Surface Radiative Forcing By Soil Dust Aerosols and the Hydrologic Cycle

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Abstract

For absorbing aerosols like soil (or ‘mineral’) dust, the surface radiative forcing differs substantially from the value at the top of the atmosphere (TOA). The climate response depends not only upon the TOA value, but its difference with respect to the surface value, which represents radiative heating within the atmosphere. Moreover, the surface forcing impacts the hydrologic cycle, which can feed back upon the aerosol burden through the efficiency of wet deposition. We calculate the surface forcing by soil dust aerosols and compare its global sensitivity to aspects of the dust distribution that are poorly constrained by observations. Ignorance of the global dust burden corresponds to a forcing uncertainty of over a factor of two, with smaller uncertainties due to imprecise knowledge of particle optical properties and the particle size distribution.

While global evaporation and precipitation are reduced in response to surface radiative forcing by dust, precipitation increases locally over desert regions, so that dust emission can act as a negative feedback to desertification. The global reduction in precipitation potentially increases the particle lifetime by reducing the efficiency of wet deposition, while increasing the source area for emission by reducing vegetation. Both of these mechanisms represent a positive feedback to the global dust burden and its radiative forcing. For the current climate, we calculate the reduction in wet deposition and find that the dust burden is increased only modestly. However, the global dust burden is substantially higher during glacial climates, so that the amplification of the dust load by these feedbacks is larger. By extrapolating from its radiative forcing in the current climate, we estimate that dust reduces precipitation during glacial times by as much as half the reduction due to the colder climate alone.
1 Introduction

Soil (or ‘mineral’) dust particles are the most common aerosol by mass (Andreae 1995), entering the atmosphere through wind erosion of dry soils. While the atmospheric dust load is largest near deserts and soils disturbed by cultivation, the smallest particles, which are the most radiatively active by mass, can travel thousands of kilometers downwind of their source region. African dust is observed over both North and South America (Prospero and Nees 1977; Prospero et al. 1981; Perry et al. 1997; Prospero 1999), and to the east over the Indian Ocean (Meywerk and Ramanathan 1999). Soil dust from Asian deserts is measured at remote islands within the Pacific (Duce et al. 1980; Parrington et al. 1983; Rea 1994), and as far downwind as North America and Greenland (Biscaye et al. 1997; Reader et al. 1999; Bory et al. 2002; Bory et al. 2003), and even Europe (Grousset et al. 2003). While wind-blown particles have accumulated in deep-sea sediments throughout geologic time (Rea 1994; Kohfeld and Harrison 2001), increasing dustiness is observed in certain regions (N’Tchayi Mbourou et al. 1997; Prospero 1999; Nicholson 2000).

While radiative forcing by soil dust has been recognized for decades (Coakley and Cess 1985), its precise value remains unknown. The anthropogenic component, attributed to dry soils disturbed by agriculture, overgrazing, and deforestation, is possibly comparable to that of other tropospheric aerosols (Tegen and Fung 1995; Tegen et al. 1996; Sokolik and Toon 1996; Tegen et al. 1997), although this is still debated (Haywood and Boucher 2000; Ginoux et al. 2001; Mahowald et al. 2002). Unlike radiative forcing by greenhouse gases, which is fairly uniform throughout the troposphere (Hansen et al. 1997), forcing by dust and other tropospheric aerosols exhibit large regional variations due to their short lifetime. Downwind of major source regions, dust radiative forcing often dominates (Li et al. 1996; Chiapello et al. 1999).

The present-day distribution of dust must be inferred from a combination of data sources, as there is no single definitive set of observations. Multi-decadal surface measurements exist downwind of certain major source regions like the Sahara and Sahel (Prospero 1996). Column integrals of aerosols are provided by satellite retrievals (Moulin et al. 1997; Husar et al. 1997; Herman et al. 1997; King et al. 1999; Torres et al. 2001) and sun photometers (Holben et al. 1998).
1998), although the dust contribution is entangled with that of other aerosol species. The vertical distribution of dust within the column is revealed by LIDAR (Light Detection and Ranging) measurements, usually within detailed observing campaigns of limited duration (Karyampudi et al. 1999). Dust models are an effort to estimate the three dimensional distribution of dust and its evolution in time subject to all of these observational constraints (Tegen and Fung 1994; Guelle et al. 2000; Perlwitz et al. 2001; Ginoux et al. 2001; Tegen et al. 2002; Luo et al. 2002; Zender et al. 2003).

In contrast to sulfate aerosols formed by industrial pollution, dust particles absorb as well as scatter sunlight (Lacis and Mishchenko 1995; Tegen and Lacis 1996). Absorption and scattering have offsetting effects upon solar forcing at the top of the atmosphere (TOA), while acting in concert to reduce forcing at the surface. Dust particles exhibit the greenhouse effect by absorbing at thermal wavelengths, resulting in positive forcing at both TOA and the surface. The effect of aerosols upon climate is often characterized by their forcing at TOA (IPCC 2001). However, for absorbing aerosols, the climate response depends not only upon the TOA forcing, but also upon its difference with respect to the surface value, which represents radiative heating within the atmosphere (Miller and Tegen 1998; Miller and Tegen 1999; Ramanathan et al. 2001). Indeed, the change in surface temperature in the former study is largely the result of this heating rather than the TOA forcing per se.

This article calculates surface radiative forcing by soil dust aerosols, with emphasis upon its global uncertainty. The climate response to dust depends upon the global average of the forcing, or at least the average over the scale of the perturbed circulation. Despite estimates by Miller and Tegen (1998) and Woodward (2001), global radiative forcing at the surface remains uncertain due to imprecise knowledge of the dust distribution. Table 1 shows that among current models, estimates of the global dust burden vary by over a factor of two, while the ratio of clay aerosol to larger silt particles varies at least fourfold. We quantify the global sensitivity of the surface forcing to these uncertainties, along with those in the particle optical properties, where there is disagreement among laboratory (Patterson et al. 1977) and in situ measurements (Kaufman et al. 2001; Sinyuk et al. 2003). Single column models have been used to investigate the forcing sensitivity in the context of a
particular regional climate (Sokolik and Golitsyn 1993; Claquin et al. 1998; Liao and Seinfeld 1998). Here we rank contributions by poorly known aspects of the dust distribution to the global forcing uncertainty. Our calculation of the global sensitivity of surface forcing complements the calculation of TOA sensitivity by Myhre and Stordal (2001), who used a similar distribution (Tegen and Fung 1995). In Section 2, we present our dust distribution, calculated as a radiatively active tracer within the NASA Goddard Institute for Space Studies (GISS) atmospheric general circulation model (AGCM), using a parameterization described by Tegen and Miller (1998). In Section 3, we calculate the surface forcing, and identify aspects of the dust distribution making the greatest contribution to the forcing uncertainty.

The surface radiative forcing corresponds mainly to a reduction of sunlight beneath the dust layer. This is balanced largely by a reduction of surface evaporation, which weakens the hydrologic cycle (Coakley and Cess 1985; Miller and Tegen 1998; Ramanathan et al. 2001). However, we show in Section 4 that while evaporation and precipitation are reduced globally by dust, rainfall increases over deserts: a negative feedback by dust emission to desertification. The global reduction in precipitation increases the dust burden by reducing the efficiency of wet deposition. By comparing the dust distribution computed by the AGCM with and without dust radiative forcing, we calculate this reduction, and its amplification of the dust burden. Yung et al. (1996) speculate that the increased dust load during glacial times (Petit et al. 1990; Biscaye et al. 1997; Reader et al. 1999; Mahowald et al. 1999; Kohfeld and Harrison 2001) is due to the weakening of the hydrologic cycle by the colder climate. For comparison, we estimate to what extent the hydrologic cycle was diminished by dust radiative forcing. Our conclusions are presented in Section 5.
2 Dust Distribution

Surface radiative forcing by soil dust is calculated by first computing the atmospheric distribution of dust, parameterized as a tracer within the NASA GISS AGCM (Tegen and Miller 1998), coupled to a mixed-layer ocean (Miller et al. 1983). The distribution of dust depends upon the atmospheric circulation, which is perturbed by dust radiative heating, so that the aerosol is an interactive part of the AGCM.

The AGCM used to calculate the dust distribution has horizontal resolution of 4° latitude by 5° longitude, with 12 vertical layers extending from the surface to 10 mb. Tracers (including dust) are advected using a quadratic upstream scheme. This computes not only the tracer value at each grid point but its gradient and curvature, resulting in higher effective resolution of tracers (Prather 1986). Calculation of realistic dust emission depends upon a model’s ability to simulate the observed surface wind and soil moisture. The AGCM surface wind is calculated using similarity theory (Hartke and Rind 1997), with a second-order closure scheme used to relate the interior winds to the surface value (Mellor and Yamada 1982; Galperin et al. 1988). The model’s hydrologic cycle was ranked in the top quartile by the Atmospheric Model Intercomparison Project (Lau et al. 1996).

