Impact of vegetation and preferential source areas on global dust aerosol: Results from a model study

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[1] We present a model of the dust cycle that successfully predicts dust emissions as determined by land surface properties, monthly vegetation and snow cover, and 6-hourly surface wind speeds for the years 1982–1993. The model takes account of the role of dry lake beds as preferential source areas for dust emission. The occurrence of these preferential sources is determined by a water routing and storage model. The dust source scheme also explicitly takes into account the role of vegetation type as well as monthly vegetation cover. Dust transport is computed using assimilated winds for the years 1987–1990. Deposition of dust occurs through dry and wet deposition, where subcloud scavenging is calculated using assimilated precipitation fields. Comparison of simulated patterns of atmospheric dust loading with the Total Ozone Mapping Spectrometer satellite absorbing aerosol index shows that the model produces realistic results from daily to interannual timescales. The magnitude of dust deposition agrees well with sediment flux data from marine sites. Emission of submicron dust from preferential source areas are required for the computation of a realistic dust optical thickness. Sensitivity studies show that Asian dust source strengths are particularly sensitive to the seasonality of vegetation cover.

INDEX TERMS: 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 0322 Atmospheric Composition and Structure: Constituent sources and sinks; 0365 Atmospheric Composition and Structure: Troposphere—composition and chemistry; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; KEYWORDS: aerosols, mineral dust, tracer modeling, global modeling


1. Introduction

[2] The importance of atmospheric aerosol as a climate-forcing factor is well established, though there are still considerable uncertainties in the estimates of the direct and indirect effects of aerosols on the radiation regime [Intergovernmental Panel on Climate Change, 2001]. Soil-derived dust is one of the largest contributors to the global aerosol loading and can be expected to have a strong impact on regional and global climates. In addition, inputs of soil dust impact the productivity of marine [Martin, 1991; Coale et al., 1996; Hutchins and Brunland, 1998] and terrestrial [Swap et al., 1992; Chadwick et al., 1999] ecosystems, thus potentially influencing the global carbon cycle. Dust sedimentation rates, reconstructed from ice core or marine records, indicate that dust loads were considerably higher during cold periods in the past [Petit et al., 1990; Legrand, 1995; Steffensen, 1997; Petit-Maire, 1999] with greater potential to impact the climate [Claquin et al., 2002]. Dust emissions are likely to change in the future in response to anthropogenic changes in climate and consequent changes in natural vegetation pattern [Harrison et al., 2001] and also as a consequence of changes in land use.

[3] Global simulations of the modern dust cycle [e.g., Gentoth, 1992; Tegen et al., 1996; Andersen et al., 1998; Mahowald et al., 1999; Reader et al., 1999] are capable of reproducing the first-order patterns of dust transport and deposition under modern climate conditions. None of the current models is capable of correctly reproducing the magnitude of dust emissions from Sahara and Asia at the same time, and all of the global dust models overestimate dust emission and transport from Australia. These shortcomings probably reflect oversimplifications in the treatment of dust emissions in the current generation of models.

[4] The emission of dust from the land surface is strongly controlled by the presence and density of the vegetation
cover [Wolfe and Nickling, 1996; Marticorena et al., 1997; Wyatt and Nickling, 1997]. In semiarid areas, for example, the rapid growth of annual grasses and herbs after rain is sufficient to suppress dust emissions over a matter of days to weeks, while the presence of shrubby vegetation substantially reduces emissions even during the period when the plants are not in leaf [Wyatt and Nickling, 1997]. Although models of the dust cycle allow vegetation type to implicitly or explicitly influence dust sources, the dust source schemes currently employed do not take into account seasonal variations in vegetation cover and their impact on the extent of dust source areas. [5] Even in sparsely vegetated areas, not all surfaces are active dust sources. Dust emissions tend to be concentrated in areas where the recent geomorphological history has resulted in concentration of fine-grained material and the creation of large areas with low surface roughness. Dry lake beds, the relics of formerly more extensive lakes in the past, are a good example of these so-called preferential source areas [Pye, 1987; Prospero et al., 2002] because lacustrine sediments are characteristically fine-grained. A classic example of a paleolake preferential dust source area is the Bodele depression, to the north of Lake Chad, which is an area of little or no surface relief consisting of fine-grained lacustrine sediments deposited by paleolake Chad in the early to mid-Holocene [McTainsh, 1987; Livingstone and Warren, 1996]. Paleolake beds in, for example, central Australia have also been shown to be significant sources of dust today [Middleton, 1984]. All topographic depressions that contain fine-grained materials could act as preferential dust sources, including, for example, glacial outwash plains, riverine floodplains, alluvial fans, and washes. There are some situations where dust deflation from these sources appears to have been significant. However, at a global scale, paleolakes are areally the largest of such dust sources. Models of the dust cycle that define source areas using modern observational data implicitly include the occurrence of preferential sources, but only Ginoux et al. [2001] have explicitly defined preferential source areas in a dust emission scheme. They consider all large-scale topographic depressions as preferential source areas. [6] In this paper, we present a new dust emissions scheme which incorporates an explicit treatment of both preferential sources and the role of seasonal changes in vegetation cover. We show that this model is capable of reproducing the regional patterns in seasonal and annual dust loadings and deposition under modern climate conditions and yields plausible estimates of interannual changes in dust emissions.

2. Model Description

2.1. Dust Source Model

[7] Global time-dependent dust emissions are calculated explicitly, taking into account vegetation type and cover, preferential dust source areas, particle size distribution, snow cover, soil moisture, and surface wind speed. The dust source model has a 6-hour time step and 0.5° horizontal resolution.

2.1.1. Simulation of Vegetation Type and Cover

[8] We use an equilibrium terrestrial biogeography model (BIOME4, [Kaplan, 2001]) to determine the distribution of potential vegetation types (biomes). BIOME4 predicts the distribution of 27 biomes as a function of monthly mean temperature, precipitation, net radiation, and soil type on a 0.5° grid. The simulated distribution of the biomes agrees well with observations [Kaplan, 2001]. Nonforest biomes (barren land, desert, tropical xerophytic shrubland, temperate xerophytic shrubland, tropical grassland, temperate grassland, graminoid and forb tundra, erect dwarf shrub tundra, prostrate dwarf shrub tundra, and cushion forb tundra) are considered as potential dust sources; that is, dust emissions can occur from these biomes when other criteria (including sparse vegetation cover, soil dryness, and absence of snow cover) are satisfied.

[9] We calculate seasonal and interannual changes in vegetation cover from observations, as the fraction of absorbed photosynthetically active radiation (FPAR) derived from monthly retrievals of the normalized difference vegetation index (NDVI) from the advanced very high resolution radiometer (AVHRR) satellite instrument [Braswell et al., 1997] (available for the years 1982 to 1993 at 0.5° horizontal resolution) using the empirical relationship FPAR = 1.222 · (NDVI/0.559–0.1566) [Knorr and Heimann, 1995]. Negative values are set to zero.

[10] In the grass-dominated biomes (i.e., desert, temperate grasslands, tropical grasslands, and graminoid and forb tundra) we assume that the aboveground biomass varies over the seasonal cycle and that dust deflation can occur whenever there is no green vegetation. For each model gridcell characterized by a grass-dominated biome the dust source area is assumed to be zero in any month when FPAR >0.25 and is assumed to increase linearly with decreasing vegetation cover to a maximum at FPAR = 0. This formulation accounts for the prevention of dust emissions by even a modest grass cover.

[11] Shrubs can protect the soil surface from deflation even when there are no leaves present (i.e., FPAR = 0). Thus it is not possible to use this simple FPAR formulation to determine the extent of the dust source area in shrub-dominated biomes. However, the degree to which the presence of dormant shrubs protects the surface varies with the density and spacing of the shrubs and the nature of the ground cover between the shrubs. In those tundra biomes (erect dwarf shrub tundra, prostrate dwarf shrub tundra, and cushion forb tundra), in which shrubs are either strongly dominant or generally associated with long-lived perennial forbs, we use the maximum observed FPAR as an index of the density of the shrub cover. We assume that the effective vegetation cover at each gridcell is the same throughout the year and corresponds to the annual maximum of FPAR. Thus shrub-dominated biomes with annual maximum FPAR >0.25 do not act as dust sources, but shrub-dominated biomes with annual maximum FPAR <0.25 have a dust source area that is the same throughout the year but increases linearly to a maximum at annual maximum FPAR = 0.

