Sensitivity of climate forcing and response to dust optical properties in an idealized model

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An idealized global climate model is used to explore the response of the climate to a wide range of dust radiative properties and dust layer heights. The top-of-the-atmosphere (TOA) shortwave forcing becomes more negative as the broadband shortwave single scattering albedo increases and the broadband shortwave asymmetry parameter decreases, but the sensitivity is highly dependent on the location of the dust layer with respect to clouds. The longwave TOA forcing is most affected by the height of the dust layer. The net TOA forcing is most sensitive to the shortwave single scattering albedo and shortwave asymmetry parameter. The surface and atmospheric temperature responses are approximately linear with respect to the TOA forcing, as opposed to the surface or atmospheric forcings. Thus the TOA forcing can be used to estimate both the surface and atmospheric temperature responses to dust. The corresponding changes in latent and sensible heat fluxes are essential for the close relationship of the surface temperature response to the TOA forcing. Estimating the hydrological cycle response requires knowledge of the vertical distribution of dust with respect to clouds or other reflective particles. The sensitivity of the latent heat flux to variations in the shortwave single scattering albedo changes sign with dust height. The latent heat flux change becomes less negative as the shortwave single scattering albedo increases if the dust layer is below clouds. However, when the dust is above clouds, the latent heat response becomes more negative as the single scattering albedo increases.


1. Introduction

Airborne mineral, or soil, dust affects the earth’s climate by altering the surface and atmospheric radiative budgets. The radiative effect of airborne dust is difficult to determine because the current dust distribution and optical properties are not well known. Dust remains in the atmosphere for at most a few weeks; thus it is highly regional and episodic. The chemical and physical properties of dust, which determine optical properties, depend on the source region and the history of uplift, transportation, and deposition. Ground or aircraft-based observations can provide specific information about particle concentrations and optical properties [e.g., Holben et al., 2001]. However, since dust properties can vary greatly in time and space, many such measurements are needed to determine the climatology of dust. On the other hand, satellite data [Herman et al., 1997; Husar et al., 1997] provide continual coverage of large areas, but the derived aerosol properties are subject to biases [e.g., Chu et al., 2005] and differ from instrument to instrument [Abdou et al., 2005].

Uncertainties in the optical properties of dust prevent an accurate determination of the direct radiative effect. Reflection of solar radiation by dust increases the planetary albedo, but this shortwave cooling effect is opposed by a shortwave heating due to absorption of solar radiation. The longwave effect is one of heating [Ramaswamy et al., 2001]. However, the magnitude of the longwave, shortwave, and net effects are unknown and depend strongly on the dust optical properties.

One large uncertainty is the reflectivity of dust at solar wavelengths. The reflectivity is characterized by a wavelength-dependent single scattering albedo, \(\omega\), the ratio of incident radiation that is reflected to the amount that is reflected or absorbed. The single scattering albedo depends on the mineral composition, size, and shape of dust particles and thus varies from region to region [Sokolik et al., 1993; Claquin et al., 1999]. Presently, no single model includes all of these components, and much more observational data are necessary before there is sufficient information to properly account for this variation.

Estimates of a representative broadband dust shortwave single scattering albedo in the literature vary widely. Sokolik and Toon [1996, Figure 1] show that the single scattering albedo varies significantly based on the assumed size distribution and mineralogy (as well as the wavelength of the radiation). While they use \(\omega = 0.85\) as a representative value for shortwave wavelengths, many recent measure-
ments suggest almost zero solar absorption ($\omega \sim 0.97$) [Wang et al., 2003; Kaufman et al., 2001; Christopher et al., 2003; Haywood et al., 2003; Clarke et al., 2004]. Additionally, the shortwave forcing depends on the specifics of the atmospheric column, i.e., the location of clouds or other atmospheric constituents which interact with solar radiation. Similarly, the longwave effect of dust depends on the longwave optical properties and the atmospheric state (temperature, humidity, etc.), though this has been less studied. The uncertainties in these optical properties lead to a large uncertainty in the overall effect of dust [see Shell and Somerville, 2007, Table 1].

The sensitivity of dust radiative forcing to particle optical properties and vertical distribution has been explored using single column radiative transfer models [Liao and Seinfeld, 1998; Claquin et al., 1998], global radiative transfer models [Myhre and Stordal, 2001], and general circulation models [Miller et al., 2004]. Computational requirements have limited these previous dust sensitivity studies to consideration of a few specific cases. In addition, most previous sensitivity work has looked at the forcing, not the climate response (an exception to this is the work by Miller et al. [2004], which considers the hydrological cycle response to three different single scattering albedo cases). In this work, we explore the climate sensitivity (both forcing and response) to the direct radiative effect of dust using a computationally efficient model of the climate system [Shell and Somerville, 2007], which allows a more thorough exploration of parameter space. Rather than focusing on a few specific configurations, we are able to simulate hundreds of different steady states in less than a day. This advantage of our idealized model is especially useful for the current dust problem, since dust concentration and optical parameters are not currently well-constrained by data. We take advantage of this speed to determine how sensitive the dust effect is to these dust properties, including the asymmetry parameter, which has not been considered in as much detail as the single scattering albedo and height. We also examine whether or not the dust forcing and climate responses (temperature and hydrological cycle changes) exhibit the same sensitivities to optical properties. Finally, the simplicity of our model allows for an easier understanding of the feedbacks and processes within the dust-climate system. This model property helps us understand the temperature and hydrological cycle response to the imposed dust forcing, as well as interactions between the two.