The model transports four size categories of soil dust, including clay particles (defined by radii less than 1 μm), and three silt categories (Tegen and Miller 1998). Clay particles settle gravitationally with nearly identical fall speeds, and can be advected to a good approximation as a single size class (Tegen and Lacis 1996).

Calculation of emission depends upon identifying potential source regions and determining whether the meteorological conditions for emission are satisfied. Intuitively, emission should occur over dry regions where there is abundant loose soil vulnerable to wind erosion. In practice, this has led to a number of plausible physical criteria, summarized by Ginoux et al. (2001), that lead to markedly different estimates of emission in some regions. Following Tegen and Fung (1994), we use the vegetation cover data set compiled by Matthews (1983) to allow dust emission in desert, grassland, and shrub regions, where vegetation is low and sparse. Vegetation is fixed, even though Tegen et al. (2002) show that Asian dust
emission is increased if grass cover emerges only after the onset of spring rains. We also allow emission in dry soils disturbed by agriculture, overgrazing or deforestation (Tegen and Fung 1995; Tegen et al. 1996). By these criteria, emission can occur potentially over a third of the global land surface. Recent dust models attempt to identify ‘preferred’ source regions, such as the exposed bed of former lakes, where there are fine, erodible particles in abundance (Ginoux et al. 2001; Tegen et al. 2002; Luo et al. 2002; Zender et al. 2003). Our model does not explicitly include these sources. While their introduction brings the model dust load into better agreement with satellite retrievals in the vicinity of source regions, the large-scale distribution, upon which our global forcing depends, is comparatively insensitive (Zender and Newman 2003).

Wind-tunnel measurements suggest that emission occurs when the friction speed (defined as the square-root of the wind-stress divided by the air density) exceeds a threshold (Gillette 1978). The friction speed is related to the 10 meter wind speed by the surface roughness parameter (Arya 1988). In the absence of a global data set of surface roughness, whose resolution is comparable to source variations, we calculate emission solely in terms of surface wind speed, a method which has been shown to represent the seasonal pattern of Saharan emission reasonably well (Marticorena et al. 1999). We assume that saltation is the dominant means of lifting small particles from the surface (Shao et al. 1993), so that the threshold wind speed is interpreted as the value that lifts the larger, saltating particles (Schulz et al. 1998). We allow the threshold to vary geographically so that emission approximately equals that computed by the offline transport model of Tegen and Lacis (1996), where ECMWF analyzed surface winds with $1\frac{1}{8}^\circ \times 1\frac{1}{8}^\circ$ horizontal resolution were used. Where wind speeds beneath the scale of the AGCM grid box are large, as indicated by the ECMWF analyses, the threshold wind speed is reduced (Tegen and Miller 1998). The global average threshold (weighted geographically by emission) is $4.5\text{ ms}^{-1}$, compared to $6.5\text{ ms}^{-1}$ in the offline model (Tegen and Fung 1994).

Then, emission $\mathcal{E}$ is calculated according to the formula (Tegen and Fung 1994):

\[
\mathcal{E} = \begin{cases} 
C f(r) w^2 (w - w_T) & \text{for } w \geq w_T, \\
0 & \text{for } w < w_T.
\end{cases}
\] (1)
Here, $C$ is a constant of proportionality with units of $\text{kg s}^2\text{m}^{-5}$, $f(r)$ is the fraction of the grid box within which dust particles of radius $r$ are available, $w$ is the surface wind speed, and $w_T$ is the threshold speed above which emission takes place. To account for the binding effect of soil moisture, emission is permitted only if evaporation exceeds precipitation for a specified duration ranging from a day for sandy soils to ten days for soils comprised of small clay particles.

Emission of a particular particle size depends upon its availability within a grid box, which is represented by $f(r)$ in (1). The soil type and particle size data sets of Zobler (1986) and Webb et al. (1991) give the fractional area exhibiting each size category; $f(r)$ represents this fraction multiplied by $\frac{1}{6}$ for clay and $\frac{1}{8}$ for each silt category to account for the erodible component (Tegen and Fung 1994). In order that our model lie within the estimated range of global emission, we choose the coefficient of proportionality $C$ in (1), which is model dependent, equal to $52 \mu\text{g s}^2\text{m}^{-5}$. Averaged globally (weighting by the annual mean emission), the product $Cf$ is $1.7 \mu\text{g s}^2\text{m}^{-5}$ for clay particles, and $3.2 \mu\text{g s}^2\text{m}^{-5}$ for the sum of the three silt categories. On average, roughly twice as much silt as clay is emitted for each wind event exceeding the threshold.

Downwind transport of dust within the AGCM is calculated using model winds, rather than observed values. Dust radiative forcing can alter the circulation, a process not accounted for in current reanalysis models (Alpert et al. 1998; Weaver et al. 2002). On the other hand, AGCM winds may not agree everywhere with actual values, and these will distort the model trajectories of dust.

Dust is removed from the atmosphere by a combination of dry deposition through gravitational settling and turbulence, along with wet deposition, where particles are swept out by precipitation. Gravitational deposition depends upon particle size and is efficient only for the larger silt particles. Wet deposition is proportional to the surface precipitation and scrubs the column to a fixed height that is specified based upon the climatological extent of deep convection (Tegen and Fung 1994). Table 1 shows that the resulting deposition lifetimes are within the range of recent models (Ginoux et al. 2001; Tegen et al. 2002; Luo et al. 2002; Zender et al. 2003).
Dust emission within the AGCM, averaged seasonally both for various regions and the entire globe, is shown in Fig. 1. Averages are based upon a thirty-one year integration following a nineteen year spin-up, during which time the ocean mixed layer comes into equilibrium with the dust radiative forcing. The global and annual average emission is $10^{18} \pm 69$ Tg, slightly smaller than one standard deviation below the total calculated by Perlwitz et al. (2001), using an identical model but with prescribed climatological SST as a lower boundary condition. Dust emission is not directly observed, but the present model value is among the lower estimates. Based upon measurements of ocean deposition, Duce et al. (1991) estimate a lower bound of $910$ Tg yr$^{-1}$, while an upper bound of $3000$ Tg yr$^{-1}$ is suggested by Andreae (1995). Table 1 shows that emission of particles smaller than $10 \mu$m (which includes all particles light enough to be carried out over the ocean) calculated by recent models ranges from $1100$ to $1800$ Tg yr$^{-1}$ (Ginoux et al. 2001; Tegen et al. 2002; Luo et al. 2002; Zender et al. 2003).

According to Fig. 1b, global emission is dominated by the Sahara and Sahel in all seasons. The NH springtime peak in this region is consistent with observed Sahel visibility (N‘Tchayi Mbourou et al. 1997) along with observational estimates of emission in the Western Sahara (d’Almeida 1986). Emission is also large during the NH spring in Central and East Asia, and during the Southern Hemisphere spring and summer in Australia. While a third of the land surface has the potential to emit dust, two-thirds of the total emission originates from thirty grid boxes. These thirty represent a tenth of the potential source grid boxes, corresponding to a few percent of the total land surface. This emphasizes the localized nature of dust emission (Goudie and Middleton 2001; Prospero et al. 2002); the high winds associated with emission are confined to a few specific regions.

The global dust load is only loosely constrained by observations. While the models of Ginoux et al. (2001), Tegen et al. (2002), Luo et al. (2002), and Zender et al. (2003) reproduce observed surface concentrations (Prospero 1996), along with such measures of column amount as the Aerosol Robotic Network (AERONET) optical thickness (Holben et al. 1998) or the Total Ozone Mapping Spectrometer retrievals (Herman et al. 1997; Torres et al. 2001), global and annual averaged burdens in these models range from 17.4 to
35.9 Tg (Table 1). Our model value of 14.6 Tg is at the low end, suggesting that our forcing derived in the next section may underestimate the actual value.

The contribution of each size category to our global dust load is shown in Fig. 2. While nearly twice as much silt is emitted globally in comparison to clay (Table 2), the latter dominates the aerosol load. Clay particles are removed from the atmosphere mainly by wet deposition, which is less efficient compared to gravitational and turbulent settling of the larger silt particles. In Table 2, the particle residence time is estimated by dividing the aerosol load by the emission rate. Clay particles reside for nearly 10 days, three time longer than the larger and less buoyant silt particles (Table 2). Table 1 shows that our particle lifetimes are within the range computed by recent models.