[12] We have adopted a different procedure for temperate and tropical shrublands, which are characterized by a mixture of shrubs and grasses. In general, the relative importance of grasses increases as the overall productivity of the vegetation decreases. We have used mean annual FPAR as an index of the productivity of these shrublands. In cases where the mean annual FPAR is >0.5 we assume that the temperate/tropical shrubland at that location is both...
dense and highly productive and is therefore characterized by the absolute dominance of shrubs and the relative unimportance of grasses. These gridcells are treated as shrub-dominated vegetation. The effective vegetation cover at each gridcell throughout the year is therefore determined by the annual maximum FPAR. Thus highly productive temperate and tropical shrublands do not act as potential dust sources. We assume that less productive temperate and tropical shrublands (mean annual FPAR <0.5) are characterized by widely spaced shrubs and a significant presence of grasses. The widely spaced shrubs do not act as a significant impediment to dust deflation, and the grasses will only impede deflation when they are present. Thus we use the same relationship between FPAR and dust source area in less-productive temperate/tropical shrublands as in pure grassland biomes: The dust source area is 0 when the monthly FPAR is >0.25 and increases linearly to a maximum at FPAR = 0.

The effective surface $A_{\text{eff}}$ for dust emission is thus calculated as

$$A_{\text{eff}} = 1 - \left(\text{FPAR}_{\text{max}} \cdot f_{\text{shrub}} + \text{FPAR}_{\text{month}} \cdot f_{\text{grass}}\right) \cdot \frac{1}{0.25}.$$  

(1)

where $f_{\text{shrub}}$ and $f_{\text{grass}}$ are the relative contribution for coverage with shrub or grass vegetation for each biome type (here $f_{\text{shrub}}$ is 1 for shrub-dominated vegetation and $f_{\text{grass}}$ is 1 for grass-dominated biomes; for other biome types $f_{\text{shrub}}$ and $f_{\text{grass}}$ are 0). The factor $1/0.25$ is the slope in the linear relationship between dust emission and FPAR, when assuming a limit of FPAR = 0.25 for dust emissions and assuming a linear relationship between the fraction of unvegetated area per gridcell and dust emission.

### 2.1.2. Simulation of Preferential Dust Source Regions

It would be possible to specify the extent of paleolake deposits for certain regions on the basis of detailed geological and geomorphological mapping (see, e.g., maps of paleolake extent in the Great Basin, western United States given by Street-Perrott and Harrison [1985]). Similar maps have been produced for northern Africa [e.g., Petit-Maire, 1991; Hoelzmann et al., 1998]. Unfortunately, such maps are not available for other regions, and therefore we have adopted an alternative strategy involving explicit simulations of the extent of paleolake beds across the globe. We assume that climate variations during the Quaternary have been sufficiently large to have allowed lakes to form in closed basins at some time in the past, although the timing of lake formation would be different for different regions. We specify the extent of these lakes using a high-resolution water routing and storage model, HYDRA [Coe, 1998]. HYDRA uses land surface topography at 5° resolution to determine the location and extent of lakes and wetlands as a function of runoff, precipitation, and surface evaporation.

### 2.1.3. Soil Texture and Particle Size Distribution

The particle size distribution for individual soils is described by four populations: clay, silt, medium/fine sand, and coarse sand (Table 1). There is no global data set that would give the relative proportions of each of these populations in specific soils. We therefore derived global estimates from the soil texture class data given in the Food and Agriculture Organization/United Nations Educational, Scientific, and Cultural Organization soil map of the World [Zobler, 1986]. This classification describes the soil texture of the top 30 cm of the dominant soil in each 0.5° grid cell. The texture categories are fine, medium, coarse, or mixtures of these (e.g., medium-fine results when in 50% of the grid cell area the dominant soil is fine and in 50% of the area the dominant soil is medium-textured). In terms of the standard soil textural triangle [see e.g., Fitzpatrick, 1980] we assume that the coarse texture category includes sands, loamy sands, and sandy loams; the medium texture category includes sandy loams, loams, sandy clay loams, silt loams, silt, silty clay loams, and clay loams with <35% clay; and the fine texture category includes clays, silty clays, sandy clays, clay loams, and silty clay loams with >35% clay. The percentage of sand, silt, and clay particles in each texture category is estimated from the centroids of the appropriate texture classes in the textural triangle. In the case of mixed soil categories (e.g., medium-fine) we assume that the soil type lies close to the boundaries between categories. Thus soils classified as medium-fine probably vary in texture between silt loams, silty clay loam, and clay loams, and we used the centroid of this more limited set of texture classes to determine the relative abundance of sand, silt, and clay (Table 2). We use the textural description to determine

### Table 1. Properties of the Four Particle Size Populations

<table>
<thead>
<tr>
<th>Type</th>
<th>Diameter range, µm</th>
<th>Mean diameter, µm</th>
<th>σ</th>
<th>Density, g cm⁻³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse sand</td>
<td>5000–1000</td>
<td>710</td>
<td>2</td>
<td>2.65</td>
</tr>
<tr>
<td>Medium/fine sand</td>
<td>50–500</td>
<td>160</td>
<td>2</td>
<td>2.65</td>
</tr>
<tr>
<td>Silt</td>
<td>2–50</td>
<td>15</td>
<td>2</td>
<td>2.65</td>
</tr>
<tr>
<td>Clay</td>
<td>0–2</td>
<td>2</td>
<td>2</td>
<td>2.5</td>
</tr>
</tbody>
</table>

We determined the potential maximum areal extent of lakes assuming that the supply of precipitation was unlimited. The difference between the simulated maximum areas of lakes and the actual present-day areas of lakes is assumed to indicate the extent of paleolake deposits formed under wetter climate conditions at some time in the recent geological past. These exposed paleolake areas are then used as preferential source areas for dust emission, provided they meet the additional criterion for dust deflation imposed by vegetation, soil moisture conditions, and wind strength. Although HYDRA explicitly simulates rivers, the 5° topography data set is at too coarse a resolution to make it possible to determine the width of the river channels. Thus we cannot explicitly determine the extent of preferential sources in fluvial environments. These fluvial source areas are, however, less important than paleolake basins over most regions of the world, and thus this omission is unlikely to significantly impact on our results.

### Table 2. Relative Proportions (Percent) of Coarse Sand, Medium/Fine Sand, Silt, and Clay Attributed to Each Soil Texture Category

<table>
<thead>
<tr>
<th>Texture Class</th>
<th>Coarse Sand</th>
<th>Medium/Fine Sand</th>
<th>Silt</th>
<th>Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse</td>
<td>43</td>
<td>40</td>
<td>17</td>
<td>0</td>
</tr>
<tr>
<td>Medium</td>
<td>0</td>
<td>37</td>
<td>33</td>
<td>30</td>
</tr>
<tr>
<td>Coarse-medium</td>
<td>10</td>
<td>50</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Coarse-medium/fine</td>
<td>23</td>
<td>23</td>
<td>19</td>
<td>35</td>
</tr>
<tr>
<td>Fine</td>
<td>0</td>
<td>0</td>
<td>34</td>
<td>67</td>
</tr>
<tr>
<td>Coarse-fine</td>
<td>0</td>
<td>50</td>
<td>12</td>
<td>38</td>
</tr>
<tr>
<td>Medium-fine</td>
<td>0</td>
<td>27</td>
<td>25</td>
<td>48</td>
</tr>
<tr>
<td>Lacustrine sources</td>
<td>0</td>
<td>0</td>
<td>100</td>
<td>0</td>
</tr>
</tbody>
</table>
whether the sand present was coarse or medium/fine and also to decide whether clay was present as clay-sized particles or as silt-sized aggregates. Clay loams are highly unlikely to contain coarse sand, for example, while sandy clay loams could contain both coarse and medium/fine sand. Thus any sand present in those categories which are characterized by clay loams is considered to be medium/fine sand, whereas sand present in those categories characterized by sandy clay loams would be allocated equally to the coarse and to the medium/fine sand populations. Similarly, although a small percentage of clay is allowed in the coarse category, loamy sands tend to contain aggregated clay particles. Thus the percentage of clay in categories characterized by loamy sands was added to the silt fraction.