The simple model [Shell and Somerville, 2007] determines the radiative effect of a given dust distribution on the climate system and is based on the two-layer energy-balance model of Shell and Somerville [2005]. The model produces a realistic climate forcing and response, for both the present-day dust distribution and the volcanic aerosol distribution for the Mount Pinatubo eruption in 1991. In this paper, we perform a number of sensitivity studies to analyze the climate sensitivity to dust properties. Our goal is to identify the areas of largest uncertainty in the direct radiative effect of dust on climate.

2. Model Description

The simple model consists of a longitudinally averaged atmosphere layer above a surface layer, which represents the combined influences of ocean and land. Both layers are composed of the same number of grid points (generally 100) from pole to pole, so that different latitude bands are resolved. The model determines the steady state temperatures of the atmosphere ($T_a$) and surface ($T_s$) for each latitude such that the different heating and cooling mechanisms balance:

$$\frac{C_a}{\partial T_a}{\partial t} = 0 = S_a + L_a + F_s + F_I + D_a$$

$$\frac{C_s}{\partial T_s}{\partial t} = 0 = S_s + L_s - F_s - F_I + D_s$$

where $C$ is the heat capacity, $t$ is time, $S$ is the net solar (shortwave) heating, $I$ is the net infrared (longwave) heating, $F_s$ and $F_I$ are the sensible and latent heat fluxes from the surface to the atmosphere, and $D$ represents the meridional transport of heat due to dynamical effects. The subscripts $a$ and $s$ refer to atmospheric terms and surface terms respectively. Although the model includes a single atmospheric layer, it does approximate the variation of temperature with height by explicitly calculating a lapse rate. In the tropics, the lapse rate is set to the moist adiabatic lapse rate, based on a constant relative humidity in the boundary layer. At higher latitudes, the lapse rate is set to the critical lapse rate for baroclinic adjustment [Stone, 1978].

Dust influences the shortwave and longwave heating terms in equations (1) and (2). These modifications are described in detail by Shell and Somerville [2007] and summarized here. The dust distribution is specified and constant in time but varies latitudinally based on the present-day annual average dust concentration of Tegen and Lacis [1996] and Tegen et al. [1997]. This concentration is calculated using a tracer transport model [Tegen and Fung, 1994] of the uplift, transport, and deposition of dust and corresponds to a source strength of 1200 Tg/yr and a mean load of 36 mg m$^{-2}$ (18 Tg). In comparison, a compilation of recent dust statistics by Zender et al. [2004] gives an emissions range of 1000–2150 Tg/yr and a load range of 8–36 Tg.

The shortwave parameterization calculates the net solar absorption by the atmosphere and surface, allowing for multiple reflections between the two, based on a latitudinally varying surface albedo and atmospheric absorptivity, reflectivity, and transmissivity. Dust modifies these atmospheric shortwave optical properties. Using a specified and latitudinally invariant broadband dust single scattering albedo, $\omega$, and asymmetry parameter (mean cosine of the scattering angle), $g$, the dust layer is approximated as a delta-Eddington layer before being combined with atmospheric layers above and below it to obtain the net atmospheric properties used in the solar heating calculation. The dust optical properties are chosen to represent the spectrally integrated effect of dust on solar radiation, so that the resulting broadband shortwave radiative effect corresponds to what would be obtained from calculations over the whole spectral range.

We also specify the vertical location of the dust layer with respect to other reflective atmospheric constituents (e.g., clouds or other aerosols). The vertical coordinate, $\zeta$, is defined...
Figure 1. Change in global average solar radiation absorbed by (top) the climate (surface plus atmosphere), (middle) surface, and (bottom) atmosphere as a function of dust single scattering albedo and (left) asymmetry parameter or (right) vertical coordinate $z$ of the dust layer. See section 2 for the definition and physical interpretation of the vertical coordinate $z$. Negative numbers correspond to an instantaneous shortwave cooling, while positive values indicate that the instantaneous shortwave effect of dust warms the climate. The instantaneous shortwave effect of dust always cools the surface. The shortwave effect of dust is generally a warming of the atmosphere.
as the atmospheric pressure at the height of the dust layer if reflection within the atmosphere were evenly distributed as a function of pressure and the desired fraction of atmospheric reflection occurred above the dust layer. Assuming most of the atmospheric reflection is caused by clouds, high values of \( \zeta \) indicate that the dust layer is below most clouds, while low values of \( \zeta \) correspond to dust layers predominately above clouds. For example, if \( \zeta = 200 \) mbar, then about 25% of the atmospheric reflection occurs above the dust layer. Note that the dust-free shortwave atmospheric properties are specified, so there is no cloud feedback in this model.

[12] The effect of dust on the longwave heating budget is calculated separately from the dust-free longwave heating. The dust-free longwave budget includes emission and absorption from the surface and atmosphere. The atmospheric terms are influenced by an interactive latitudinally varying emissivity, which depends on boundary layer temperature. Thus the longwave calculation includes a simple water vapor feedback parameterization. To determine the longwave heating due to dust, we use the simple model of Markowicz et al. [2003]. This model assumes the effect of dust is in the atmospheric window (8–12 \( \mu m \)) and that no gaseous absorption occurs in the window. The longwave calculations do not include the effects of clouds, which have been shown to decrease longwave forcing [Liao and Seinfeld, 1998]. Thus our model may overestimate the magnitude of longwave forcing for cloudy situations. However, our goal is not to determine the exact dust forcing value but rather to understand how this value depends on the optical properties of dust and the atmospheric column.