The model size distribution is consistent with that measured a few thousand kilometers downwind of the Sahara, where clay particles are the most prevalent (Duce 1995). Clay particles also dominate the model distribution over gridboxes where emission is strongest (not shown). This is in contrast to AERONET retrievals (Ginoux et al. 2001; Ginoux 2003), where close to source regions, silt particles make the greatest fractional contribution. Fig. 2b shows the size distribution calculated by an offline transport version of the AGCM’s dust parameterization (Miller and Tegen 1998), where the silt burden is 1.4 times that of clay. The peak contribution to the aerosol load comes from the smallest silt category, rather than the largest clay category that makes the greatest contribution in the AGCM. The contrast between the two distributions in Fig. 2 is largely due to the planetary boundary layer parameterization used to compute the surface wind speed. Among recent models, the clay fraction of our global burden is large (Table 1), although there is little consensus about the correct value. Because of this disagreement, we consider the sensitivity of the surface forcing to particle size in the next section.

The geographic distribution of the atmospheric dust load is shown in Fig. 3. Over Africa, the dust plume swept westward by the Trades is centered at Sahelian latitudes during NH winter and downwind of the Sahara during summer. Dust from Central and East Asia extends over the Pacific during NH spring and summer. These features are qualitatively consistent with AVHRR retrievals of aerosol optical depth (Husar et al. 1997), although the
plume of Saharan dust extends insufficiently westward over the Atlantic during NH summer, mainly due to wet deposition by excessive AGCM precipitation along the trajectory (R. Cakmur, personal communication). A more precise comparison to AVHRR was carried out by Perlwitz et al. (2001), using a dust distribution computed by the same AGCM, but with climatological SST rather than a mixed-layer ocean as a lower boundary condition. [Despite the different lower boundary condition, the climatological load of the present model is highly correlated with that of Perlwitz et al. (2001), with a median spatial correlation of 0.97 over twelve months.] Perlwitz et al. (2001) demonstrate that the AGCM generally reproduces the spatial extent of the dust load, although it underestimates the global integral. Despite the global underestimate, the model exceeds the aerosol optical thickness inferred by AVHRR over Australia. By using ‘preferred’ source regions corresponding to dry, topographic depressions, Ginoux et al. (2001), Tegen et al. (2002), and Zender et al. (2003) simulate Australian emission in better agreement with surface observations, although this agreement may be sensitive to the surface wind analyses used to compute emission (Luo et al. 2002).

Regional averages of dust mixing ratio are given by Fig. 4. Each region is defined where the dust load exceeds 50 mg m\(^{-2}\) (Fig. 3). The African and Asia plumes are arbitrarily divided by the meridian at 25\(^\circ\) E poleward of 36\(^\circ\)N, and the meridian at 95\(^\circ\)E to the south. The mixing ratio falls off rapidly above the surface, although it remains non-negligible in the upper troposphere over Asia during NH spring and summer, and as high as the lower stratosphere during NH summer over Africa and the Arabian peninsula.
3 Surface Radiative Forcing

Radiative forcing is calculated using the geographic and size distribution of dust particles described in the previous section. Per unit mass, extinction by scattering and absorption peaks at a wavelength comparable to the particle size (Andreae 1995), and clay particles have radii comparable to the wavelength of maximum solar irradiance. While their nearly identical lifetimes allow them to be transported as a single size category, radiative forcing by clay particles depends upon the precise form of the size distribution within this category. We compute their radiative forcing by subdividing the clay category into four bins, with an effective radius of 0.1, 0.2, 0.4, and 0.8 \( \mu \text{m} \), respectively. The subdivision by mass is based upon the distribution of clay particles calculated explicitly by Tegen and Lacis (1996). The effective radius of each silt category is 1, 2, and 4 \( \mu \text{m} \), respectively. The contribution of larger particles to the global forcing is negligible (Tegen et al. 1996).

Radiative fluxes in the GISS AGCM are computed using the single Gauss point doubling/adding algorithm (Lacis and Mishchenko 1995). The radiative effect of dust is calculated using Mie theory, where particles are treated as homogeneous spheres, and absorption and scattering of radiation at a particular wavelength depend solely upon the index of refraction and particle size. We assume a globally uniform index of refraction, taken from laboratory measurements at infrared (Volz 1973) and solar wavelengths (Patterson et al. 1977) of far-traveled Saharan dust arriving at Barbados. A different index of refraction would be appropriate for other source regions, given their distinct mineralogical distributions; we subsequently examine the forcing sensitivity to changes in the particle optical properties.

In the GISS AGCM, aerosol radiative effects at solar and thermal wavelengths are treated separately. The solar spectrum is divided into six wavelength bands, with the shortest spanning 0.20 to 0.77 \( \mu \text{m} \). Multiple scattering effects are included in the solar calculation. At thermal wavelengths, scattering is small in comparison to absorption, and is neglected (Tegen and Lacis 1996). Dufresne et al. (2002) show that this neglect reduces the thermal contribution to forcing at TOA by half, although the reduction at the surface (and to radiative
forcing within the atmosphere) is less than 15%. Because the thermal contribution to global forcing will be shown to be relatively small, this neglect has little effect upon our conclusions.

Radiative forcing is a function of the size distribution within the dust layer (Andreae 1995). Fig. 5a shows the ratio of the scattering extinction to the total extinction at solar wavelengths. Both extinctions are integrated globally, annually, and over the entire range of particle sizes, creating a bulk single scattering albedo, \( \varpi_0 \). Also shown for the shortest solar wavelength band (0.20–0.77\( \mu \)m) is the single scatter albedo of each model size category (denoted by its equivalent radius). The bulk single scatter albedo lies near the value corresponding to the predominant particle size (Fig. 2a). The geographic distribution of the annual and column average \( \varpi_0 \) at the shortest solar band (0.20–0.77\( \mu \)m) is shown in Fig. 5b. Geographic variations in \( \varpi_0 \) result from corresponding variations in the particle size distribution and its evolution downstream of the source region. The single scatter albedo is smallest near the source due to the presence of larger particles, reaching a minimum of 0.89 in Fig. 5b. Downwind, where the size distribution changes due to sedimentation of larger particles, \( \varpi_0 \) increases to 0.91.

Dust radiative forcing is defined as the perturbation to the radiative budget by dust prior to any response by the climate. We calculate the forcing by first integrating an AGCM without dust through one annual cycle. The model temperature, humidity, and cloud cover are archived every five hours, and then used to calculate radiative fluxes with and without the the climatological dust distribution described in the previous section. The forcing is defined as the flux difference between the two calculations. The forcing is sensitive to the albedo of the column underlying the dust layer, which requires an accurate simulation of clouds. The GISS AGCM has been shown to be one of the most accurate in model comparisons to observed cloudiness and its variability (Weare et al. 1995), along with cloud forcing (Cess et al. 1995).

Although the forcing is strictly calculated independent of any climate response, the distinction between the two is imprecise. The dust distribution is a function of climate, which in turn responds to the dust radiative forcing. In practice, one measure of the interaction is the change in the dust load by this forcing. Perlwitz et al. (2001) calculate that the inclusion
of dust radiative forcing reduces the AGCM dust load by about 15%. Because this reduction is a small fraction of the total load, we ignore this ambiguity in the forcing calculation.

The global and annual average radiative forcing is shown in Table 3 (center column). Dust attenuates the incident solar flux through a combination of absorption and reflection. At thermal wavelengths, dust acts like a greenhouse gas, absorbing radiation from the surface, reducing outgoing emission at the top of the atmosphere, while increasing the flux back toward the surface. Because most dust particles are small compared to thermal wavelengths — especially beyond a few hundred kilometers from the source region (Duce 1995) — solar forcing dominates the thermal component on a global scale, even though dust particles are highly absorbing at longer wavelengths (Sokolik and Toon 1999). The net forcing at the surface is -1.64 Wm$^{-2}$. At TOA, the net forcing is negative but nearly zero; negative forcing by shortwave reflection is almost completely offset by absorption at solar and thermal wavelengths. Hansen et al. (1997) estimate 0.91 as the global value of $\omega_0$ at 0.55 $\mu$m separating positive from negative radiative forcing at TOA by an aerosol layer. This is consistent with our small TOA forcing and the global average $\omega_0$, equal to 0.905 within the 0.20 to 0.77 $\mu$m spectral band (Fig. 5b). The difference between the surface and TOA forcing is proportional to diabatic heating within the atmosphere. This heating, corresponding to roughly 0.01 K per day, is dominated by the solar component. Cooling at thermal wavelengths — a signature of the greenhouse effect — is a comparatively small offset. Fig. 6 depicts the vertical distribution of radiative forcing averaged over the spatial extent of the dust layer, defined as the region where the column burden exceeds 50 mg m$^{-2}$ (c.f. Fig. 3). Within the dust layer, the heating rate is on the order of 0.1 K per day (Fig. 6d), offsetting by roughly 10% the tropic-wide radiative cooling rate (Holton 1992), whose magnitude determines the strength of the large-scale tropical circulation (Pierrehumbert 1995). The heating extends throughout the troposphere, despite the concentration of dust near the surface. The extension of the heating to well above the dust layer is partly due to decreasing density and thermal inertia with height, as well as reflection of sunlight at the top of the dust layer, which reduces the downward flux and associated forcing below. Given the small TOA forcing and comparatively large surface value, the effect of dust is to displace radiative heating from the surface into the atmosphere (Miller and Tegen 1999; Ramanathan et al. 2001).
The model’s aerosol burden is at the low end of current estimates. The burdens listed in Table 1 are larger by as much as a factor of two and a half. To examine the sensitivity of our global forcing estimate to a higher aerosol amount, we increased our baseline dust burden (Fig. 3) by a globally uniform factor and recomputed the forcing. Fig. 7 shows the global average surface and TOA forcing, normalized by their baseline values (Table 3, center column), for global loads ranging from half the baseline estimate to five times this amount. To a good approximation, the forcing is linear in this range for all seasons. If the present-day global burden is as large as calculated by Ginoux et al. (2001), then the actual surface forcing is over two times our model’s estimate. In this case, the atmospheric radiative heating within the dust layer (c.f. Fig. 6d) is as large as a quarter of the tropical radiative cooling rate.