[16] We have created a new texture class for preferential sources (Table 2). Deep-water lacustrine deposits are typically formed from clay-sized material [Street-Perrott and Harrison, 1985]. However, in drying or seasonally dry lakes in arid to semiarid regions a variety of physical and biological processes operate to create aggregates of the clay-sized material. The aggregates can vary in size but are typically in the range of fine to medium silt (Table 3) and are sufficiently robust to remain intact during deflation and aeolian transport. These aggregates eventually break down to their component clay-sized particles, particularly under more humid conditions, although the environmental conditions under which this takes place are not fully documented. Clearly, the creation of silt-sized aggregates is likely to enhance aeolian deflation of drying or dry lake sediments. We have therefore assumed that the surface material in paleolake basins identified as preferential source areas consists of silt-sized aggregates and is deflated as such. The median radius of the aggregates derived from lacustrine sediments has been measured as 15 μm for dry lake beds in western North America and 27 μm for an Australian location (Table 3). These values were used in the global calculations of dust deflation from preferential sources: We assumed that the median particle radius for silt aggregates in preferential source areas was 15 μm in the northern hemisphere (NH) and 27 μm in the southern hemisphere (SH). The threshold wind stress required to lift particles of these sizes are 30 and 20 cm s⁻¹, respectively; so the assumption that all paleolake basins are characterized by silt-sized aggregates makes these regions highly effective dust sources.

[17] The ratio α of vertical dust flux to horizontal soil particle flux reflects the availability of fine dust, which is partly a function of particle size distribution and partly a function of surface properties (interparticle cohesion and crusting). Following the approach of Marticorena et al. [1997] (who summarize the results from measurements from seven soil samples by Gillette [1978]) we assume α ratios of 10⁻³ cm⁻¹, 10⁻⁶ cm⁻¹, and 10⁻⁷ cm⁻¹ for silt, fine/medium sand, and coarse sand, respectively. We take into account that the high cohesive forces of clay particles inhibit soil deflation in soils with high clay content by assuming emission factors for clay populations of 10⁻⁶ cm⁻¹ for soils with clay content of <45% and 10⁻⁷ cm⁻¹ for those with >45% clay. We use an α ratio of 10⁻⁵ cm⁻¹ for our “lacustrine source” texture class, which is the highest used for any soil type (see Table 2). The use of a high α ratio, together with the low threshold velocities required to lift dust particles from such surfaces, causes high dust emissions from the preferential source regions.

[18] The kinetic energy released during saltation can cause emission of dust particles at wind speeds below their inherent threshold velocity. We assume that particles smaller than those that would be dislocated at a given wind speed are removed during dust events in proportion to their abundance in the parent soil. According to a wind tunnel study, saltating sand grains cause the dislocation of individual clay particles at wind speeds >10.5 m s⁻¹ [Alfaro et al., 1997]. We therefore assume that silt-sized aggregates from preferential sources are separated into clay-sized particles when the surface wind speed exceeds 10 m s⁻¹.

2.1.4. Snow Cover and Soil Moisture

[19] We exclude snow covered areas as dust sources. The daily snow cover for the individual years was taken from the European Centre for Medium Range Weather Forecast (ECMWF) reanalysis (ERA) data on T106 resolution, interpolated to 0.5°. The snow covered area A_{snow} is calculated from the snow depth C (in meters) as

\[ A_{snow} = \min[1, \left( \frac{C}{0.015 \text{ m}} \right)] \]

[DKRZ Model User Support Group, 1992]. Comparisons with passive microwave snow data derived from the Nimbus-7 Scanning Multichannel Microwave Radiometer indicate that this approximation produces a reasonable estimate of snow area [Foster et al., 1996].

[20] Dust emission is suppressed when the soil is wet. However, the strong winds required for dust emission increase evaporation and typically cause rapid drying of the soil surface. Thus deflation can take place, even shortly after a precipitation event [Gillette, 1999], when the upper-
most part of the soil (skin) is dry, even though the remainder of the soil may be wet. The BIOME4 model calculates soil moisture using a two-layer soil scheme [Haxeltine and Prentice, 1996]. We have assumed that the surface layer could be dry (permitting dust emission) unless the uppermost soil layer is at field capacity, in which case dust emission is suppressed.

### 2.1.5. Wind Speed

Soil deflation from bare surfaces occurs when the surface winds exceed a certain threshold velocity, which depends on soil particle size distribution as well as on surface conditions like vegetation cover and soil moisture. The optimum particle diameter for mobilization lies in the range of 60–100 μm [Bagnold, 1941; Iversen and White, 1982]. The threshold velocities for each size fraction were calculated following Marticorena and Bergametti [1995]. The wind-blown particles can mobilize smaller particles through saltation [Shao et al., 1993].

The horizontal particle flux $G$ and the vertical dust particle flux $F$ are predicted as follows [White, 1979]:

$$F = \alpha \cdot A_{\text{eff}} \cdot G \cdot (1 - A_{\text{snow}}) \cdot I_0$$

where $\rho_0$ is the air density, $g$ is the gravitational constant, $u_*$ is the surface wind stress, $u_{tr}$ is the threshold wind stress, $I_0$ is 0 when upper-layer soil moisture is at field capacity and 1 otherwise, and $s_i$ is the relative surface area covered by each size fraction $i$. We used ERA 10-m surface wind products with $1.25^\circ \times 1.25^\circ$ horizontal resolution and 6-hourly time step to compute dust emissions.

### 2.2. Dust Transport Model

Dust transport and deposition are simulated with the tracer transport model TM3 [Heimann, 1995]. TM3 has $3.75^\circ \times 5^\circ$ horizontal resolution and 19 vertical levels. Advection of the tracers is computed using the slopes scheme [Russell and Lerner, 1981], and mixing is computed for dry and moist convection. We ran the model with 6-hourly ECMWF ERA wind fields for the years 1987–1990. The 6-hourly global dust fluxes from each 0.5 degree gridcell were summed up to the $3.75^\circ \times 5^\circ$ resolution required for input into TM3. Seven dust size classes are transported as independent tracers. The limits of the size classes were set at 0.1 μm, 0.3 μm, 0.9 μm, 2.6 μm, 8. μm, 24 μm, 72 μm, and 220 μm.

The parameterization of dust deposition largely follows Tegen and Fung [1994]. Dry deposition of dust particles is parameterized as gravitational settling (Stokes-Cunningham) and turbulent mixing to the surface from the first atmospheric layer. The gravitational settling velocity is size dependent and calculated by

$$v_n = \frac{2r^2g\rho}{9\eta} C_{\text{Cunn}}$$

where $r$ is the particle radius, $g$ is the gravitational constant, $\rho$ is the particle density, and $\eta$ is the dynamic viscosity of air with

$$C_{\text{Cunn}} = 1 + \frac{\lambda}{r} \left(A + B \cdot \exp\left(-\frac{C\theta}{\lambda}\right)\right)$$

where $A = 1.249$, $B = 0.418$, $C = 0.847$, and $\lambda$ is the mean free path of the air molecules [Kasten, 1968]. The Cunningham correction factor $C_{\text{Cunn}}$ is important at higher layers in the atmosphere, where the mean free path of the air molecules is large. For large particles that cross several atmospheric layers during one model time step the dust is redistributed into the layer the dust reaches at the end of the time step. To calculate dust sedimentation velocities, the radius that is representative for the mass deposition per size bin is calculated according to Tegen and Fung [1994].