[13] In order to maintain the clarity and computational efficiency of the model, we have made a number of simplifications to the climate system [Shell and Somerville, 2007]. For example, dust and climate properties vary longitudinally, yet we use zonal averages. The surface albedo of land is quite different from that of ocean, and thus the effect of dust is not the same in the two regions. The single scattering albedo varies as well. For example, Miller et al. [2004] calculate longitudinal variations of up to 0.02 in the bulk shortwave single scattering albedo based on variations in dust particle size, while Claquin et al. [1999] obtain 550 nm single scattering albedo differences of around 0.05 caused by different mineralogies at the same latitude. Correlations between zonal variations in dust and climate properties, such as surface albedo, are neglected in our model. However, the model is still able to produce a reasonable global dust effect compared with general circulation model (GCM) results and generates similar sensitivities to dust properties [Shell and Somerville, 2007].

3. Sensitivity to Shortwave Radiative Properties

[14] We begin our exploration of dust-climate system behavior by focusing on the uncertainties in instantaneous dust forcing and steady state climate response when we vary the shortwave single scattering albedo, asymmetry parameter, and vertical coordinate \( \zeta \) over a range of values. The dust concentration, longwave single scattering albedo, longwave asymmetry parameter, and longwave height are held constant. Figures 1 and 2 show the instantaneous forcing and climate response from this sensitivity study. In Figures 1 (left) and 2 (left), \( \zeta \) is held constant, while the single scattering albedo and asymmetry parameter vary. Figures 1 (right) and 2 (right) show the results when \( \zeta \) and the single scattering albedo are varied while holding the asymmetry parameter constant. We calculated globally averaged dust forcing and climate response for all combinations of these parameter values; the included plots are representative of the complete results.

3.1. Shortwave Forcing Sensitivity

[15] For the purposes of this study, we define the shortwave dust forcing as the difference in instantaneous shortwave heating between the version of the model with dust and the dust-free version. Thus forcing corresponds to the effect of the total mineral dust, not just the anthropogenic component. The TOA shortwave forcing (Figure 1, top) can be either negative or positive and becomes more sensitive to the single scattering albedo and asymmetry parameter as the layer moves above clouds or other aerosols. When the dust layer is above clouds, most atmospheric (i.e., non-dust) absorption occurs below the dust layer. If the dust is reflective, it reflects sunlight before it can be absorbed by the atmosphere. If the dust is absorptive, it absorbs incoming solar radiation before the atmosphere has a chance to reflect it to space. Thus high dust layers have a larger effect on the shortwave TOA forcing than lower layers do. In addition, higher dust is more likely to have a net TOA warming effect, in agreement with radiative transfer model results [Liao and Seinfeld, 1998; Myhre and Stordal, 2001]. A higher dust layer can absorb reflected sunlight from clouds below it, increasing the likelihood that the TOA forcing is positive.

[16] Miller et al. [2004] use a few general circulation model cases to show that TOA shortwave forcing becomes more negative as the single scattering albedo increases. We confirm that this relationship holds for a wide range of configurations. As the single scattering albedo increases, the TOA shortwave forcing becomes more negative because increased reflection of sunlight by dust reduces the planetary albedo. As the asymmetry parameter increases, the TOA shortwave forcing becomes more positive. A higher \( g \) corresponds to more forward scattering of incoming solar radiation and thus less reflection backward to space. Claquin et al. [1998], Liao and Seinfeld [1998], and Myhre and Stordal [2001] show that shortwave TOA forcing becomes more positive as dust particle size increases, corresponding to smaller shortwave single scattering albedos and larger asymmetry parameters [Tegen and Lacis, 1996], though clouds can alter this result. In our model, we vary the single scattering albedo and asymmetry parameter separately, so we can see that the decrease of single scattering albedo and increase in asymmetry parameter with dust size both contribute to a more positive forcing.

[17] The largest TOA cooling occurs when dust is above most of the other reflective atmospheric particles, the shortwave single scattering albedo is high, and the shortwave asymmetry parameter is low. The greatest warming occurs when the dust is above most of the reflective particles, the single scattering albedo is low, and the asymmetry parameter is high. For each asymmetry parameter, there is a specific single scattering albedo where the height is unimportant. For a \( g \) of 0.75, this corresponds to an \( \omega \) of about 0.9. As \( \omega \) moves away from this value, the TOA...
forcing becomes increasingly more sensitive to height. Note that the location of height-insensitivity does not correspond to a zero forcing.

[18] The surface forcing (Figure 1, middle) is always negative. A dust layer always cools the surface by reducing the total amount of incoming solar radiation which reaches the surface. Liao and Seinfeld [1998] find that the surface forcing is less sensitive to dust height than the TOA forcing. Our model confirms this result over a large range of parameters. The surface shortwave forcing is not very sensitive to height, since it depends on the total amount of solar radiation absorbed or reflected by the atmosphere rather than the details of where this absorption or reflection occurs. As the single scattering albedo increases, the forcing becomes less negative, similar to GCM results [Miller et al., 2004], because absorbed solar radiation does not reach the surface while scattered solar may be scattered forward toward the surface. As the asymmetry parameter increases, the surface forcing becomes less negative. The dust scatters more sunlight forward, toward the surface, rather than backward to space. In general, the surface forcing is larger in magnitude than the TOA forcing. However, due to the different surface and TOA sensitivities to changes in $\omega$, the TOA forcing is actually more negative than the surface forcing for some values of high $\omega$ and high height.