Figs. 8a and 9a show the geographic distribution of surface forcing for NH winter and summer, respectively. Regional variations in the surface forcing largely follow variations in the column dust load. Where the dust layer is optically thick, the surface forcing is generally the most negative. The TOA forcing is modulated by regional variations in the planetary albedo (not shown), which measures the combined reflectivity of the atmospheric column and underlying surface (Andreae 1995). During NH winter (Fig. 8b), the negative forcing is largest where the dust cloud extends over the dark, verdant coast of the Gulf of Guinea and the eastern Atlantic. During NH summer (Fig. 9b), when the plume is centered farther north over the comparatively bright Sahara, the TOA forcing in this region changes sign to positive. Because the forcing at TOA is small compared to the surface value, regional variations in atmospheric radiative heating generally follow surface variations, which in turn are dominated by regional variations in the column dust load (Fig. 8c and 9c).

Surface forcing of \(-14\,\text{W}\,\text{m}^{-2}\) was estimated by the Indian Ocean Experiment (between January and April of 1998) and can be compared to the model value (Fig. 8a). Over the northwestern Indian Ocean, the model forcing is negative and of magnitude \(1\text{–}2\,\text{W}\,\text{m}^{-2}\). Ramanathan et al. (2001) estimate that 10% of the aerosol burden is due to soil dust, and if we assume the different aerosol types contribute linearly to the forcing (Satheesh et al. 1999; Podgorny et al. 2000), then our model is in agreement.

Our model’s overestimate of Australian emission, noted in the previous section, may bias
our estimate of the global forcing. To quantify this effect, we calculate the contribution to the forcing by each source region (Table 4). The regions are demarcated using the $10 \text{mg m}^{-2}$ contour (Fig. 3); the Australian plume is spatially distinct at this threshold. We divide the Asian and NH African/Arabian plumes as in the previous section (Fig. 4). The downstream boundary of the Asian plume is set at $90^\circ \text{W}$, and African dust is distinguished from sources to the west using the minimum in dust load over the Atlantic. Because dust from each source region extends beyond its prescribed boundary, its contribution listed within the table is an underestimate. However, for the case of Australian dust, we found that the TOA value is insensitive to thresholds between 10 and $20 \text{mg m}^{-2}$, suggesting that this overestimate is slight. According to Table 4, Australian dust contributes just over ten percent of the global forcing at the surface. The contribution to the global TOA value is larger, yet still under a quarter, in part because forcing beneath the extensive Asian plume largely cancels due to the compensating albedo effect of dark ocean and bright clouds beneath its trajectory. Even if the actual Australian emission were half the model value, our global forcing estimates would be only slightly reduced.

Forcing has been calculated assuming a uniform index of refraction for all source regions, based upon laboratory measurements of far-travelled Saharan dust. These particles consist mainly of clay and quartz, with enhanced solar absorption due to trace amounts of the iron oxide hematite (Chester and Johnson 1971; Volz 1973). The Sahara dominates emission throughout the year (Fig. 1), and downwind of the Atlantic coast, the mineralogical composition is nearly invariant (Patterson et al. 1977). Thus, the use of far-travelled Saharan dust worldwide is a reasonable first approximation that is especially attractive because of the added computational cost of 4 tracer variables needed to represent the size distribution of each additional mineral type.

Nonetheless, other source regions have distinct optical properties (Sokolik et al. 1993). Even downwind of the Sahara, variations in the aerosol mineralogy are well-known (Carlson and Prospero 1972). Solar absorption can be reduced by condensation of water on the surface of dust particles (Sokolik et al. 1993), as clouds form within the aerosol layer. In situ sun photometer and satellite measurements suggest that solar absorption by Saharan dust is
smaller than that implied by laboratory measurements of the particle index of refraction (Kaufman et al. 2001; Sinyuk et al. 2003). To quantify this discrepancy in our model, we compare our size-integrated single scatter albedo (which varies inversely with particle absorption) to AERONET measurements taken downwind of the Sahara at Capo Verde during NH summer (Kaufman et al. 2001). At this location, the model $\omega_0$ is just over 0.90 in the spectral band between 0.20 and 0.77 $\mu$m (Fig. 5b). The AERONET values at comparable optical thicknesses are 0.905 at 0.44 $\mu$m and 0.935 at 0.67 $\mu$m with an estimated uncertainty of 0.02. While our model is marginally in agreement if the measurement uncertainty is taken into account, this agreement may result from our unusually large fraction of clay aerosol. Were our size distribution weighted toward larger particles, as in some models (Table 1), our excess absorption with respect to AERONET would be larger (Fig. 5a).

Given the global variations in particle mineralogy, along with variations in water coating along the trajectory due to cloud processing, we calculate the forcing sensitivity to variations in the optical properties. We carry out two additional experiments where the single scattering albedo $\omega$ (equal to the ratio of scattering to total extinction for a single particle) is either increased or else decreased by 10%. In either case, total extinction, corresponding to scattering plus absorption, is held constant. A 10% increase in $\omega$, corresponding to an increase in scattered radiation at the expense of absorption, might represent dust particles comprised solely of clay or quartz, whose solar absorption is nearly zero. Conversely, a 10% decrease in $\omega$ can represent Saharan particles whose absorptivity is increased by the aggregation of a small amount of additional hematite (Sokolik and Toon 1999). Note that a 10% increase or decrease in $\omega$ is within the range measured globally (Sokolik et al. 1993), and used in aerosol radiation models (Sokolik and Toon 1996). This variation also bounds the reduction to solar absorption inferred by Kaufman et al. (2001) and Sinyuk et al. (2003).

The sensitivity of the global and annual average forcing to particle absorption is given by Table 3. The sensitivity of the forcing is large in comparison to the modest changes in the single scattering albedo. For example, the 10% decrease in $\omega$ results in a 50% increase in the surface forcing, while the forcing at TOA even changes sign. This large sensitivity is due to the domination of the solar forcing over the longwave component. At solar wavelengths,
absorption by far-travelled Saharan dust is small, and \( \tau \) is near 0.9 (Fig. 5a). Thus, a 10% increase reduces particle absorption (proportional to \( 1 - \tau \) for \( \tau \) near unity) to nearly zero, while a corresponding decrease doubles the absorption. In contrast, \( \tau \) at thermal wavelengths is closer to 0.5, so that a 10% change has little effect.

Column heating by dust shows a similarly large sensitivity to the particle absorption. Figs. 10 and 11, show regional profiles of the forcing as perturbations to the net flux (in Wm\(^{-2}\)) and diabatic heating (in K per day), respectively. The regions correspond to the vertical profiles of dust mixing ratio in Fig. 4. In all regions, the heating shows the same sensitivity as the surface forcing, doubling for a 10% reduction in \( \tau \). For more absorbing particles, the heating rate ranges from 0.2 to 0.3 K per day, even in the upper troposphere. Again, this is a non-negligible offset to the tropical radiative cooling rate of 1 K per day (Holton 1992).

The forcing is also uncertain as a result of the contrasting size distributions calculated by current models (Table 1, Fig. 2), where the ratio of silt particles to clay varies over a factor of four. We calculate the sensitivity of the surface forcing to particle size by computing the forcing for each size category separately. This decomposition assumes that the total forcing is equal to the sum of the contributions from each category. In fact, the sum of the forcings calculated separately slightly exceeds the forcing of the entire distribution (Table 5). Taken together, each size category reduces the radiation incident upon other size particles. However, for the present dust load, this difference is very small — a result of the modest magnitude of the present-day dust burden.