Wet deposition of aerosol particles occurs by in-cloud and subcloud removal. Dust is assumed to be nonhygroscopic and therefore does not form cloud condensation nuclei. Scavenging by subcloud removal is parameterized using a scavenging ratio of 750, as given by Tegen and Fung [1994] for large-scale precipitation. Washout can be effective in wet deposition during deep convection events, since cloud formation can occur through the ice phase and form ice nuclei. Here we assume complete washout of dust particles in wet convection events above the first atmospheric layer. This reduces the mixing of dust particles into the upper troposphere in the tropics.

### 2.3. Strategy for Validation

We have evaluated three aspects of our simulations of the modern dust cycle: the large-scale patterns of atmospheric dust loading using satellite-derived data, the magnitude and temporal patterns of dust concentrations at key sites using continuous dust monitoring data, and the magnitude and spatial patterns in dust deposition using marine sediment trap data.

We have used the absorbing aerosol index derived from the Total Ozone Mapping Spectrometer satellite instrument (TOMS AI, see Herman et al. [1997]) in order to evaluate the large-scale patterns of atmospheric dust loading and to see whether the model is capable of reproducing the transport patterns associated with individual dust events. We used daily values of the TOMS AI for the 4 years 1987–1990. The TOMS instrument detects the presence of dust and other absorbing aerosols (primarily black carbon, BC) based on the observed departure of backward-scattered UV radiation from pure Rayleigh scattering. The magnitude of the TOMS AI depends on the optical thickness of the absorbing aerosols, the height of the aerosol layer, the optical properties of the aerosols, the viewing angle of the instrument, and the presence of clouds. In order to compare these data with the dust optical thickness simulated by our model, we have modified the TOMS AI data set by removing those gridcells where the BC aerosols are likely to be the most important contribution to the observed optical thickness or where cloud cover could have affected the observed optical thickness significantly. We used the simulation of Lioussse et al. [1996] as a guide to the importance of BC aerosols from industrial and biomass burning sources under modern climate and removed those cells for those
months where the average simulated BC optical thickness was >0.01. We also screened the TOMS AI data set for cells affected by cloud cover and removed those cells where the reflectivity at 380 nm was >12%. The approach is similar to that employed by Hsu et al. [1999], but they used a slightly lower reflectivity (9%), which excludes cells in which the impact of clouds on the registration of absorbing aerosol is likely very slight. If the lower value of 9% is used to exclude cloudy scenes, for example, no dust signal is retrieved for the extremely active dust source of the Bodele depression in the southern Sahara. Since the TOMS instrument cannot detect absorbing aerosols in the lowermost 1–1.5 km of the atmosphere, we confine our comparisons to simulated dust in the atmospheric layers above 1.5 km. Given that a number of assumptions about the optical properties of aerosols have to be made a priori in the retrieval algorithm of aerosol products from satellite instruments [King et al., 1999], there are large uncertainties attached to quantitative reconstructions of the optical thickness. We therefore make no attempt to make quantitative comparisons between the simulated dust loading and the TOMS AI.

[29] In addition to checking spatial patterns and temporal variations of atmospheric dust qualitatively with the TOMS AI retrievals, the simulated dust can be quantitatively checked against measurements of atmospheric dust concentration and deposition. We have used measurements of dust concentrations at 21 observation sites operated by the University of Miami (J. Prospero and D. Savoie, personal communication, 2001) to compare with the simulated concentration of dust in the first atmospheric layer of the model at the same locations. The 21 sites included in this study represent both near- and far-field locations for dust transport from each of the major dust sources regions. At each site, high-volume bulk aerosol filter samples have been collected under selected meteorological conditions (to minimize the possible impact of local sources) on a daily basis. Dust concentrations were computed either from aluminum concentrations based on a crustal abundance of 8% Al or from the weights of filter samples ashed at 500°C after extracting with water. The length of the records varies from 2 to 15 years. The data were available as monthly averages over the sampling period and are compared with the simulated lower-atmosphere dust concentration.

[29] Modern rates of dust sedimentation to the ocean can be used to supplement the quantitative records of atmospheric dust concentration. Existing data sets [see e.g., Ginoux et al., 2001] contain a mixture of actual dust flux measurements and estimates of the dust flux to the ocean derived from measurements of atmospheric dust concentrations using a simple deposition model. We have therefore compiled a new data set of dust deposition rates to the ocean using marine sediment trap data archived in the Dust Indicators and Records in Terrestrial and Marine Paleoenvironments (DIRTMAP) [Kohfeld and Harrison, 2001] database.

[29] Marine sediment traps are generally deployed for several seasons, although some of the available records are based on a single dust season or even a single dust event. We have screened the sites and removed records from sediment traps which were deployed for <50 days and thus represent a short term signal. We have also removed records from sites where the measurements could have been affected by fluvial inputs or hemipelagic reworking. As a result of these screening procedures, the data set used for comparison purposes consists of 47 sites (Table 4). Terrestrial accumulation rates were calculated by isolating terrigenous material from organic carbon, carbonate content, and biogenic opal (e.g., Wefer and Fischer, [1993], except for sites from the north Pacific derived from Saito et al., [1992]). The terrigenous content at the north Pacific sites was estimated from Al concentration measurements assuming that terrigenous material is 8% Al. The data set has limited coverage in the northern Pacific Ocean and the SH but provides a picture of the gradients in the magnitude of dust deposition to the Atlantic Ocean and Arabian Sea that can be quantitatively compared with the simulated dust deposition fluxes.

[31] Additional estimates of dust deposition rates to the ocean could be made from the dust measurements available from the uppermost part of marine sediment cores. Estimates of recent dust deposition based on marine core top records from the DIRTMAP database have been used in previous evaluations of the modern dust cycle [e.g., Mahowald et al., 1999]. Deposition fluxes to the ocean are rather low, and most of these core top records represent the integrated dust flux to the ocean over a time span ranging from at least hundreds to possibly a few thousands of years [see Kohfeld and Harrison, 2001]. The use of such a long-term integration to represent the "modern" dust flux is fully justified when the modern simulation of the dust cycle is run for comparison with a simulation of the dust cycle at some key time in the geologic past (e.g., the last glacial maximum [Mahowald et al., 1999]). However, it is not appropriate to use these core top data to evaluate simulations designed to reproduce the dust cycle within specific years.

[32] Measurements of dust deposition on land are available from a number of regions, including the southwest United States [Reheis and Kihl, 1995], northern Africa [Ramsperger et al., 1998], China [Derbyshire et al., 1998], and Australia [McTainsh and Lynch, 1996]. However, dust flux measurements made using suspended dust traps can be inflated by the presence of organic components, plant residues, or industrial pollutants [Reheis and Kihl, 1995; Ramsperger et al., 1998; Derbyshire et al., 1998]. Some of the available measurements are from sites close to roads or settlements, and it is likely that a significant proportion of the dust is due to anthropogenic disturbances (e.g., from construction, off-road traffic, or agriculture) that are not taken into account in our simulation. The evaluation of the continental records is complex, and we have therefore not used these records for comparison with simulated dust deposition rates over the continents.

3. Results
3.1. Results of the Dust Source Model

[31] The simulated distribution of preferential source areas (Figure 1) is in good agreement with the known extent of Late Quaternary paleolakes [Street-Perrott et al., 1989] for those areas for which geomorphic information exists. Paleoenvironmental evidence supports the existence of extensive paleolakes in now arid regions of northern Africa during both the last interglacial and the early to
<table>
<thead>
<tr>
<th>Sample ID</th>
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<th>Latitude</th>
<th>Lith. Fluxes, gm^-2 yr^-1</th>
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<td>[Jickells et al., 1996], [Wefer and Fischer, 1993]</td>
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mid-Holocene, with a significant expansion of Lake Chad into the Bodele Depression, extensive areas of paleolakes in the Niger Bend region, and large (interglacial) lakes in Morocco/Tunisia [Petit-Maire, 1991]. All of these regions are captured in our simulation of preferential sources. The preferential source to the north of the Caspian Sea and within the Great Basin of the southwestern United States were occupied by large paleolakes during the last glacial maximum [Street-Perrott and Harrison, 1985; Thompson et al., 1993]. The expanded source areas within the Lake Eyre Basin and in the Riverine Plains of eastern Australia correspond to paleolakes dating largely from the last glacial [Harrison and Dodson, 1983; Magee et al., 1995]. Many of these simulated sources are known to be major sources of dust under modern conditions, for example, the Bodele depression north of Lake Chad and the Willandra Lakes region in southeastern Australia [Middleton, 1984]. 