[19] Claquin et al. [1998] find a more negative shortwave surface forcing as dust particle size increases, corresponding to smaller single scattering albedos and larger asymmetry parameters, whereas Liao and Seinfeld [1998] find less negative shortwave forcing. This discrepancy is due to differences in experimental configurations. Claquin et al. [1998] use a constant optical depth, while Liao and Seinfeld [1998] specify a constant column burden as they vary the dust size, so that optical depth decreases as particle size increases. Since we do not vary optical depth in these sensitivity studies, our results correspond more to those of Claquin et al. [1998]. The decrease of single scattering albedo with increased dust size results in a more negative surface forcing, which is partially cancelled by the more positive surface forcing associated with a larger asymmetry parameter.

[20] The atmospheric shortwave forcing (Figure 1, bottom) is positive for most optical property values except for some high $\omega$ cases. As the single scattering albedo increases, more sunlight is reflected, and the forcing becomes more negative, in agreement with limited GCM experiments [Miller et al., 2004]. The atmospheric forcing is not very sensitive to the
Table 1. Percent Change in Dust Forcing or Climate Response Caused by Variations in Different Shortwave Optical Properties*  

<table>
<thead>
<tr>
<th>Variable</th>
<th>Default</th>
<th>$\omega$</th>
<th>$g$</th>
<th>$\zeta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>TOA SW, W/m²</td>
<td>0.65</td>
<td>5.1%</td>
<td>-5.4%</td>
<td>-2.7%</td>
</tr>
<tr>
<td>TOA net, W/m²</td>
<td>-0.42</td>
<td>8%</td>
<td>-8.5%</td>
<td>-4.2%</td>
</tr>
<tr>
<td>Surface SW, W/m²</td>
<td>-1.3</td>
<td>-9.1%</td>
<td>-3.4%</td>
<td>0.14%</td>
</tr>
<tr>
<td>Surface net, W/m²</td>
<td>-0.88</td>
<td>-13%</td>
<td>-4.9%</td>
<td>0.2%</td>
</tr>
<tr>
<td>Surface temperature, K</td>
<td>-0.1</td>
<td>2.5%</td>
<td>-7.6%</td>
<td>-2.8%</td>
</tr>
<tr>
<td>Atmospheric temperature, K</td>
<td>-0.11</td>
<td>8.9%</td>
<td>-8.7%</td>
<td>-4.2%</td>
</tr>
<tr>
<td>Latent heat, W/m²</td>
<td>-0.66</td>
<td>-5.1%</td>
<td>-6.2%</td>
<td>-1.3%</td>
</tr>
</tbody>
</table>

*Optical properties are 0.01 for $\omega$ and $g$ and 100 mbar for $\zeta$. One property is varied, while the other two are held constant at their default values. Positive percentages indicate that an increase in the optical property increases the magnitude of the forcing, temperature, or latent heat value.

3.2. Climate Sensitivity

Next we allow the climate to adjust to the imposed shortwave forcings and compare the results to those of the climate without dust. Both the surface (Figure 2, top) and atmospheric (Figure 2, bottom) steady state responses are more closely like the TOA forcing (Figure 1, top), as opposed to the surface (Figure 1, middle) or atmospheric (Figure 1, bottom) forcing. For example, the surface temperature change (Figure 2, top) can be either positive or negative, despite the fact that the surface forcing (Figure 1, middle) is always negative. Miller et al. [2004] obtain a similar response when they vary the single scattering albedo alone. For global average TOA forcing values of 0.76, -0.18, and -0.82 W/m², they obtain global average surface air temperature responses of 0.05, -0.21, and -0.4 degrees C, respectively (R. Miller, personal communication, 2005). The configuration with the surface temperature increase corresponds to the most negative surface forcing of the three cases.

While the atmospheric and surface temperature sensitivity patterns are similar to the TOA forcing pattern, the latent heat change has a notably unique pattern (Figure 3, top). For a given asymmetry parameter, the latent heat change forms a saddle pattern in phase space. For some $\zeta$ values, the magnitude of the latent heat change increases with single scattering albedo, while for others, the magnitude decreases. Thus the latent heat pattern matches none of the forcing patterns (Figure 1). Instead, the sign of the sensitivity to $\omega$ depends on the location of the dust layer with respect to clouds or other atmospheric reflectors. If the dust layer is below all the clouds, the magnitude of the hydrological cycle change decreases as the single scattering albedo increases. On the other hand, if the dust layer is above all the clouds, the magnitude of the hydrological cycle change increases with increasing single scattering albedo. The sensible heat pattern (Figure 3, bottom) is similar, though smaller in magnitude, to the surface forcing (Figure 1, middle).