The contribution of each size category to the global and annual forcing is shown in Fig. 12. The contribution reflects the global abundance of each size category (Fig. 2), along with the efficiency of forcing by each category per unit mass — denoted in Fig. 13 by diamonds for the current model distribution (Fig. 2a) and by crosses for the distribution used by Miller and Tegen (1998) (Fig. 2b). The similar estimates of forcing per unit mass for the two size distributions, despite different geographical distributions, suggests that this quantity is robust. According to Mie theory, extinction by a single particle is negligible at wavelengths larger than the particle radius, but is roughly constant for shorter wavelengths.
Moreover, the number of particles decreases per unit mass of aerosol as the radius increases. Thus, solar forcing per unit mass peaks for particles with radii comparable to the wavelength of maximum solar irradiance (around 0.5 \( \mu \text{m} \)), and thermal forcing increases slowly with particle size for the size range considered here. This accounts for the forcing efficiency shown in Fig. 13. At TOA, where solar absorption and reflection make offsetting contributions to the forcing, thermal forcing by larger particles can exceed the solar residual to create net positive forcing. In contrast, solar absorption and reflection both create negative forcing at the surface, and the compensating positive forcing at thermal wavelengths is relatively small, even for larger particles. Because of this cooperation, the sign of the forcing is less sensitive to the size distribution at the surface than at TOA.

To estimate the effect of current uncertainties in the size distribution upon the global forcing, we take two size distributions with identical total loads but varying ratios of silt to clay, and multiply the mass of each size category by the forcing per unit mass computed by the AGCM (Fig. 13, diamonds). The first size distribution is taken from the AGCM (Fig. 2a), and has a ratio of silt to clay aerosol of 0.59. The second size distribution has a ratio of 1.41, and is taken from Fig. 2b (Miller and Tegen 1998). The latter ratio is similar to that computed by Ginoux et al. (2001), whose distribution compares well with that retrieved by AERONET. These two ratios of silt to clay are representative of the range among models in Table 1. The forcing is listed in Table 6. As the silt burden increases, both the TOA and surface forcing tend toward positive values due to the longwave effect of the larger silt particles. However, despite the large increase in silt fraction, the surface forcing and atmospheric radiative heating remain within about 10% of the value calculated from the AGCM distribution.
4 Dust Forcing of the Hydrological Cycle

In a radiative equilibrium atmosphere, negative forcing at the surface by dust is balanced by a reduction in upward thermal radiation, caused in part by a decrease in ground temperature. The effect of atmospheric dynamics is to allow the surface to respond alternatively by reducing its turbulent fluxes of sensible and latent heat into the atmosphere. The latter reduces precipitation, which feeds back upon the dust load by reducing the efficiency of wet deposition. In this section, we examine the perturbation of the AGCM’s hydrologic cycle in response to radiative forcing by dust. The effect of dust upon climate is assessed by comparing two experiments with the mixed-layer AGCM, one with radiatively active dust (whose dust distribution was used to compute the forcing in the previous section), and one where radiative forcing by dust is omitted. More precisely, the response is defined as the difference between the climatological values of a variable in the two experiments, where averages are constructed over the final thirty-one years of each simulation, following a nineteen year spin-up. Because the surface forcing is especially sensitive to absorption by the dust particles, we do two additional experiments, where the particle single scatter albedo is increased or else decreased by 10%.

In each experiment, the surface energy balance is written as:

\[ R_{SW} = R_{LW} + LE + S, \]  

(2)

where \( R_{SW} \) is the net shortwave radiation into the surface, \( R_{LW} \) is the net longwave radiation from the surface into the atmosphere, \( LE \) is the turbulent flux of latent heat leaving the surface, equal to the latent heat of vaporization times the evaporation rate, and \( S \) is the turbulent sensible heat flux. If we perturb the surface by adding radiative forcing \( F_{SRF} \) due to dust, then the budget of the anomalies (denoted by \( \delta \)) is:

\[ F_{SRF} + \delta R_{SW} = \delta R_{LW} + \delta (LE) + \delta S. \]  

(3)

[Note the distinction between the surface radiative forcing \( F_{SRF} \) — calculated offline in the previous section using the AGCM dust distribution along with temperature, humidity, and cloud cover from an experiment without dust — and the surface radiative flux anomaly]
\[ \delta R_{SW} - \delta R_{LW}, \] which is computed allowing the climate of the AGCM with dust to respond to its radiative forcing.] For the sake of discussion, we assume that the forcing is entirely at solar wavelengths, and that the radiative response is restricted to thermal wavelengths. (We examine the validity and consequence of these approximations below.) Thus:

\[ \delta R_{LW} + \delta(LE) + \delta S = SRF, \]  

where the surface forcing \( SRF \) is negative as calculated in the previous section. Each of the surface fluxes on the left-hand side of (4) is shown in Fig. 14. The global and annual average response of the surface fluxes is summarized in Table 7 (center column). On a global scale, dust radiative forcing at the surface is balanced predominately by a reduction in the latent heat flux, and secondarily, by the sensible heat flux. The reduction in the latent heat flux is largest over oceanic regions downwind of dust source regions — such as the eastern subtropical Atlantic, the Arabian Sea, and the southeast Indian Ocean — along with continental areas with abundant soil moisture, such as the West African coast. In comparison to the anomalous turbulent fluxes, the direct radiative response through the net upward thermal flux anomaly is small. Our attribution of the net surface longwave anomaly as a response to dust assumes that the surface longwave forcing is negligible. While this may not be the case near source regions (Claquin et al. 1998), where the fraction of large particles is greatest, it is approximately true over the global extent of the dust layer (Table 3). A non-zero, positive longwave component of the forcing corresponds to a net upward thermal radiative response that is slightly greater than the anomaly defined by the difference between the model values.

Although the surface latent heat flux is reduced globally by dust (Coakley and Cess 1985; Miller and Tegen 1998), this effect decreases with increasing particle absorption (Table 7, right column). This is counterintuitive, because surface radiative forcing increases sharply with particle absorption (Table 3). The explanation is that despite their global decrease, evaporation and rainfall are increased by dust over desert regions, such as the Sahara and Mojave, even with the reduction of incident sunlight (Fig. 14a). This increase is largest for the experiment with the most absorbing particles (Fig. 16), accounting for the otherwise unexpected decrease in global evaporation with decreasing \( \omega \).
To understand the contrasting evaporative response to dust, note that atmospheric radiative cooling, which determines the rate of subsidence in the absence of dust, is smallest over deserts and dry continental interiors (Fig 15a). This cooling is weak because of the small column water vapor, which allows efficient emission of longwave radiation by the surface directly to space, without absorption and reemission within the intervening atmosphere. Over deserts, the column radiative forcing by dust (Figs. 8c and Figs. 9c) causes the greatest fractional reduction of subsidence. Because subsidence inhibits convection and rainfall by mixing dry upper tropospheric air into the boundary layer, a decrease in the rate of descent increases the likelihood of precipitation (Fig. 17).

The evaporative increase over the Sahara and Mojave deserts is related to the rainfall anomaly in these regions (Fig. 17b). Radiative heating within the dust layer changes the column diabatic heating, causing a spatial reorganization of rainfall, and in particular, an extension of the ITCZ into the western Sahara. The precipitation increase with particle absorptivity (Fig. 17c) is related to the increased offset of radiative cooling and reduction of local subsidence as $\varpi$ decreases.

This reduction depends upon the magnitude of the atmospheric radiative heating anomaly, defined as:

$$
\delta H \equiv F_{TOA} - F_{SRF} + \delta LP + \delta S + \delta R_{TOA} - \delta R_{SRF}.
$$

where $F_{TOA}$ and $\delta R_{TOA}$ are the forcing and radiative anomaly at TOA, respectively, and $P$ is the column precipitation. If rainfall is supplied by evaporation beneath the dust layer, then $\delta(LP) = \delta(LE)$, and we can use the anomalous surface energy balance (3) to write:

$$
\delta H \equiv F_{TOA} + \delta R_{TOA}.
$$

According to (6), $\delta H$ is largest over deserts, where the TOA forcing is positive (Figs. 8a and 9a), due to the bright underlying surface compared to oceanic or vegetated regions. This contrast, along with the larger dust load and smaller background radiative cooling over dry regions, means that dust is more likely to create rainfall over a desert than over the ocean.

Menon et al. (2002) show a similar increase in precipitation beneath an absorbing aerosol layer, although their prescribed SST experiment lacks a surface energy constraint over the
oceans. Miller and Tegen (1998) show that inclusion of the surface energy budget reduces the anomalous diabatic heating and precipitation: compare eqs. (5) and (6).