The average global dust emission simulated by the model is $805 \pm 82$ Mt yr$^{-1}$, with $475 \pm 51$ Mt yr$^{-1}$ for particles with $<10 \mu$m radius. Of this total dust emission, 12% consists of small ($<1 \mu$m radius) particles, which can be transported long distances and are responsible for most of the dust radiative forcing. Preferential source areas represent only 17% of the total area from which dust emissions are simulated at some time during the year. However, these source areas contribute 40% of the total mean annual dust emissions (particles with $<10 \mu$m radius) and 84% of the radiatively important submicron dust (Figure 2a). Thus failure to include preferential sources would lead to a substantial underestimation of the emissions (Figure 2b) and of the dust optical thickness.

[34] Allowing vegetation phenology to influence the seasonal evolution of dust emission produces a significant improvement in the simulation of dust emissions, particularly from central Asia (Figure 3). Dust emissions from

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Region</th>
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<th>Latitude</th>
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<td>46.12</td>
<td>175.03</td>
<td>4.09</td>
<td>[Kawahata et al., 1998a]</td>
<td></td>
</tr>
<tr>
<td>NP-B</td>
<td>W Pacific</td>
<td>46.82</td>
<td>162.12</td>
<td>0.86</td>
<td>[Saito et al., 1992]</td>
<td>estimated from Al concentration, assuming 8% Al terrigenous</td>
</tr>
<tr>
<td>48N21W-1</td>
<td>Mid Lat Atl</td>
<td>47.72</td>
<td>-20.87</td>
<td>3.10</td>
<td>[Honjo and Manganini, 1993]</td>
<td></td>
</tr>
<tr>
<td>48N21W-2</td>
<td>Mid Lat Atlantic</td>
<td>47.83</td>
<td>-19.50</td>
<td>2.80</td>
<td>[Jickells et al., 1996]</td>
<td></td>
</tr>
<tr>
<td>P</td>
<td>Station Papa</td>
<td>50.00</td>
<td>-144.98</td>
<td>0.30</td>
<td>[Fischer et al., 1988]</td>
<td></td>
</tr>
<tr>
<td>* 75N7W</td>
<td>Greenland Sea</td>
<td>74.5</td>
<td>-6.72</td>
<td>3.10</td>
<td>[Fischer et al., 1988]</td>
<td>possible ice rafted detritus</td>
</tr>
<tr>
<td>* 79N1E</td>
<td>Fram Strait</td>
<td>78.87</td>
<td>1.37</td>
<td>4.00</td>
<td>[Fischer et al., 1988]</td>
<td>possible ice rafted detritus</td>
</tr>
</tbody>
</table>

The asterisk indicates site not used. Latitude and longitude are given in decimal degrees, where + represents N or E, and − represents E or S, respectively.
Asia are largest during the spring months, giving rise to the so-called “Kosa” dust events [e.g., Iwasaka et al., 1983; Parungo et al., 1995]. These emissions result from strong winds occurring before or during the early part of the growing season, before vegetation growth is sufficiently established to suppress dust emissions. When dust sources are determined using annual average FPAR (Figure 3a), Asian dust sources are suppressed. Dust emissions from the Sahara are reduced by 4% as a result of taking vegetation phenology into account. This reduction is minor, and the magnitude of the simulated emissions from the Sahara is plausible. Furthermore, the timing of dust events, which is likely to be sensitive to the vegetation changes, is captured well.

[36] We performed two experiments to examine the sensitivity of dust emissions to the treatment of shrublands and, in particular, the arbitrary choice of FPAR = 0.5 to differentiate dense and sparse shrubland. In the first experiment all potential source areas were treated as though the vegetation consisted of dense shrubs (i.e., the area of potential sources was determined by the maximum FPAR during the year). In the second experiment all potential source areas were treated as though the vegetation was completely removed as FPAR decreased (i.e., as though the vegetation acted like grass). These two experiments represent end-members in the treatment of the control of vegetation on source areas, and the differences in the magnitude and seasonal representation of dust emissions between the two experiments provides a measure of the realism (or otherwise) of our actual parameterization of shrubland behavior. Figures 3b and 3c show the differences in dust fluxes for these cases compared with the reference case (see Figure 2a). In the first experiment (Figure 3b), in which all sources are treated as grass-covered, dust emissions increase by 10%. The largest changes occur in the southern Sahara/Sahel, around the Caspian and Aral Seas, and in the Canadian Arctic. The increase in emissions in the Sahara/Sahel is plausible, but the increases in Asia and the Canadian Arctic are unrealistic. In the second experiment (Figure 3c), in which all sources are treated as dense shrublands, dust emissions are reduced by 27% and emissions from Asia and Australia are completely suppressed. This second experiment produces a completely unrealistic simulation of the modern dust cycle. These experiments demonstrate the importance of vegetation phenology in controlling dust emissions and suggest that we have adopted a reasonable approximation in the treatment of shrub-dominated vegetation.

3.2. Size Distribution of Transported Dust

[37] During atmospheric transport there are major changes in the size distribution of dust. For example, the atmospheric size distribution in the first atmospheric layer
close to the source region in the western Sahara reflects the size distribution of emitted particles, with a maximum at 5 μm radius and secondary maxima around 40 μm (where the threshold for dust particle emission is at minimum) and 1 μm (Figure 4). The paucity of dust particles with radii >100 μm reflects the absence of wind speed events that are higher than the threshold velocity required to dislocate large particles. The lower end of the dust size distribution reflects the availability of fine dust particles in the parent soil. At more remote locations, 1000–3000 km downwind of the Sahara over the North Atlantic, the size distribution peaks at 1 μm radius, in good agreement with observations of remote dust [Schütz et al., 1981]. The absence of particles >10 μm radius is a
Figure 3. (a) Difference between annual dust emissions for the case of seasonally invariant vegetation cover, (b) vegetation cover consisting only of grasses, and (c) vegetation cover consisting only of shrubs, and the reference case for the year 1988. See color version of this figure at back of this issue.
over the North Atlantic.

and 3000 km (dashed line) downwind of the source region
source region (Sahara) (solid line) and 1000 (dotted line)
imixing ratio, g dust/kg air) in the first model layer for a

Figure 4. Modeled dust mass size distributions (X: mass
mixing ratio, g dust/kg air) in the first model layer for a
source region (Sahara) (solid line) and 1000 (dotted line)
and 3000 km (dashed line) downwind of the source region
over the North Atlantic.

3.3. Simulated Atmospheric Distribution:

Comparison With TOMS

Comparison of TOMS AI (Figure 5, top) and TM3
model results (Figure 5, bottom) on a daily basis for a
Saharan dust episode (30 March through 4 April 1988,
Figure 5a), for an Asian dust episode (25 May through
29 May 1989, Figure 5b) and an Australian dust episode
(28 November through 2 December 1987, Figure 5c) shows
that the spatial patterns of dust transport for specific events
are well reproduced. The daily model dust aerosol optical
thickness (AOT) above 1.5 km is compared with the
averaged daily TOMS AI for the period from 1987 to
1990 for several regions where dust aerosol dominates the
total absorbing aerosol AOT (central United States, Caspian
Sea region, northwest China and Mongolia, Australia,
eastern North Atlantic, Arabian Sea, Sahel and Sahara)
(Figure 6). The regionally averaged seasonal cycles of
modeled dust and retrieved AI are in relatively good
agreement, except for Australia. Though the timing of
the simulated individual events is consistent with the
observations, the relative magnitudes of the individual
peaks are not captured. This mismatch in the magnitude
of the individual peaks may be caused by variability in the
dust vertical distribution.

