S1.0 Model description

The dust source mechanism follows the Dust Entrainment and Deposition Module [*Zender et al.*, 2003a] and work conducted in the offline Model of Atmospheric Transport and Chemistry (MATCH) [*Mahowald et al.*, 2002; *Luo et al.*, 2003; *Mahowald et al.*, 2003; *Mahowald and Luo*, 2003]. The sources of dust are assumed to be dry, unvegetated regions with strong winds based on our present understanding of dust generation [e.g. *Mahowald et al.,* 2005]. When the total leaf area index plus the stem area index is below 0.1, the area of the gridbox available for dust generation increases linearly with decreasing vegetation cover. The source scheme parameterizing dust entrainment into the atmosphere is described in detail in *Zender et al* [2003a].

The transported aerosols are assumed to have a sub-bin distribution based on a log-normal distribution with a mass median diameter of 3.5 μm within each bin, larger than in our previous studies following the results of *Grini and Zender* [2004]. We use four bins with the following source apportionment : 0.1-1.0 um 3.8%; 1.0-2.5 μm 11%; 2.5-5.0 μm 17% and 5.0-10.0 μm 67% (all values in diameter) in order to better match the data, as described in *Grini and Zender* [2004]. This distribution of particles into the size bins has more large particles than previous studies [e.g. *Zender et al.,* 2003a; *Mahowald et al.,* 2002]. Deposition processes include dry gravitational settling, turbulent dry deposition and wet deposition during precipitation events. Both dry depositional processes are modeled using parameterizations described in *Zender et al.* [2003a], with a mass flux advection scheme in order to parameterize vertical fall rates correctly [*Rasch et al.,* 2001; *Ginoux*, 2003]. Wet depositional processes are parameterized within the CAM3 similar to *Rasch et al.* [2001]. We assume that mineral aerosols are incorporated into clouds because mineral aerosols readily take up water [*Koretsky, et al.*, 1997], but do not link cloud precipitation processes with aerosols.

Shortwave radiative effects are calculated within CAM3 every hour. A delta-Eddington approximation is adopted for the shortwave using 19 discrete intervals [*Collins,* 1998] for each vertical layer in the model [*Collins et al.,* 2004]. Longwave effects are calculated every 12 hours in CAM3, which uses an absorptivity/emissitivity formulation for longwave heating [*Ramanathan and Downey,* 1986]. A broadband approach is used with 7 bands, which accounts for the water vapor window regions [*Collins et al.,* 2002]. The indices of refraction have been derived from *Patterson* [1981] for the visible wavelengths, *Sokolik et al.* [1993] for the near infrared and *Volz* [1973] for the infrared. The imaginary part of the visible wavelength indices of refraction were scaled to match the new observation-based estimates of *Sinyuk et al.* [2003] and *Dubovik et al.* [2002] for the region 0.33 to 0.67 μm. Scattering of longwave radiation by dust is neglected in radiative calculations. This may lead to underestimates of longwave radiative forcing by up to 50% at top of the atmosphere (TOA) and 15% at the surface (SFC) [*Dufresne et al.,* 2002].

Simulations are conducted until there is a 15 year period where globally averaged surface temperature appears to be in equilibrium. The evolution of the globally mean surface temperature with time in the simulations, along with the time period over which the means are averaged for this study are shown in Figure S1.

S2.0 Radiative Forcing

For the current climate, the companion paper [Yoshioka et al, submitted], discusses in some detail comparisons to previous studies. Because this discussion is important to this paper, we briefly review some of the results from Yoshioka et al. [submitted]. For the top of atmosphere, our model produces -0.92, +0.31 and -0.60 W/m^2 for the shortwave, long wave and net radiative forcing. We compare these to some recent modeling studies in Table S1.

Table S1: Dust radiative forcing (W/m^2)

Radiative Forcing (W/m2)	This paper	Woodward [2001]	Miller et al. [2004]	Miller et al., in press
TOA				
SW	-0.92	-0.16	-0.33	-0.62
LW	0.31	0.23	0.15	0.22
Net	-0.6	0.07	-0.18	-0.4
Surface				
SW	-1.59	-1.22	-1.82	-1.29
LW	1.13	0.4	0.18	0.45
Net	-0.46	-0.82	-1.64	-0.84

As discussed in more detail in Yoshioka et al. [submitted—available www.cgd.ucar.edu/tss/staff/mahowald)], this model tends to have larger long wave forcing and smaller shortwave forcing at the top of atmosphere. For the cases of *Woodward* [2001] and *Miller et al.* [2004], the differences are largely due to our larger size distribution (more big particles) and to the smaller imaginary part of the index of refraction we use, following *Sinyuk et al.* [2003]. These result in a larger single scattering albedo than earlier studies (less absorption). We obtain more similar results to *Miller et al.* [in press], which has similar optical properties assumptions, and more similar size distributions. Comparisons to observations suggest that our radiative forcings are within the uncertainties of the observations [*Yoshoika et al.*, submitted available at www.cgd.ucar.edu/tss/staff/mahowald].

The only previous study on LGM dust impacts on climate using a realistic dust distribution is the *Claquin et al.* [2002] study, and that study only calculated radiative forcings, not feedbacks. That study included several different optical properties for the calculation of dust radiative forcing, and thus includes a range of values. In Table 2 we compare our dust optical depths and radiative forcings with the results from that study. In general, our results are within the range of the sensitivity studies conducted in that study, except in the tropics, where our dust burdens and radiative forcings are lower in both the modern and the last glacial maximum simulations. Our global average top of atmosphere forcings are lower than the range predicted in *Claquin et al.* [2002] due to the differences in the tropics. These differences are due to differences in the dust amount, due to differences in model parameterizations of dust entrainment and deposition, as well as differences in the physical models surface winds and precipitation. In addition, differences in the radiative forcing calculation can be from differences in the optical properties assumed (see *Yoshioka et al.,* [submitted: available at www.cgd.ucar.edu/tss/staff/mahowald] for more details on our assumptions).