Our analysis of the anomalous surface energy balance neglects the radiative response at solar wavelengths: for example, by a reduction in cloud cover that offsets the reduction in incident sunlight. We measure the importance of this effect by comparing the surface radiative forcing to the net solar flux at the surface, as in Fig. 18. While comparison is complicated by internal variability of the model, the forcing and solar anomaly are almost identical beneath the dust layer. Over the Sahara, the solar anomaly is slightly more negative than the forcing, indicating an increase in cloud cover associated with the precipitation anomaly (Fig. 19; similar cloud cover anomalies are found at higher levels.) Over the ocean, the anomalous cloud cover is less spatially coherent and more difficult to interpret, although it generally increases with absorptivity. As noted above, the rate of subsidence, which flushes lower levels with dry upper tropospheric air and thus inhibits cloud formation, is smallest for the most absorbing particles. The increase in cloud cover with increasing aerosol absorption is counter to the ‘semi-indirect’ effect described by Hansen et al. (1997) in the same AGCM, where heating of an aerosol layer leads to a decrease in relative humidity and cloud cover. In that study, aerosol heating was confined to a single model layer, rather than the deeper layer indicated in Fig. 11, with a correspondingly smaller reduction of dry air subsiding into the boundary layer.

The AGCM demonstrates that while dust generally reduces evaporation and rainfall beneath the dust layer, these quantities may actually increase over dry regions, where dust is originally emitted. This represents a negative feedback to desertification, where dessication of the soil and subsequent dust emission cause an expansion of the rainy zone into the desert, inhibiting further emission. In our model, Saharan rainfall increases by 0.3 mm/day (Fig. 17b), which corresponds to roughly 10 cm/year. This is about 5% of the annual average rainfall within the ITCZ to the south (Fig 15b), but a more substantial perturbation to Saharan rainfall — it could be larger given evidence that our model underestimates the actual dust load. Note that soil particles from the Sahel are enriched in hematite, a mineral which efficiently absorbs at solar wavelengths (Sokolik and Toon 1999; Claquin et al. 1999), which
would increase the column warming and the precipitation anomaly (Fig. 17c). The negative feedback to desertification opposes the reduction in rainfall by dust due to microphysical effects, where dust increases the available cloud condensation nuclei (CCN), reducing the average cloud droplet size. Rainfall is decreased as a consequence of the increased number of collisions required to form a raindrop (Nicholson 2000; Rosenfeld et al. 2001). Modeling of this aerosol indirect effect of dust is currently restricted to case studies within a limited domain, and is not included in our AGCM.

Despite the increase in desert rainfall by dust, the global effect of its surface forcing is to reduce rainfall. This in turn reduces wet deposition, as shown in Fig. 20. This reduction is calculated by comparing the deposition rate between AGCM experiments with and without dust radiative forcing. Dust is a prognostic variable in both experiments, but alters the climate — and in turn is altered by these climate anomalies — only in the former. [In contrast, the dust distribution was specified externally in the experiments of Miller and Tegen (1998), and was unchanged, despite the climate anomalies it forced.] The global and annual mean wet deposition and wet deposition lifetimes are listed in Table 8. Dust radiative forcing increases the wet deposition lifetime from 12.5 to 12.8 days, compared to the experiment where dust is present but has no radiative effect — with an additional increase of over a full day to 13.9 for more absorbing particles. While dust may increase rainfall and inhibit emission over deserts, its global effect is to increase the dust load; it represents a local negative feedback to desertification but a positive feedback to the global dust burden.

In the current climate, this positive feedback is modest. For fixed emission and dry deposition, the dust load is increased by only 1.0% due to the lengthened wet deposition lifetime. This is comparable to the 1.3% reduction of global evaporation and precipitation, based upon the -1.13 W m$^{-2}$ global anomaly of the surface latent heat flux (Table 7). Nonetheless, the actual percentages could be between two and three times larger if the present-day dust burden is as large as estimated by Ginoux et al. (2001).

The reduction in evaporation and increase in the wet deposition lifetime by dust radiative forcing were greater during the Last Glacial Maximum (LGM), when dust deposition increased by an order of magnitude in deep sea sediments (Rea 1994) and high latitude ice
cores (Petit et al. 1990; Biscaye et al. 1997; Reader et al. 1999; Mahowald et al. 1999; Kohfeld and Harrison 2001). Estimates of the LGM dust burden range from one-and-a-half (Lunt and Valdes 2002) to two-and-a-half (Mahowald et al. 1999) to three (Werner et al. 2003) times the present-day value. If we assume our present-day burden is an underestimate by a factor of two, and that the LGM burden is two times greater, then assuming a corresponding increase in surface radiative forcing, the reduction in evaporation due to the radiative effect of dust alone is just over 5%. Yung et al. (1996) hypothesize that the glacial dust load was larger due to a reduction in rainfall by the cooler climate together with a corresponding decrease in wet deposition. Bush and Philander (1999) calculate this reduction to be 11%, so that the additional reduction due to dust could be a significant source of aerosol. Decreased precipitation by dust can also reduce vegetation, and increase potential source areas for emission, another positive feedback to the dust load. Mahowald et al. (1999) show that the decrease in vegetation by the reduction in temperature, precipitation, and atmospheric CO$_2$ during the LGM result in substantially greater mobilization of dust.

Extrapolation of the current forcing to LGM values is highly uncertain (Harrison et al. 2001), due to the greater fraction of the LGM dust load at high latitudes, where there are large changes in surface albedo due to the extensive continental ice sheets and reduced vegetation. Both of these effects would combine to make the surface brighter, which would increase the surface forcing and the associated reduction in evaporation and precipitation beyond that of our estimate. Despite this caveat, our extrapolation suggests that the reduction of the LGM hydrologic cycle by dust radiative forcing is large enough — compared to the reduction due solely to the cooler climate — to justify its explicit calculation with an AGCM.
5 Conclusions

Although its importance to climate has long been recognized (Coakley and Cess 1985), radiative forcing by dust aerosols remains highly uncertain. While aerosol forcing is typically characterized by its TOA value (IPCC 2001), the surface forcing is also fundamental to the climate response, especially for absorbing aerosols like dust, where forcing at the two levels may be quite different. This difference represents radiative heating displaced from the surface into the dust layer (Miller and Tegen 1999; Ramanathan et al. 2001).

Using a global dust distribution derived by the NASA GISS AGCM, we estimate the surface forcing and atmospheric radiative heating, emphasizing their sensitivity to aspects of the dust distribution that vary widely among models, and are poorly constrained by observations. Global surface forcing is negative due to absorption and reflection of incident solar radiation within the overlying dust layer, where radiative heating is roughly 10% of the tropicwide radiative cooling rate. While the dust mixing ratio is largest near the surface, radiative heating remains substantial throughout the troposphere. This is important to the climate response, which is largest when the radiative forcing is separated from the boundary layer (Miller and Tegen 1999). In this case, dust heating is balanced in the tropics by reduced adiabatic subsidence, or by reduced eddy heat convergence in mid-latitudes, which leads to a change in surface air temperature. In contrast, radiative heating within the boundary layer is compensated mainly by a change in the surface sensible heat flux via the ground temperature, which allows the air temperature to remain relatively unperturbed.

Our forcing estimate is most sensitive to the current uncertainty in aerosol burden. The global burden computed by the AGCM is just under 15 Tg. This is at the low end of recent model estimates that can exceed ours by more than a factor of two, while remaining consistent with observed values of column amount and surface concentration (Ginoux et al. 2001; Tegen et al. 2002; Luo et al. 2002; Zender et al. 2003). Because the global forcing is nearly linear within this range, the actual forcing and atmospheric radiative heating are possibly over a factor of two greater than our model estimate. Observational estimates of the aerosol burden have a similar uncertainty: satellite inferences of the total aerosol optical thickness vary by a factor of two over the oceans (Myhre et al. 2003). The particle size distribution also varies
greatly among models. The ratio of silt particles to clay varies fourfold, although the surface forcing and atmospheric radiative heating vary only by 10% over this range.

Dust radiative forcing is calculated by assuming particle optical properties that are globally uniform, taken from laboratory measurements of far-traveled Saharan dust (Volz 1973; Patterson et al. 1977). Solar absorption by Saharan dust remains under debate. The absorption inferred from sun photometers is substantially smaller (Kaufman et al. 2001; Sinyuk et al. 2003). In contrast, Weaver et al. (2002) find that our adopted optical properties lead to the best agreement of dust radiative forcing with satellite measurements. In addition to the uncertainty of measured absorption, the actual forcing will depart from our calculated value due to geographic variations in the mineralogy of source regions, along with the presence of water on the aerosol surface. We calculate that a 10% increase in the particle single scatter albedo $\omega$ (corresponding to reduced absorption) halves the surface forcing, and reduces the associated radiative heating to nearly zero. This change in $\omega$ bounds the reduction in solar absorption proposed for Saharan dust by the \textit{in situ} measurements of Sinyuk et al. (2003). Conversely, a 10% decrease in particle absorption nearly doubles the surface forcing and atmospheric radiative heating.