Figure 4 illustrates the relationship between the climate response and the imposed TOA and surface forcings for each set of single scattering albedo, asymmetry parameter, and $\zeta$ values. We also include results from different versions of the model. The black dots correspond to the regular version of the model, while the colored points indicate climate responses in versions of the model where different feedbacks are omitted. For example, the blue points correspond to a version of the model where the lapse rate is held constant to the values from the dust-free version of the model. Similarly, the red points indicate results when the water vapor feedback is omitted. In the left-hand plots, points with the same x-axis values correspond to steady states with the same TOA forcing but different surface forcings. Similarly, the plots on the right side show different steady states based on the surface forcing, but with different TOA forcings. Again, we see that the atmospheric and surface temperature responses closely correspond to the TOA forcing, rather than the surface forcing. This correspondence occurs regardless of the specific feedbacks used (with the exception of the surface temperature response in the case where latent and sensible heat fluxes are held constant). Thus changes in surface temperature and atmospheric temperature can be estimated from the TOA forcing. The latent heat change, on the other hand, depends on both the TOA and surface forcings.

This correspondence between the temperature change and the TOA forcing due to dust has been previously explained [Cess et al., 1985; Miller and Tegen, 1998, 1999]. In a steady state, TOA dust forcing is primarily balanced by changes in the amount of outgoing longwave radiation (assuming shortwave...
feedbacks are small compared to the total dust forcing). Since very little radiation from the surface passes through the atmosphere without being reabsorbed, most of the longwave radiation is emitted to space from the atmosphere, and the atmosphere emission must change appropriately. Assuming that lapse rate and emissivity changes are second-order effects, the only way the atmospheric emission can change is if the atmospheric temperature changes. Thus the atmospheric temperature increases when the TOA forcing is positive and decreases when the forcing is negative. Although our model omits shortwave feedbacks (for example, cloudiness changes), more complicated models confirm that this atmospheric temperature adjustment is the primary way the climate responds to an imposed dust forcing [Miller and Tegen, 1998].

[27] The surface temperature change is linked to the atmospheric temperature change through convective coupling [Cess et al., 1985; Miller and Tegen, 1998, 1999], which is included in our model by the interactive lapse rate. Since the lapse rate changes are small, the surface temperature changes approximately follow the atmospheric temperature changes. Our lapse rate parameterization assumes that there is enough convection in the tropics to maintain a moist adiabatic lapse rate, and thus the lapse rate does not change very much with the climate. If convection decreases sufficiently (as a result of atmospheric warming and surface cooling by dust), this lapse rate parameterization is no longer valid. In this case, the surface temperature would become decoupled from the atmospheric temperature, resulting in surface cooling [Cess et al., 1985]. However, GCM experiments suggest that, for present-day dust concentrations, there is sufficient convection to maintain coupling between the surface and atmosphere [Miller and Tegen, 1998].

[28] To determine how dependent these results are on various feedbacks within the system, we recalculate the steady state responses using versions of the model that hold various climate components constant. The close relationship between the surface temperature change and the TOA forcing holds for versions of the model which omit the lapse rate and water vapor feedbacks. However, when latent

Figure 3. Global average steady state (top) latent and (bottom) sensible heating change as a function of dust single scattering albedo and (left) asymmetry parameter and (right) $\zeta$. Negative values correspond to decreases in the flux from the surface to the atmosphere and represent warming (cooling) effects on the surface (atmosphere).
and sensible heat flux feedback is omitted (green dots in Figure 4), the surface temperature is no longer coupled to the boundary layer temperature, so there is no convective coupling between the atmospheric temperature and surface temperature. Thus these surface flux changes are essential for the close relationship between TOA forcing and surface temperature.

[29] We again consider the effects of a standardized change in optical properties on our “reference” state. According to Table 1, in the region of phase space near

<table>
<thead>
<tr>
<th></th>
<th>Base model</th>
<th>$F_s$ and $F_L$ constant</th>
<th>$L_r$ constant</th>
<th>$e$ constant</th>
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</thead>
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<tr>
<td>TOA</td>
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<td><img src="image7.png" alt="Graph" /></td>
<td><img src="image8.png" alt="Graph" /></td>
</tr>
<tr>
<td>Latent heat change</td>
<td><img src="image9.png" alt="Graph" /></td>
<td><img src="image10.png" alt="Graph" /></td>
<td><img src="image11.png" alt="Graph" /></td>
<td><img src="image12.png" alt="Graph" /></td>
</tr>
</tbody>
</table>

**Figure 4.** Global average steady state (top) atmospheric temperature, (middle) surface temperature, and (bottom) latent heat change as a function of (left) TOA and (right) surface shortwave dust forcing for each set of optical parameters in the shortwave sensitivity experiment. Also shown are results from versions of the model which omit the latent and sensible heat flux, emissivity, or lapse rate feedback.
the reference parameter values, the atmospheric temperature responds the most to variations of 0.01 in the single scattering albedo and asymmetry parameter, while the surface temperature is most influenced by the asymmetry parameter. Thus the surface temperature is more sensitive to the asymmetry parameter than the TOA shortwave forcing is. This difference can also be seen in the different slopes of isolines in the Figure 2 (top left) and Figure 1 (top left) and also in the spread of points in the surface temperature plots of Figure 4. However, the surface temperature still behaves most like the TOA forcing than any of the other forcings. Finally, both the single scattering albedo and asymmetry parameter can change the latent heat flux.