For the years 1987–1990 the model (Figure 7a)
reproduces observed interannual changes in the global
dust load (Figure 7b) for many regions. Year-to-year
changes in biomass burning in addition to variations in
dust loading would be visible in the difference maps of
the TOMS AI (Figure 7b). To minimize this contribution,
we only show the results for the NH spring, when
biomass burning is relatively low. The patterns of dust
loading over the southern Sahara, Mediterranean, and the
Takla Makan are in good agreement with observations for
these years. The high TOMS AI over eastern Siberia in
1987 is not captured by the simulation. This may be due
to either an unusual biomass burning event in this year or
a strong dust event not captured by the model. The
modeled dust AOT in the northwestern Sahara appears
to be underestimated in 1987 but overestimated in 1989.
Interannual changes in the atmospheric dust load are
for the most part determined by changes in the dust source
strength. Our simulations of dust transport were only made
for a limited period (1987–1990), but we have calculated
dust sources for a longer period from 1979 to 1993. We
have compared the dust fluxes (particles <10 µm radius) for
individual source regions with averaged TOMS AI (Figure
8) as simulated from our baseline experiment (where FPAR
for the individual years was used to calculate potential
source areas) from a simulation made using a 12-year
average of monthly NDVI FPAR and from a simulation
made using annual average FPAR to calculate the potential
dust source areas. The different treatment of vegetation
cover has no impact on Saharan emissions. For the Asian
dust regions of Gobi and Takla Makan the absence of the
vegetation seasonal cycle leads to much lower simulated
dust emissions, but the interannual changes are still matched.
The treatment of the vegetation cover has a significant
impact on the interannual changes in the dust signal over
Australia. However, both cases produce a reasonable
match to the observations. This is an indication that the
interannual vegetation changes that actually occurred in
the source regions during the period from 1982 to 1990
did not cause major changes in dust production.

3.4. Atmospheric Concentration

The seasonal dust aerosol mixing ratio for the first
atmospheric layer averaged for all modeled years captures
the well-known features of the global dust distribution
(Figure 9). The dominant feature of the simulation is the
Saharan dust plume that crosses the North Atlantic and
shifts seasonally. In the NH winter and spring, Saharan dust
is transported as far as the United States. High dust
concentrations are simulated over the Arabian peninsula
in late spring (during the season of “Shamal” winds [Pye,
1987]), while the maximum dust concentrations over the
Arabian Sea occurs later during the summer months. This
seasonal shift agrees with the maximum dust signal
(aerosol optical thickness product) registered by the
National Oceanic and Atmospheric Administration/AHRR
instrument in this region [Husar et al., 1997]. Asian dust
originating from the Gobi and Taklamakan deserts is tran-
sported over the North Pacific in the simulation, with a spring
maximum. This spring maximum is in good agreement with
observations that show maximum dust activity during the
spring Kosa events in Asia [e.g., Parungo et al., 1995]. The
location of a dust maximum around the Caspian and Aral
seas in spring agrees with the occurrence of dust storms in
this region of Central Asia [Pye, 1987]. In the United States
the simulated springtime maximum of atmospheric dust
occurs 1000 km further north than shown by visibility-
derived dust storm statistics [e.g., Orgill and Sehmel,

Figure 5. Daily AOT (left) and AI (right) for the
 modeled springtime maximum of atmospheric dust
occurs 1000 km further north than shown by visibility-
derived dust storm statistics [e.g., Orgill and Sehmel,
This discrepancy probably reflects the importance of agriculturally disturbed sites as sources of dust in the central United States. BIOME4 simulates potential vegetation, and we have not attempted to take land use into account in estimating source area locations in our simulation. In the SH, dust is transported from South America and Australia. The dust production and transport from these regions is smaller than shown by previous models of the global dust cycle [e.g.,

Figure 5. Comparison of daily dust distributions for (a) Saharan, (b) Asian, and (c) Australian dust episode. (top) Maps of daily TOMS A1 and (bottom) simulated aerosol optical thicknesses above 1.5 km height. Note that the color bar describes different units. See color version of this figure at back of this issue.
Tegen and Fung, 1994; Mahowald et al., 1999; Reader et al., 1999) and in better agreement with the dust loading implied by satellite retrievals of dust optical thickness [Husar et al., 1997] or the TOMS AI [Herman et al., 1997].

The modeled seasonal atmospheric dust concentration in the first atmospheric layer is generally in good agreement with surface observations at 21 sites operated by the University of Miami [Prospero, 1996], both in magnitude and timing (Figure 10). However, simulated dust concentrations in the western North Atlantic (Bermuda, Miami, Barbados) are up to an order of magnitude lower than observations during summer. Winter concentrations at Barbados are also underestimated, although the observed concentrations at Bermuda and Miami are reproduced well by the model. It seems unlikely that these discrepancies reflect incorrect placement of the simulated dust plume, given that we use analyzed wind fields. It is possible that simulated dust is not sufficiently mixed down from higher layers at these locations in summer. The rather good agreement between our simulations and observations at the elevated Izana station lends support to this hypothesis, as does the fact that modeled dust concentrations in summer at 2 km height at Barbados are quite similar to the surface concentrations observed at this station. However, modeled dust concentrations in the upper layers of the atmosphere over Bermuda and Miami are still considerably less than observed near the surface. Thus incorrect vertical transport of dust in our model can only explain part of the discrepancy with observations from the western North Atlantic. The model results agree relatively well with dust observations at remote North Pacific locations downwind of the Asian continent (Midway, Oahu, Enetewak, Fanning, Nauru) in both seasonality and magnitude. However, dust concentrations are underestimated at locations close to the Asian continent. At Cheju the simulated concentrations are half of the observed concentrations, while at Hedo the simulated concentrations are one quarter of the observed concentrations. Dust deflation at Cape Grim, south of Australia, is underestimated over the first part of the year.

3.5. Dust Deposition

The simulated annual dust deposition, averaged over the 4 years 1987–1990, shows peak deposition rates of up to 150 g m$^{-2}$ yr$^{-1}$ close to source regions in Central Asia and northern Africa (Figure 11a). The simulation shows significant regions of high deposition downstream from these source regions. Dust deposition fluxes at remote sites in the NH are of the order of 0.3–3.0 g m$^{-2}$ yr$^{-1}$. In the SH, dust deposition is also relatively high adjacent to the source areas in the Namib, Patagonia, and central Australia. However, since the SH source areas are of limited extent, the dust fluxes over most of the SH are generally low (<0.1 g m$^{-2}$ yr$^{-1}$). There is thus a significant difference in background levels of dust between NH and SH.

The simulated patterns of dust deposition are in rather good agreement with dust deposition rates from ocean sediment traps (Figure 11b). The observations show, for example, peak deposition rates of 10–27 g m$^{-2}$ yr$^{-1}$ off western Africa between 11° and 22°N [Jickells et al., 1996; Ratmeyer et al., 1999; Wefer and Fischer, 1993]. Both the magnitude and the positioning of this peak are captured in the simulation, which also shows the decline in deposition rates both with distance from the coast and
Figure 6. Comparison of daily model dust aerosol optical thickness (red line) with TOMS AI (blue line) averaged for eight regions over the period 1987–1990. See color version of this figure at back of this issue.
Figure 7. Anomaly maps showing the difference in (a) the aerosol optical thickness above 1.5 km and (b) the TOMS AI for NH spring for the individual years 1987–1990 minus the 4-year average values of these variables. These maps show whether the dust load was above or below average in a particular region during an individual year. See color version of this figure at back of this issue.
northward and southwards, shown by the data. The northwest-southeast gradient in observed dust deposition rates in the Arabian Sea is also well captured in the simulation. The model tends to underestimate the magnitude of dust deposition in the western Pacific close to the Asian source regions, but the latitudinal position of the plume appears to be consistent with the data available from this region. Although there are marine sediment trap data from the high-latitude SH [e.g., Fischer et al., 1988; Noriki and Tsunogai, 1986; Wefer and Fischer, 1993; Wefer et al., 1988], most of these sites do not provide good records of aeolian accumulation, as they are impacted by both resuspension and ice-rafted detritus. The limited observations that are available for aeolian sediment fluxes east of