Regional variations in mineralogical composition of soil particles are fundamental to the forcing and regional climate response, along with the associated anthropogenic ‘fingerprint’ of soil dust that must be distinguished from that due to greenhouse gases. We are currently introducing distinct aerosol mineralogies using recent data sets by Claquin et al. (1999), describing the regional distribution of various mineral types, and Sokolik and Toon (1999), who tabulate wavelength-dependent indices of refraction for these minerals. Biogeochemical processes are sensitive to the precise mineral content of the deposited aerosol. Deposition of iron-rich minerals, such as hematite, fertilize the ocean mixed layer, increasing the productivity of marine ecosystems and the ocean uptake of CO$_2$ (Martin 1991; Bishop et al. 2002). By this mechanism, dust is hypothesized to have modulated the atmospheric CO$_2$ concentration during the last glacial cycle (Archer et al. 2000; Broecker 2000). Certain minerals also supply crucial nutrients to soils far downwind of the source region (Swap et al. 1992; Chadwick et al. 1999).
Through radiative forcing at the surface, dust alters the hydrologic cycle. The reduction in sunlight beneath the dust layer is balanced globally by diminished evaporation, with a secondary reduction in the surface sensible heat flux (Coakley and Cess 1985). Despite its global reduction, we find that evaporation increases along with rainfall over the Sahara. We suggest three reasons for this contrast. First, in the absence of dust, radiative cooling is smallest over deserts, where the column moisture is small, and surface emission of longwave radiation directly to space is efficient. Consequently, adiabatic subsidence, which balances radiative cooling and inhibits convection by flushing the boundary layer with dry upper level air, is weak. Second, radiative heating of the column by dust, which reduces the rate of subsidence, is large over the bright desert surface. Finally, the column dust burden is largest over arid regions, where the absence of soil moisture leaves particles susceptible to wind erosion. Dust radiative heating of the column, combined with weak background subsidence, favors precipitation. This represents a negative feedback to dust emission resulting from desertification. The strength of this feedback depends upon the column radiative heating by dust, which increases with particle absorption. This accounts for the reduction in magnitude of the global evaporative anomaly as particle absorption is increased, despite the increase in the magnitude of surface forcing. The local negative feedback opposes the decrease in rainfall due to the aerosol indirect effect upon cloud microphysics (Nicholson 2000; Rosenfeld et al. 2001). Quantifying the relative importance of these two feedbacks will require more precise measurements of particle absorption, along with a better understanding of the interaction between dust and cloud microphysics on a global scale.

Despite the increase in desert precipitation, the global reduction of rainfall by dust reduces the wet deposition efficiency. Thus, while radiative forcing by dust creates a local negative feedback to desertification, it increases the aerosol burden globally. For the current climate, this effect is modest, as dust reduces precipitation by only a percent or two. The associated wet deposition lifetime increases from 12.5 to 12.8 days, although this lifetime is as long as 13.9 days if the particle absorption is increased by 10%. However, for glacial climates, when the inferred load is substantially larger (Kohfeld and Harrison 2001), we estimate that the reduction in precipitation could exceed 5%, roughly half of the precipitation decrease resulting simply from the colder climate (Bush and Philander 1999). This suggests
that dust radiative forcing contributes non-negligibly to the reduction of the hydrologic cycle during glacial climates. In addition to its effect upon the aerosol lifetime, the reduction in rainfall has the potential to reduce vegetation, expanding potential regions of dust emission. Mahowald et al. (1999) show that the contraction of vegetated areas by the colder and drier LGM climate contributes substantially to the greater mobilization of dust. The ability of dust to reduce vegetative cover is a potentially large positive feedback, given that the reduction of rainfall by dust is roughly half that due to all other processes. Our estimated radiative forcing by dust for a glacial climate, based upon extrapolation from the present, is highly uncertain, due to the substantial differences in the geographic distribution of dust as well as the surface albedo between the two climates (Kohfeld and Harrison 2001). However, the reflective continental ice sheets and diminished forest area might actually increase the surface forcing beyond that suggested by our estimate. Despite its uncertainty, our estimated reduction in the hydrologic cycle by dust during glacial times is large enough to suggest that its explicit calculation is worthwhile.

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References


34


### Tables

<table>
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<th>Emission (Tg yr(^{-1}))</th>
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Table 1: Comparison of global dust emission, load, proportion of clay to silt aerosol, lifetime, between the present model and recent studies of dust transport. Clay particles are defined with radii less than 1 µm, except for the studies of Luo et al. (2002) and Zender et al. (2003) where the radii extend up to 1.25 µm.
Table 2: Global and annual average emission (mg m\(^{-2}\) d\(^{-1}\)) and aerosol load (mg m\(^{-2}\)) of clay and silt particles, along with their calculated aerosol lifetime (d).

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Table 3: Annual and global average radiative forcing by dust aerosols (Wm\(^{-2}\)), with the net forcing (positive downward) decomposed into solar (SW) and longwave (LW) contributions. The baseline case is denoted by 1.0 \(\times\) \(\varpi\). The experiments with more absorbing and more reflecting particles are denoted by 0.9 \(\times\) \(\varpi\) and 1.1 \(\times\) \(\varpi\), respectively.

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<td><strong>Surface:</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NH Africa/Arabia</td>
<td>-0.75</td>
<td>-1.18</td>
<td>-1.08</td>
</tr>
<tr>
<td>Asia</td>
<td>-0.03</td>
<td>-0.47</td>
<td>-0.60</td>
</tr>
<tr>
<td>Australia</td>
<td>-0.44</td>
<td>-0.07</td>
<td>-0.04</td>
</tr>
<tr>
<td>Other</td>
<td>-0.27</td>
<td>-0.25</td>
<td>-0.33</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>-1.49</td>
<td>-1.97</td>
<td>-2.05</td>
</tr>
<tr>
<td><strong>TOA:</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NH Africa/Arabia</td>
<td>-0.22</td>
<td>-0.15</td>
<td>-0.04</td>
</tr>
<tr>
<td>Asia</td>
<td>0.00</td>
<td>-0.02</td>
<td>-0.02</td>
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<tr>
<td><strong>Total</strong></td>
<td>-0.32</td>
<td>-0.16</td>
<td>-0.12</td>
</tr>
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Table 4: Contribution of the NH African/Arabian, Asian, and Australian source regions to the total global average forcing by dust aerosols (Wm\(^{-2}\)).

<table>
<thead>
<tr>
<th></th>
<th>Baseline</th>
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<tr>
<td><strong>TOA:</strong></td>
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<td></td>
</tr>
<tr>
<td>DJF</td>
<td>-0.32</td>
<td>-0.33</td>
</tr>
<tr>
<td>MAM</td>
<td>-0.16</td>
<td>-0.17</td>
</tr>
<tr>
<td>JJA</td>
<td>-0.12</td>
<td>-0.13</td>
</tr>
<tr>
<td>SON</td>
<td>-0.15</td>
<td>-0.16</td>
</tr>
<tr>
<td>ANN</td>
<td>-0.18</td>
<td>-0.20</td>
</tr>
<tr>
<td><strong>Surface:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DJF</td>
<td>-1.49</td>
<td>-1.52</td>
</tr>
<tr>
<td>MAM</td>
<td>-1.97</td>
<td>-1.99</td>
</tr>
<tr>
<td>JJA</td>
<td>-2.05</td>
<td>-2.07</td>
</tr>
<tr>
<td>SON</td>
<td>-1.04</td>
<td>-1.05</td>
</tr>
<tr>
<td>ANN</td>
<td>-1.64</td>
<td>-1.66</td>
</tr>
</tbody>
</table>

Table 5: Global and seasonal average radiative forcing calculated by the baseline experiment, and the sum of the forcings for each size class calculated separately.
### Table 6: Global and annual average radiative forcing by dust aerosols (Wm\(^{-2}\)), for different ratios of silt to clay aerosol. The forcing is calculated by multiplying the mass corresponding to each category by the global and annual average forcing per unit mass shown in Fig. 12.