Table S2: Comparison of LGM vs. current radiative forcings (top of atmosphere).

	Claquin et al. (2003)			This study				
	AOD		TOA RF (W/m2)	AOD			TOA RF (W/m2)	
	Current	LGM	(LGM-Current)	Current	LGMC	LGMT	LGMC	LGMT
45-90N	0	0.07	-1.2 to -0.3	0.02	0.09	0.15	-0.62	-0.80
15-45N	0.13	0.19	-1.9 to 0.8	0.14	0.22	0.21	-1.30	-1.88

0-15N	0.14	0.26	-3.8 to -2.3	0.10	0.16	0.14	-1.27	-1.38
0-15S	0.01	0.13	-3.1 to -1.0	0.01	0.03	0.03	-0.33	-0.36
15-45S	0.01	0.15	-3.1 to -1.0	0.02	0.02	0.04	-0.04	-0.59
45-90S	0	0.02	-0.4 to +0.2	0.00	0.00	0.01	-0.03	-0.04
Global	0.05	0.14	-2.3 to -1.0	0.05	0.09	0.10	-0.60	-0.91

Section S3: Efficacies

Hansen et al. [2005] use the nomenclature of efficacy to explore how effective different radiative forcings are in causing a change in temperature. Climate efficacy is defined as the global mean temperature change per unit forcing produced by the forcing agent relative to the response by carbon dioxide. For the T42 version of CAM3, the temperature response to a doubled carbon dioxide forcing (3.8 W/m^2) is 2.5 ◦C [*Kiehl et al.,* 2006]. For the current climate, our surface temperature response is -0.20, while the top of atmosphere instantaneous radiative forcing change is -0.60 W/m^2 (SOMF case). This results in an efficacy of 0.50. This is much lower than the instantaneous efficacy seen for mineral aerosols (0.87) in the GISS model [*Hansen et al., 2005*]. Under different climates, the efficacy varies between 0.49 to 1.19, with the highest effacies in the last glacial maximum, and thus appears to be sensitive to climate. The differences in efficacy between this model and *Hansen et al.* [2005] may be due to differences in the physical model, the aerosol distribution or aerosol optical properties assumed here. As discussed in detail in *Hansen et al.* [2005], the instantaneous forcing (which we calculated here) is not as good an indicator of climate response as other forcings, which were not calculated for this study.

Section S4: Estimates of anthropogenic radiative forcing and climate response

We simulate a 0.43 W/m^2 increase in radiative forcing in the current climate relative to the pre-industrial (Table 1) due to a decrease in dust from carbon dioxide fertilization of the vegetation, and changes in surface wind and soil moisture, ignoring a possible 50% dust source from land use change [*Mahowald and Luo,* 2003; *Tegen, et al.*, 2004; *Mahowald, et al.*, 2004; *Mahowald et al.,* 2005]. Estimates of preindustrial to current climate changes are a 60% increase in dust to a 24% decrease in dust in the only published study looking at the impacts of land use, climate change and carbon dioxide fertilization on vegetation [*Mahowald and Luo,* 2003]. Using the relationship between radiative forcing and temperature response to AOD we obtain in this study (–14.7 W/m^2/AOD and 12.7 K/AOD) and the changes in dust loading between current and pre-industrial climate estimated in *Mahowald and Luo* [2003] (-24% to +60%) based on our current climate AOD value (we use the more conservative value of 0.039 from the SOM case, instead of the larger value from the SOMB case), we estimate that the anthropogenic influence on dust (including climate and land use) is –0.15 to 0.33 W/m^2 at the top of atmosphere radiative forcing. We calculate here the anthropogenic forcing as the difference between preindustrial dust forcing and current climate forcing (-0.46 W/m^2). Similarly, we can estimate the impact of the –24% to +60% change in dust would have on surface temperatures, assuming the linear relationship we find in this study (–12.3 °C/AOD) as –0.13 to 0.29 °C; these values are the change from current climate dust loadings impact on surface temperature (-0.20°C). These results are significant relative to the ~1.5 W/m^2 radiative forcing impact of anthropogenic carbon dioxide and 0.4-0.8 °C change in surface temperature observed [*Houghton et al.,* 2001], and suggest a better understanding of changes in dust over the past 100 years is important.

Figure captions

Figure S1: Evolution of globally averaged surface temperature with time (first year is not plotted) for each of the cases discussed in the text. The dotted line shows the time period of the averages used for the analysis in this paper, as well as the averaged surface temperature. The color scheme is the same as that used in Figure 1.

Figure S2: Top of atmosphere radiative forcing (W/m2) for dust aerosols in each climate scenario. These plots represent the case of including dust minus not including dust within the same climate.

Figure S3: Surface radiative forcing (W/m2) for dust aerosols in each climate scenario, similar to Figure 1S.

Figure S4: Scatter plots showing the relationships between aerosol optical depth and radiative forcing (triangles are top-of-atmosphere forcings, while squares are for the surface values)(a), surface temperature response (b) and precipitation response (c) in the 5 climate simulations shown in Table 1. The color scheme is the same as in Figure 1.

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