3.3. Longwave Response and Sensible and Latent Heating Changes

The latent heat flux response (Figure 3) behaves differently from the TOA, atmospheric, and surface forcings (Figure 1). To determine how the latent heat flux interacts with other climate components and understand what determines this unique pattern, we examine results from a version of the model that excludes changes in the latent and sensible heat fluxes (Figure 5). By specifying these fluxes to be their dust-free steady state values, the surface and atmospheric solar forcings can be balanced (in the global average) only by an altered longwave budget. The atmospheric temperature change (Figure 5, bottom) is very similar to the change found in the case where latent and sensible heating vary (Figure 2, bottom), though slightly warmer. The surface temperature change pattern (Figure 5, top), on the other hand, is drastically different (compare it with Figure 2 (top)), in addition to being cooler overall. If the latent and sensible heat fluxes were allowed to vary, they would decrease as the surface temperature decreases, warming the surface and cooling the atmosphere. Thus the overall surface and atmospheric temperature changes are simply related to an elimination of this feedback. However, to explain the different pattern for the surface temperature change, we examine the details of the longwave response.

On the global average, when there is no latent or sensible heat flux change, the atmospheric and surface solar forcings must be balanced by equal and opposite changes in net longwave radiation, to achieve a steady state. We develop a simple analytical model, based on the full
computational model [Shell and Somerville, 2005] to estimate the longwave forcing change. Assuming the longwave forcing due to dust is insignificant,

$$\Delta S = \Delta I_{s,e} - \Delta I_{s,ab}$$  \hspace{1cm} (3)$$

$$\Delta S = \Delta I_{s,e} - \Delta I_{s,ab}$$  \hspace{1cm} (4)$$

where $\Delta S$ is the shortwave forcing, and $\Delta I$ is the change in a longwave term for the atmosphere, subscript $a$, or surface, subscript $s$. The second subscript for the longwave radiation terms refers to either emission, $e$, or absorption, $ab$. For example, $\Delta I_{s,e}$ corresponds to a change in the longwave emission by the atmosphere. Increases in shortwave radiation absorption are therefore balanced by a combination of increases in longwave emission and decreases in absorption.

[32] The atmosphere absorbs a fraction $\epsilon$ of the emitted surface longwave radiation. Assuming $\epsilon$ does not change significantly with the climate,

$$\Delta I_{s,ab} = \epsilon \Delta I_{s,e}$$

We use a value of 0.9, the global average emissivity in the full model, for $\epsilon$. We also assume that the fraction of radiation emitted by the atmosphere which is in the direction of the surface (3/5) does not change. Thus

$$\Delta I_{s,ab} = 0.6 \Delta I_{s,e}$$

[33] Combining these equations yields the change in longwave emission from the atmosphere and surface based on the shortwave radiative forcing:

$$\Delta I_{s,e} = \frac{\Delta S_s + \epsilon \Delta S_s}{1 - 0.6 \epsilon}$$

$$\Delta I_{s,e} = \frac{0.6 \Delta S_s + \epsilon \Delta S_s}{1 - 0.6 \epsilon}$$

The predicted changes in surface longwave emissions are shown in Figure 6 (top). These changes are very similar to the changes obtained from the computer model without latent and sensible heating changes (Figure 6, bottom). Changes in atmospheric longwave emissions are also similar between the two methods (not shown).

[34] Thus the analytic model approximates the results of our computer model without the latent and sensible heat flux feedback. The results differ slightly because the emissivity and the fraction of atmospheric longwave emission toward the surface change slightly as the solar forcing varies. Also, we have ignored longwave forcing due to dust.

[35] When we allow the latent and sensible heat fluxes to vary, equations (3) and (4) are no longer valid. Latent and sensible heat fluxes modify the exchange of energy between the surface and the atmosphere. The latent heat change pattern from the full model (Figure 3, top) is very similar to the change in surface emission in the constant latent heating model (Figure 6, bottom), though smaller in magnitude. The sensible heating change pattern, on the other hand, most closely matches the surface shortwave forcing. The latent heating is generally more sensitive to changes in temperature than the surface longwave emission or sensible heat flux. Changes in latent heating account for most of the surface-atmosphere flux changes necessary to balance the imposed forcing. Our model assumes that enough surface moisture is available to supply the necessary evaporation. GCM experiments indicate that, in arid regions, the surface radiative forcing is balanced primarily by changes in the sensible heat flux [Miller et al., 2004]. Thus the local sensible heat flux behavior may be more complex than in our model. However, on a global scale, although dust is normally generated in arid regions, it is generally advected over ocean or vegetated land, where there is sufficient surface moisture.

4. Sensitivity to Longwave Radiative Properties

[36] Less attention has been paid to uncertainty in longwave optical properties, since the longwave forcing (i.e., the “greenhouse effect” of dust) is generally smaller in magnitude than the shortwave forcing. However, the longwave forcing has a different vertical distribution than the shortwave forcing. Thus changes in longwave optical properties may result in different climate behavior. In addition, in regions of parameter space where the shortwave forcing is small enough, the longwave forcing will dominate.

[37] Liao and Seinfeld [1998] and Claquin et al. [1998] investigate the sensitivity of the dust longwave forcing to changes in dust optical properties and vertical distribution using 1-dimensional column radiative transfer models. While they only consider a few cases and varied parameters in isolation, we explore the global dust longwave forcing more thoroughly, considering simultaneous variations in longwave single scattering albedo, asymmetry parameter, and height over a range of values. In addition, we extend these calculations to include the climate response to this forcing, which requires the use of a global model.

