CAM4   | 0.01    | −0.19 | 0.09 | 0.09    | 0.15  | 0.06 | −0.08   | −0.35 | 0.03 |
| CAM5   | 0.07    | −0.01 | 0.11 | 0.15    | 0.44  | 0.04 | −0.08   | −0.44 | 0.07 |
| GFDL   | 0.10    | 0.03  | 0.14 | 0.18    | 0.64  | −0.01| −0.08   | −0.61 | 0.14 |
|        | RES     |       |      | RES_SW |       |      | RES_LW |       |      |
| CAM4   | 0.10    | 0.28  | 0.03 | −0.03   | −0.12 | 0.01 | 0.13    | 0.41  | 0.02 |
| CAM5   | 0.05    | 0.15  | 0.01 | −0.02   | −0.08 | 0.001| 0.07    | 0.23  | 0.01 |
| GFDL   | 0.03    | 0.18  | −0.03| −0.22   | −0.52 | −0.10| 0.25    | 0.70  | 0.07 |

Note. Rapid adjustments (RAP_ADJ) are calculated as the effective radiative forcing (ERF) minus the instantaneous direct forcing (−2.18 W/m²). ERF is calculated as the difference in TOA radiative fluxes. Quantities are decomposed into shortwave (e.g., SDE_SW) and longwave (e.g., SDE_LW) components. Noncloud adjustments are calculated as the residual (RES = RAP_ADJ − SDE). Also included are land versus ocean adjustments. Changes significant at the 90% confidence level are denoted by bold font based on a t test for the difference of means using the pooled variance. Unit is watts per square meter. SDE = semidirect effect; TOA = top-of-the-atmosphere.

Table 2 shows that all three models yield a positive global annual mean SDE. The largest SDE occurs in GFDL AM2.1 at 0.10 W/m², followed by CAM5 and CAM4 at 0.07 and 0.01 W/m², respectively. Both GFDL AM2.1 and CAM5 yield a significant SDE. Furthermore, the positive SDE is due to shortwave radiation, with all three models yielding a significant SDE_SW increase. This is consistent with the reduction in midlevel clouds over land discussed above, with low-cloud reductions over the sea also contributing (but partially offset by the corresponding CMED increase over sea). Low clouds have a net cooling effect on the planet, because they tend to have a high albedo. Although midlevel clouds are less reflective, and possess a greenhouse effect, we find that they affect the SDE in a way similar to low-level clouds. Thus, reductions in low and midlevel clouds lead to an increase in solar radiation, a positive SDE, and warming. Furthermore, the small CAM4 SDE is directly related to its relatively large CLOW increase over land (Table 1).

The SDE due to longwave radiation (SDE_LW) is negative in all three models, thereby masking the positive SDE_SW. The negative SDE_LW is consistent with high-cloud reductions over land, which is partially offset by the CHI increase over oceans. High level clouds tend to have a warming effect, due to their greenhouse effect. Thus, reductions in high clouds cause a negative SDE, due to enhanced longwave cooling. Interestingly, due to opposing CMED and CHI TOA flux adjustments over land, the positive SDE_SW due to CMED reductions and negative SDE_LW due to CHI reductions mute one another (except in CAM4), which means the overall SDE is driven by changes over the ocean. Here, although both SDE_SW and SDE_LW are relatively small, they are both positive (due to CLOW reductions and CHI increases respectively) and lead to a significant positive SDE over ocean in all models. Similar to the limited spatial agreement in the cloud adjustment across models, limited spatial SDE agreement exists (supporting information Figure S5).

We note that the total rapid adjustment (RAP_ADJ) is positive in all models (Table 2), with the SDE (cloud rapid adjustment) dominating in two of three models (CAM4 being the exception). Noncloud rapid adjustments (calculated as the residual, RES = RAP_ADJ − SDE) are also positive in all models. This is due to a positive longwave noncloud adjustment, consistent with atmospheric cooling and less longwave emission. This is partially offset by a negative shortwave noncloud adjustment, consistent with decreases in water vapor and less solar absorption.
4. Conclusion and Discussion

We investigate the SDE using semiempirical dust and sea salt DREs. We find a robust positive global annual mean total rapid adjustment in all three models, which is generally dominated by the cloud rapid adjustment (ranging from 0.01 W/m² in CAM4 to 0.10 W/m² in GFDL AM2.1). The positive SDE is dominated by shortwave radiation, consistent with decreases in low and midlevel clouds (over sea and land, respectively) which result in an increase in shortwave radiation and warming. Longwave effects overall weaken the positive SDE, but with opposing effects over land and sea—high-level cloud is reduced over land but enhanced over sea. Due to opposing CMED and CHI TOA flux adjustments over land, the positive SDE is driven by cloud changes—including reduced CLOW and increased CHI—over the ocean. Cloud changes are primarily related to changes in relative humidity, stability, and vertical motion. These results are qualitatively consistent with Allen et al. (2019), where fine-mode aerosols without dust and sea salt led to a positive SDE dominated by shortwave radiation, consistent with low and midlevel cloud reductions. Although we use semiempirical dust and sea salt direct forcing in an attempt to reduce model uncertainties related to aerosol emissions, transport, vertical distribution, and removal (Allen & Landuyt, 2014; Bond et al., 2013; Evan et al., 2014; Shindell et al., 2013; Textor et al., 2006), our approach may be limited by several uncertainties. This includes the observed AOD and reliance on simulated optical properties, as well as uncertainty related to the discrimination between fine- and coarse-mode aerosols. Additional caveats include lack of consistency between the aerosol forcing and simulated meteorology, as well as uncertainty in the simulated cloud fields. For example, Randles et al. (2013) found that removing the feedback of meteorology on aerosol distributions can significantly impact the response depending on the parameter, region, and season considered. The largest effect of removing coupling is to enhance the AOD globally over the oceans. Any differences in the SDE across our models is not due to the aerosol forcing but due to the model’s adjustment to this fixed forcing (e.g., due to cloud or convection parameterizations). Although we find several robust adjustments on the global scale, including positive total and cloud rapid adjustments, there is little regional agreement (e.g., supporting information Figure S5). This underscores the difficulty of quantifying the SDE and implies that a model’s circulation and cloud adjustment leads to significant uncertainty in the SDE, independent of the aerosol forcing. Our results also imply that a multimodel analysis using each model’s prognostic aerosols will likely result in a larger SDE uncertainty than showed here. Nonetheless, our results show that dust and sea salt leads to a positive global mean SDE, implying cloud rapid adjustments act to warm the climate system.

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