Figure 8. Comparison of dust source fluxes at five locations for the period 1979–1982 through 1990 with averaged TOMS AI (solid line). Model results for the reference case (dashed line) are shown together with results where a 12-year average NDVI-derived FPAR was used to calculate potential source areas (dotted line), and with results which use annual average FPAR instead of monthly values to calculate the source areas (dash-dotted line).
Figure 9. Seasonal distribution of the modeled dust mixing ratio in the first atmospheric layer.
Figure 10. Comparison of the range of modeled dust concentrations at the first atmospheric layer over the years 1987–1991 (shaded area) and observed dust concentrations (open diamonds) at several sites. The shaded area envelops the minimum and maximum modeled concentrations at a given month, while the solid line is the 4-year model average. Shown are comparisons at locations, where the station data were available for at least 4 years.
Figure 11. (a) Dust deposition fluxes (4-year average for 1987–1990) compared with (b) sediment flux data. See color version of this figure at back of this issue.
Figure 12. Comparison of the range of modeled dust deposition fluxes over the years 1987–1990 (shaded area) and observed dust concentrations (open diamonds) at specific marine sites (for references, see Table 4). The shaded area envelops the minimum and maximum modeled deposition fluxes at a given month, while the solid line is the 4-year model average. The data points at each month indicate individual measurements.
Table 5. Model-Based Estimates of Global Annual Dust Emission

<table>
<thead>
<tr>
<th>Reference</th>
<th>Dust Emission, Mt yr^{-1}</th>
<th>Size Range (Radius)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>[Joussamue, 1990]</td>
<td>N/A</td>
<td>Size mode at 1 μm</td>
<td>Relative units</td>
</tr>
<tr>
<td>[Tegen and Fung, 1994]</td>
<td>3000</td>
<td>0.1–50 μm</td>
<td>Calibrated(^a)</td>
</tr>
<tr>
<td>[Tegen et al., 1996]</td>
<td>1200</td>
<td>0.1–10 μm</td>
<td>Calibrated(^a)</td>
</tr>
<tr>
<td>[Tegen and Miller, 1998]</td>
<td>940</td>
<td>0.1–8 μm</td>
<td>Calibrated(^d)</td>
</tr>
<tr>
<td>[Andersen et al., 1998]</td>
<td>2000</td>
<td>Size mode at 1 μm</td>
<td>Calibrated(^d)</td>
</tr>
<tr>
<td>[Mahowald et al., 1999]</td>
<td>3000</td>
<td>Size mode at 1.25 μm</td>
<td>Calibrated(^d)</td>
</tr>
<tr>
<td>[Reader et al., 1999]</td>
<td>430</td>
<td>&lt;2 μm</td>
<td>Calibrated(^d)</td>
</tr>
<tr>
<td>[Ginoux et al., 2001]</td>
<td>1800</td>
<td>0.1–6 μm</td>
<td>Calibrated(^d)</td>
</tr>
<tr>
<td>This work</td>
<td>800 (480)</td>
<td>0.1–1000 μm (0.1–10 μm)</td>
<td>Uncalibrated</td>
</tr>
<tr>
<td>This work</td>
<td>1700 (1100)</td>
<td>0.1–1000 μm (0.1–10 μm)</td>
<td>Maximum(^b)</td>
</tr>
</tbody>
</table>

\(^a\)Emission factor calibrated to match previous model-based global emission estimates.
\(^b\)Emission factor calibrated to match observations.
\(^c\)Maximum emission to match observations.

Australia and west of South Africa show slightly higher values than the model results.

Comparison of the seasonal distribution of dust deposition fluxes in different ocean basins with sediment trap data [Ratmeyer et al., 1999; Haake et al., 1993] shows good agreement at most locations (Figure 12). Simulated deposition is lower than observations during winter/spring at one location in the tropical North Atlantic (Figure 12e) and lower during fall/winter at one site in the eastern Indian Ocean (Figure 12k). This may indicate dust emissions from upwind dust sources that are too low. In the tropical Atlantic (Figures 12f and 12g), coastal plankton species were found in the sediment trap data, which may indicate sediment mixing from the coastal region [Wefer and Fischer, 1993]. For the Pacific the agreement between seasonal sediment fluxes is mostly good; however, in the tropical Pacific the dust fluxes are at least an order of magnitude higher than the observations (Figures 12o, 12p, and 12q). In this region the seasonal cycle indicates that the dust sediment fluxes are influenced by SH rather than Asian sources. The atmospheric station data from this region (Funafuti, American Samoa) (Figure 10) show very low dust fluxes, which is inconsistent with the relatively high sedimentation rates found in the marine sediments.

4. Discussion

The total dust emission simulated by our model is 805 ± 82 Mt yr^{-1}. This value is at the low end of estimates of global dust emissions published in the literature, which range from 60 to 3000 Mt yr^{-1} [see e.g., Duce, 1995]. This extraordinarily large range suggests that there are significant problems or uncertainties associated with the published global estimates.

Some of the previously published estimates of the global dust loading have been derived by extrapolation from regional measurements of dust loading and/or deposition. There are, however, significant discrepancies even between estimates for specific regions. Dust deposition rates to the North Pacific, for example, have been estimated as 6–12 Mt yr^{-1} [Uematsu et al., 1983] and 480 Mt yr^{-1} [Duce et al., 1991]. The larger of these estimates may be compatible with the independent estimate of 67 Mt yr^{-1} deposition to the more limited area of the China Sea [Gao et al., 1997], but clearly the lower estimate is not. Similar uncertainties exist for estimates of Saharan dust. Dust deposition to the North Atlantic has been estimated to be 220 Mt yr^{-1} [Duce et al., 1991]. This is rather low compared with the amount of dust transported from the Sahara, which has been estimated as 600–700 Mt yr^{-1} [d’Almeida, 1986]. The differences in the magnitude of the regional estimates reflects the fact that each is based on a limited number of point observations, each of which covers only a limited period of time. The problem is further compounded when poorly constrained estimates from specific regions, which are not necessarily typical of all dust source regions, are extrapolated to yield a global estimate.

Estimates of the global dust loading based on dust cycle model simulations potentially provide a way of avoiding the uncertainties caused by extrapolation from a spatially and temporally limited sampling of regional dust patterns. Unfortunately, all of the published model-based estimates (see Table 5) have been calibrated in some way, either to match previous global emission estimates [e.g., Andersen et al., 1998; Mahowald et al., 1999] or to match selected observations [e.g., Tegen and Fung, 1994; Reader et al., 1999]. The substantial range (430 to 3000 Mt yr^{-1}) in the estimates of the total dust emission, even among those models that were tuned to match observations, strongly suggests that the amount of data used was insufficient to provide a strong constraint on the calibration. Thus it is clear that all of the previous estimates of the global dust loading are poorly constrained. There is no reason to assume our uncalibrated estimate of 805 ± 82 Mt yr^{-1} for the total global dust loading is unrealistic.

Nevertheless, our systematic comparisons with atmospheric concentration records from individual sites (Figure 10) and deposition data from the DIRTMAP data set (Figure 11) suggest that our uncalibrated estimate may be somewhat too low. Site-by-site comparisons in the remote Pacific and for dust exported from the northern Sahara to the North Atlantic show good agreement, but the annual mean transport of Saharan dust to the western North Atlantic is underestimated by a factor of 2 to 6. Dust transport to locations south of Australia, in the Arabian Sea, and in the western Pacific Ocean is also underestimated by a factor of 2 to 4.