[38] Note that the longwave dust height is different than the shortwave coordinate $\zeta$. The longwave height determines the temperature of the dust layer for use in the forcing calculations. $\zeta$, on the other hand, determines the distribution of solar reflection above and below the dust layer. Thus the two are not necessarily equivalent and can vary independently. For example, if a dust layer is moved from just below a thin cloud to just above it, the shortwave coordinate $\zeta$ will change significantly, since there is now much less atmospheric reflection occurring above the dust layer. However, the temperature of the dust layer will not have changed very much, so the longwave effect will be similar before and after the dust layer is moved.

4.1. Longwave Forcing Sensitivity

[39] The instantaneous longwave atmospheric forcing (Figure 7, bottom) can be warming or cooling, depending on the height (temperature) of the dust. A greater height corresponds to a lower dust temperature and thus less emission by dust to space and a stronger greenhouse effect. Conversely, a reduced height corresponds to a higher dust temperature and increased emission. The transition from heating to cooling occurs around 5 km, regardless of the single scattering albedo. The sensitivity to height decreases
as the single scattering albedo increases. The atmospheric forcing is not sensitive to changes in the asymmetry parameter. Since the model uses the single scattering approximation, scattered longwave radiation does not contribute to atmospheric absorption, and the direction of scatter is unimportant. Results from a radiative transfer model that includes multiple longwave scattering also show that the effect of scattering on the atmosphere is small \[\text{Dufresne et al., 2002}\].

The surface longwave forcing (Figure 7, middle) is always positive because dust enhances the greenhouse effect. The surface forcing decreases with dust height, in agreement with column model results [\text{Liao and Seinfeld, 1998; Claquin et al., 1998}]. High dust emits little radiation to space or the surface. Thus the atmosphere warms, but the amount of radiation absorbed by the surface is smaller than when the dust is lower. With an increased longwave single scattering albedo, the emissivity of the dust layer decreases, and dust emits less radiation toward the ground (though this effect is partially cancelled by the increase in reflected longwave radiation back down to the ground, see equation (23) in the auxiliary material of Shell and Somerville [2007]). As the asymmetry parameter increases, longwave radiation emitted by the surface is more likely to be scattered by the dust layer forward to space as opposed to back toward the surface. Thus the most positive surface forcing occurs when the single scattering albedo, asymmetry parameter, and height are low. Claquin et al. [1998] and Liao and Seinfeld [1998] show that longwave surface forcing increases as size increases (until the mass median diameter reaches 5 \(\mu m\)), corresponding to smaller single scattering albedos and larger asymmetry parameters [\text{Tegen and Lacis, 1996}]. In our model, we vary the single scattering albedo and asymmetry parameter separately, so we can see that the decrease of single scattering albedo with dust size results in a larger forcing, which is partially cancelled by the decrease in forcing as a result of the larger asymmetry parameter associated with larger particles.

\[\text{Figure 6.} \] Global average steady state surface longwave emission change as a function of dust single scattering albedo and (left) asymmetry parameter and (right) \(\zeta\), (top) from the idealized analytic model without latent and sensible heating and (bottom) from the full model when latent and sensible heating are held constant.
Figure 7. Global average instantaneous longwave (top) TOA, (middle) surface, and (bottom) atmospheric heating change as a function of dust longwave single scattering albedo and (left) asymmetry parameter and (right) height. Note the different scales used on this and the other longwave plots compared to the shortwave sensitivity plots.
forcing is highest for low single scattering albedos and asymmetry parameters. However, the forcing increases as the height increases, as expected [Liao and Seinfeld, 1998; Claquin et al., 1998; Myhre and Stordal, 2001], since high dust traps more longwave radiation, even if it influences the surface less. Similarly to the surface forcing results, we see a partial cancellation between the single scattering albedo and asymmetry parameter effects for the case when size is varied. Previous radiative transfer model results indicate that the net effect (including extinction changes) is an increase of TOA forcing with size [Claquin et al., 1998; Liao and Seinfeld, 1998; Myhre and Stordal, 2001].

For the shortwave case, the magnitude of the TOA forcing is generally less than the magnitudes of the atmospheric and surface forcings. In the longwave case, on the other hand, the three forcings are closer in magnitude, though their maximum values occur at different optical properties and heights. These differences in the vertical distribution of forcing will influence the climate changes necessary to reach a new steady state.

### 4.2. Climate Response

Since the shortwave forcing is larger than the longwave forcing, the longwave forcing, we examined the climate sensitivity to dust with no shortwave effect to isolate the effects of the longwave optical properties. If the dust shortwave forcing is included, it alters the overall magnitude of change but leaves the pattern unchanged.

In response to longwave dust forcing, both the atmosphere (Figure 8, middle) and surface (Figure 8, top) warm. While the magnitudes of temperature change are smaller than those found in the shortwave case, sign changes in the shortwave forcing indicate that the longwave temperature response may dominate when the shortwave radiative change is small. The atmosphere and surface warm together, based on the TOA heating, with the atmospheric warming slightly more.

The latent heating change (Figure 8, bottom) again plays a role in the response. While the atmospheric temperature looks fairly similar both within and without the latent and sensible heat feedback (not shown), the surface temperature responses are significantly different (compare Figure 8 (top) to Figure 9). In the case where latent and sensible heat fluxes are held constant, the surface cannot cool by increasing these fluxes, thus it warms more in response to dust longwave forcing.