<table>
<thead>
<tr>
<th></th>
<th>Silt/Clay = 0.6</th>
<th>Silt/Clay = 1.4</th>
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<tr>
<td><strong>TOA:</strong></td>
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<tr>
<td>clay</td>
<td>-0.26</td>
<td>-0.21</td>
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<tr>
<td>silt</td>
<td>0.06</td>
<td>0.05</td>
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<tr>
<td>total</td>
<td>-0.20</td>
<td>-0.16</td>
</tr>
<tr>
<td><strong>Surface:</strong></td>
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<tr>
<td>clay</td>
<td>-1.30</td>
<td>-0.84</td>
</tr>
<tr>
<td>silt</td>
<td>-0.36</td>
<td>-0.62</td>
</tr>
<tr>
<td>total</td>
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<td>-1.47</td>
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<tr>
<td><strong>Atmos. Heating:</strong></td>
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<td></td>
</tr>
<tr>
<td>clay</td>
<td>1.04</td>
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</tr>
<tr>
<td>silt</td>
<td>0.42</td>
<td>0.67</td>
</tr>
<tr>
<td>total</td>
<td>1.46</td>
<td>1.31</td>
</tr>
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</table>

Table 7: Annual average surface forcing and surface flux anomalies (Wm\(^{-2}\)) for AGCM experiments with reflecting dust particles (1.1 × \(\varpi\)), far-travelled Saharan particles (1.0 × \(\varpi\)), and more absorbing particles (0.9 × \(\varpi\)).

<table>
<thead>
<tr>
<th></th>
<th>1.1 × (\varpi)</th>
<th>1.0 × (\varpi)</th>
<th>0.9 × (\varpi)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface Forcing</td>
<td>-1.01</td>
<td>-1.64</td>
<td>-2.54</td>
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<tr>
<td>Net Downward Solar</td>
<td>-0.84</td>
<td>-1.84</td>
<td>-3.35</td>
</tr>
<tr>
<td>Latent Heat</td>
<td>-1.30</td>
<td>-1.13</td>
<td>-1.04</td>
</tr>
<tr>
<td>Sensible Heat</td>
<td>-0.04</td>
<td>-0.48</td>
<td>-1.21</td>
</tr>
<tr>
<td>Net Upward Longwave</td>
<td>0.57</td>
<td>-0.12</td>
<td>-1.05</td>
</tr>
<tr>
<td></td>
<td>-0.77</td>
<td>-1.73</td>
<td>-3.30</td>
</tr>
</tbody>
</table>

Table 7: Annual average surface forcing and surface flux anomalies (Wm\(^{-2}\)) for AGCM experiments with reflecting dust particles (1.1 × \(\varpi\)), far-travelled Saharan particles (1.0 × \(\varpi\)), and more absorbing particles (0.9 × \(\varpi\)).
Wet Deposition Load Wet Lifetime
\(\text{mg m}^{-2}\text{d}^{-1}\) \(\text{mg m}^{-2}\) \(\text{d}\)

<table>
<thead>
<tr>
<th></th>
<th>Wet Deposition</th>
<th>Load</th>
<th>Wet Lifetime</th>
</tr>
</thead>
<tbody>
<tr>
<td>passive dust</td>
<td>2.75</td>
<td>34.3</td>
<td>12.5</td>
</tr>
<tr>
<td>(1.1 \times \varpi)</td>
<td>2.21</td>
<td>27.6</td>
<td>12.5</td>
</tr>
<tr>
<td>(1.0 \times \varpi)</td>
<td>2.24</td>
<td>28.6</td>
<td>12.8</td>
</tr>
<tr>
<td>(0.9 \times \varpi)</td>
<td>2.10</td>
<td>29.2</td>
<td>13.9</td>
</tr>
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</table>

Table 8: Wet Deposition (\(\text{mg m}^{-2}\text{d}^{-1}\)), dust load (\(\text{mg m}^{-2}\)), and wet deposition lifetime (\(\text{d}\)) for AGCM experiments with passive dust, reflecting dust particles (\(1.1 \times \varpi\)), far-travelled Saharan particles (\(1.0 \times \varpi\)), and more absorbing particles (\(0.9 \times \varpi\)). The passive dust experiment calculates the trajectory of dust aerosols, but omits the particles’ radiative effect.
Figure Captions

Fig. 1 Emission of soil dust (in terragrams), averaged seasonally and annually. Each bar is divided into DJF (bottom, light), MAM (above, dark), JJA, (above, light), and SON (top, dark) values. At the top of each bar is the sum: the annual average. a) global sum, b) Sahara/Sahel (NH Africa), Arabia (40–60°E, 16–36°N), Central Asia (25–90°E, 36–56°N), East Asia (95–140°E, 32–52°N), North America, and Australia. The vertical lines bracketing the annual value range between one standard deviation above and below.

Fig. 2 Contribution by each particle size category to the global and annual average dust load in mg m$^{-2}$ as computed by a) the AGCM, and b) an offline tracer model as described in Miller and Tegen (1998). Each bar is divided into the contribution from the lower troposphere (light; surface–720 mb), middle troposphere (darker; 720–390 mb), upper troposphere (light; 390–150 mb), and the lower stratosphere (darkest; 150–10 mb). The size-integrated burden is indicated in the upper right corner.

Fig. 3 Column average dust load (mg m$^{-2}$). a) DJF, b) MAM, c) JJA, and d) SON. and b) JJA (338 mb).

Fig. 4 Dust mixing ratio ($\mu$g kg$^{-1}$) over Africa and the Arabian Peninsula for a) DJF, b) MAM, c) JJA, and d) SON. Also for Asia for e) MAM, f) JJA and Australia for g) SON, and h) DJF. Mixing Ratio is averaged where the column dust load exceeds 50 mg m$^{-2}$.

Fig. 5 a) Annual, global, and column average single scatter albedo for dust particles, for the six spectral bands used in the calculation of solar radiative forcing. Also shown for the shortest solar band (0.20–0.77 $\mu$m) is the single scatter albedo for each individual model size category, denoted by its equivalent radius. b) Geographic distribution of annual and column average single scatter albedo in the 0.20–0.77 $\mu$m band. The single scatter albedo is computed as the ratio of the average scattering extinction and the average total extinction.

Fig. 6 Dust radiative forcing within the dust layer (where the column burden exceeds 50 mg m$^{-2}$). The forcing is computed separately for each season and averaged over the entire year. Forcing of the a) solar, and b) thermal fluxes (W$^{-2}$). c) Net (down minus up, solar plus thermal) flux forcing (W$^{-2}$), and d) net radiative heating by dust (K per day).
**Fig. 7** Global and annual average a) TOA and b) surface forcing calculated with a rescaled AGCM dust distribution. The distribution is multiplied by a constant value (equal to 0.5, 1, 2, 3, 4, and 5) that is independent of location, and the forcing is normalized to unity when the constant equals 1.

**Fig. 8** DJF dust radiative forcing at the a) surface, b) top of the atmosphere, and c) radiative heating of the atmosphere by dust (i.e. the difference in the surface and TOA values) in W m$^{-2}$.

**Fig. 9** Same as Fig. 8 but for JJA.

**Fig. 10** Dust radiative forcing (W m$^{-2}$) over Africa and the Arabian Peninsula for a) DJF, b) MAM, c) JJA, and d) SON. Also for Asia for e) MAM, f) JJA and Australia for g) SON, and h) DJF. Mixing Ratio is averaged where the column dust load exceeds 50 mg m$^{-2}$. Triangles for more absorbing particles (0.9$x\varpi_0$), b) circles for far-travelled Saharan dust (1.0$x\varpi_0$), and c) squares for more reflecting particles (1.1$x\varpi_0$).

**Fig. 11** Same as Fig. 10, but for radiative heating of the atmosphere by dust (K per day).

**Fig. 12** Contribution by each particle size category to the global and annual average dust radiative forcing (W m$^{-2}$) at a) TOA, and b) the surface. The total indicated in each panel refers to the forcing calculated with the entire size distribution and is slightly less than the sum of the forcings calculated separately for each size class.

**Fig. 13** a) TOA, and b) surface forcing per unit mass of dust for each size category in W mg$^{-1}$, computed by dividing the forcing (Fig. 12) by the dust load (Fig. 2) for each size category.

**Fig. 14** Annual average surface flux anomaly (W m$^{-2}$) in response to surface radiative forcing by dust particles. Anomalies are defined as the difference in surface flux between experiments with and without radiatively active dust. a) latent heat flux, b) sensible heat flux, and c) net upward longwave flux.

**Fig. 15** a) Atmospheric radiative heating (W m$^{-2}$), and b) precipitation (mm/day), in the absence of dust radiative forcing.
Fig. 16  Anomalous evaporation (mm/day) for the experiment with a) more reflecting particles ($1.1 \times \omega_0$), b) far-travelled Saharan dust ($1.0 \times \omega_0$), and c) more absorbing particles ($0.9 \times \omega_0$).

Fig. 17  Same as Fig. 16, but for anomalous precipitation (mm/day).

Fig. 18  Annual average a) surface radiative forcing by dust, and b) anomalous net solar flux into the surface, both in W m$^{-2}$.

Fig. 19  Same as Fig. 16, but for anomalous low cloud cover (percent).

Fig. 20  Same as Fig. 16, but for anomalous wet deposition (mg m$^{-2}$ day$^{-1}$).
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