There are at least two reasons why the regional emissions simulated by our model might be lower than implied by the observations. First, the simulations only take
natural sources of mineral dust into account. It has been estimated that 20–50% of observed dust emissions are the result of anthropogenic disturbance of soil surfaces, e.g., through agricultural practices [Tegen and Fung, 1995; Sokolik and Toon, 1996]. The most significant discrepancies between our simulations and the observations do indeed occur in regions which are heavily impacted by humans, including North America and the Sahel. A more fundamental cause of the discrepancies may be underestimation of wind strengths. We have used 6-hourly ECMWF ERA winds to drive the emissions model. Despite the relatively high spatial and temporal resolution of the ECMWF data set, these data underestimate peak wind speeds. Middleton [1986], for example, has reported maximum wind speeds of >40 m s\(^{-1}\) during dust storm events in southwest Asia. The maximum wind speed in this region according to the ECMWF data set was only 27 m s\(^{-1}\). Similarly, the wind gusts at the leading edge of squall lines crossing the Sahara in the summer months have been observed to reach 30 m s\(^{-1}\) [Sommeria and Testud, 1984], whereas the maximum wind speed in this region according to the ECMWF data set is only 15 m s\(^{-1}\). There are relatively few meteorological stations in most of the dust source regions, and the winds in the ECMWF data set are therefore extrapolated via model physics from observation sites that may be rather distant. Thus reconstructed wind speeds are likely to be less realistic than in regions where the reanalysis model is constrained by larger numbers of observations [Trenberth and Olson, 1988]. Dust emissions are particularly sensitive to small differences in wind speed because of the \(u^3\) relationship between emission flux and surface wind stress. Thus the large discrepancies between ECMWF wind speeds and observed wind speeds during dust events would be expected to have a significant impact on our simulation. We were able, for example, to more than double the simulated total global emission in a sensitivity test (not shown) in which the ECMWF wind speeds over all source areas were arbitrarily doubled for only three of the 6-hourly time steps during one year.

[52] Given sufficient observations, it would be possible to use inverse modeling tools in an attempt to provide a better-constrained estimate of the global dust loading. The data set assembled here represents the largest compilation of quantitative observations of the modern global dust cycle available, but some regions are still too poorly sampled to adequately constrain an inverse model. We have used a more subjective approach in an attempt to place a plausible upper limit on the global dust loading. For each of the regions for which we have evidence of a significant mismatch between our simulated dust loading and/or deposition and the observed dust loading and/or deposition (i.e., southern Sahara, Arabian Sea, China, and Australia), we increase the simulated total emissions by a factor corresponding to the maximum discrepancy observed at any site. Thus we increase emissions from the southern Sahara by a factor of 10 and emissions from each of the three other regions by a factor of 4. This results in an estimated total global dust loading of 1700 Mt yr\(^{-1}\), compared with our uncalibrated estimate of 800 Mt yr\(^{-1}\). Assuming there is no change in the size distribution compared with the uncalibrated model results, 1100 Mt yr\(^{-1}\) would consist of particles with radii of <10 \(\mu m\). We believe that these estimates place a maximum upper bound on the global dust loading. Thus we suggest that 800–1700 Mt yr\(^{-1}\) is the plausible range of total global dust emission under modern climate conditions and that estimates lower or higher than this are probably erroneous.

[53] Dust emissions show considerable interannual variability. Our simulations reproduce the observed interannual changes in the global dust load, as shown by the TOMS aerosol index (Figure 7), rather well. Furthermore, the interannual patterns of dust loading over the southern Sahara, the Mediterranean, and the Takla Makan are in good agreement with observations. However, the interannual variability of dust emissions from the northern Sahara is not well captured in the simulation. Visibility studies in the Sahel [Mbourou et al., 1997] and observations of dust concentrations at Barbados [Prospero and Neees, 1986] indicate that dust emissions from the Sahara/Sahel were lower during the late 1970s than in later years. This change in dust emissions is not reproduced in our simulations (Figure 8), despite the fact that we take interannual changes in surface wind strengths and vegetation cover into account. It seems unlikely that our inability to capture the observed shift in the magnitude of emissions during the early 1980s is caused by errors in the climate data used to determine vegetation cover or in the surface winds, given the correct simulation of other regions. We therefore speculate that it may be a result of our treatment of preferential source areas, which are not allowed to vary in extent as a consequence of climatically induced changes in lake extent. In the majority of situations, recent changes in lake extent are very small compared with the changes in lake extent experienced during the Late Quaternary and thus have a negligible impact on the total area of a given preferential source. However, Lake Chad has decreased in area over the last 30 years, and the change in area since 1983 has been substantial [Coe and Foley, 2001]. As a result, there has been an expansion of dust sources in the Lake Chad basin, which may have had an impact on dust emissions. Thus, in order to capture observed changes in interannual dust loadings in regions characterized by large, low-lying lake basins, it may be necessary to simulate climatically induced changes in lake extent.

5. Conclusions

[54] The ability to correctly simulate the modern dust cycle is a prerequisite for predicting changes in dust emissions in response to known changes in climate forcing during the past or anthropogenically induced changes in the future. Using a model that takes seasonal changes in vegetation cover and the existence of preferential source areas in dry lake beds into account, we have shown that it is possible to simulate modern global dust distribution patterns on daily, seasonal, and interannual timescales in qualitatively good agreement with observations. Vegetation phenology is shown to be important, for example, in capturing the spring dust emission events from central Asia. Dry lake beds are indicated to make a substantial contribution to global dust emissions and to be responsible for known “hot spots” of dust emission. Modeling dust emissions in past or future climates will therefore require computation of sea-
sonal vegetation phenology and preferential source areas. Future dust scenarios will also need to take into account changes in dust emissions caused by changes in land use.

Although the model reproduces the magnitude of dust emissions and transport to several key regions (e.g., the northern Sahara, the remote Pacific), it underestimate deposition to the western North Atlantic, south of Australia, in the Arabian Sea, and in the Pacific close to the Asian continent. The underestimation of emissions implied by this mismatch between the simulated and observed deposition fluxes is likely related to the underestimation of wind strengths in the ECMWF data set used to drive the model.

The total global dust emission simulated by the model is 800 Mt yr

1. We suggest that the range of 800–1700 Mt yr

1 represents a more realistic estimate of global dust emissions than the extreme values of 60 Mt yr

1 and 3000 Mt yr

1 previously reported in the literature.

Acknowledgments. This work is a contribution to the MAGIC (Mineral Aerosols in Glacial-Interglacial Cycles) project, which is partly funded by the Swedish Natural Science Research Council (NFR). The DIRTMAP database is partly sponsored by the Max Planck Society and by NFR. The dust concentration data from the AEROCE program were provided by J. Prospero and D. Savoie, University of Miami. We thank G. Bönsch for assistance with the DIRTMAP database.

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Thompson, R. S., C. Whitlock, P. J. Bartlein, S. P. Harrison, and W. G. Spaulding, Climatic changes in the western united states since 18,000 yr bp., in Global Climates Since the Last Glacial Maximum, edited by H. E. Wright Jr. et al., pp. 468–513, Univ. of Minn. Press, Minneapolis, 1993.


Figure 1. Areal coverage of preferential dust sources, expressed as percentage of each 0.5° × 0.5° grid cell, calculated from the extent of potential lake areas, excluding areas of actual lakes and forest biomes.
Figure 2. (a) Annual average dust emission flux for the year 1988 for particles smaller than 2 $\mu$m radius (b) compared with dust emission flux excluding preferential source areas.
Figure 3. (a) Difference between annual dust emissions for the case of seasonally invariant vegetation cover, (b) vegetation cover consisting only of grasses, and (c) vegetation cover consisting only of shrubs, and the reference case for the year 1988.

Dust emission flux (g/m²)
Figure 5. Comparison of daily dust distributions for (a) Saharan, (b) Asian, and (c) Australian dust episode. (top) Maps of daily TOMS AI and (bottom) simulated aerosol optical thicknesses above 1.5 km height. Note that the color bar describes different units.
Figure 5. (continued)
Figure 6. Comparison of daily model dust aerosol optical thickness (red line) with TOMS AI (blue line) averaged for eight regions over the period 1987–1990.
Figure 7. Anomaly maps showing the difference in (a) the aerosol optical thickness above 1.5 km and (b) the TOMS AI for NH spring for the individual years 1987–1990 minus the 4-year average values of these variables. These maps show whether the dust load was above or below average in a particular region during an individual year.
Figure 7. (continued)
Figure 11. (a) Dust deposition fluxes (4-year average for 1987–1990) compared with (b) sediment flux data.