Table 2 shows that the surface and atmospheric temperatures are most sensitive to changes in the dust height of 100 m; however, the latent heat change is most sensitive to the longwave single scattering albedo and asymmetry parameter. For all three variables, though, changes in the shortwave properties result in larger variations than changes in the longwave properties.

### 5. Conclusions

We use the simple model of Shell and Somerville [2007] to investigate the climate response to direct radiative forcing by mineral aerosols. The model produces a realistic mean climate and dust forcing [Shell and Somerville, 2007] and thus is a useful tool for exploring the dust-climate system behavior. In this paper, we explore the sensitivity of the dust forcing and climate response to dust radiative properties. Our goal is the determine the parameters most responsible for uncertainty in the climate response to the present-day dust distribution.

The TOA shortwave forcing is most negative when the single scattering albedo is high, the asymmetry parameter is low, and the dust layer is above clouds or other reflective particles. The forcing is most positive when the single scattering albedo is low, the asymmetry parameter is high, and the dust layer is above clouds. For a representative set of optical properties, the TOA shortwave forcing is more sensitive to a change of 0.01 in the single scattering albedo or asymmetry parameter than to a change of 100 mbar in the vertical coordinate $z$.

The TOA longwave forcing is largest for a high dust height, low single scattering albedo, and low asymmetry parameter. The TOA and surface longwave forcings are always positive, but the sign of the atmospheric forcing depends on the height of the dust layer. The height of the dust layer affects the sensitivity of longwave forcing to changes in single scattering albedo and asymmetry parameter. In addition, the largest forcing changes are found at low heights (for surface forcing), high heights (TOA forcing), or height extremes (atmospheric forcing). The TOA longwave forcing is more sensitive to a height change than variations in the longwave optical properties.

The net TOA forcing is the most sensitive to changes in the shortwave single scattering albedo and asymmetry parameter. Longwave properties are not as important, since the magnitude of longwave forcing is smaller than the shortwave forcing.

The sensitivities of the atmospheric and surface temperatures are similar to the sensitivity of the TOA shortwave or longwave forcing rather than the atmospheric or surface forcings. A negative shortwave surface forcing
Figure 8. Global average steady state (top) surface and (middle) atmospheric temperature and (bottom) latent heat change as a function of dust longwave single scattering albedo and (left) asymmetry parameter and (right) height for the case of no shortwave forcing.
does not necessarily correspond to a steady state surface temperature decrease. While the atmospheric temperature correspondence to TOA forcing appears independent of the various feedbacks tested in our model, latent and sensible heat fluxes play an important role in the response of surface temperature to an imposed forcing. The atmospheric and surface temperatures respond the most to changes in the shortwave single scattering albedo and asymmetry parameter. In particular, the asymmetry parameter, which is often considered secondary in uncertainty to the single scattering albedo, results in the largest surface temperature response range. These results suggest that refinement of the shortwave asymmetry parameter, as well as the shortwave single scattering albedo, is important for a better estimate of the effect of dust. Note, though, that these results are based on the specific default optical properties we used, and may change significantly if different values are used for the base case.

Both the atmospheric and surface temperatures are closely tied to the TOA forcing, as opposed to surface or atmospheric forcing, in agreement with previous work. However, this relationship had only been previously demonstrated for a few particular dust-climate configurations. With our model, we are able to explore a much larger range of parameters and to confirm that this relationship is very robust over a large number of different configurations but dependent on the ability of latent and sensible heat fluxes to respond to climate forcing.

While the atmospheric and surface temperature sensitivity patterns are similar to the TOA forcing pattern, the latent heat change in the shortwave optical property experiment has a notably unique pattern. For a given asymmetry parameter, the latent heat change forms a saddle pattern in phase space. For some values of the shortwave vertical coordinate $\zeta$, the magnitude of the latent heat change increases with single scattering albedo, while for others, the magnitude decreases. This pattern corresponds to the change in energy flux from the surface to the atmosphere necessary to achieve a steady state, which depends on the vertical distribution of dust forcing. Thus, in order to accurately predict hydrological changes, knowledge of the vertical distribution of dust with respect to other atmospheric reflectors is required. For example, if the dust layer is entirely below clouds, an increase in single scattering albedo leads to a smaller hydrological cycle decrease. If, on the other hand, the dust is above clouds, an increase in single scattering albedo leads to a larger hydrological cycle decrease. New data from satellites such as CALIPSO, which will provide vertical profiles of dust and clouds, will be essential for determining the effects of dust on the hydrological cycle.

In this paper, we have focused on the effects of variations in single scattering albedo, asymmetry parameter, and height. Uncertainty in the amount of dust also increases the uncertainty of these ranges. Recent estimates of the present-day dust loading vary by a factor of 4 [Zender et al., 2004]. Shell and Somerville [2007] demonstrate that the effect of dust varies approximately linearly with optical depth. Thus uncertainty in the amount of dust will also contribute to uncertainty in the effect of dust.

We have shown that this new model can be useful for exploring a wide range of parameters in the dust-climate system. The model can be adapted to other aerosol species, but its simplicity is especially useful for the case of dust, since many optical properties are uncertain. However, the model omits many processes and feedbacks, which may be important to the climate response. To obtain a complete picture of the effect of dust on climate, results from this model should be considered together with observations and results from more complete models (GCMs).

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Figure 9. Global average steady state surface temperature change as a function of dust longwave single scattering albedo and (left) asymmetry parameter and (right) height for the fixed latent and sensible heating and no shortwave forcing case.
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References


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