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Authors

Zender, Charles S
Bian, H.
Newman, D.

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Mineral Dust Entrainment and Deposition (DEAD) model: Description and 1990s dust climatology

Charles S. Zender, Huisheng Bian, and David Newman

Department of Earth System Science, University of California at Irvine, Irvine, California, USA

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[1] We describe a model for predicting the size-resolved distribution of atmospheric dust for climate and chemistry-related studies. The dust distribution from 1990 to 1999 is simulated with our mineral aerosol entrainment and deposition module embedded in a chemical transport model. Mobilization processes include entrainment thresholds for saltation, moisture inhibition, drag partitioning, and saltation feedback. For mobilization we assume that soil texture is globally uniform and is replete with saltators. Soil erodibility is prescribed by a new physically based geomorphic index that is proportional to the runoff area upstream of each source region. Dry deposition processes include sedimentation and turbulent mix-out. Nucleation scavenging and size-resolved washout in both stratiform and convective cloud types are represented. Simulations of the 1990s broadly agree with station observations and satellite-inferred dust distributions. Without invoking anthropogenic mechanisms the model captures the seasonal migration of the transatlantic African dust plume, and it captures the spring maximum in Asian dust outflow and concentration over the Pacific. We estimate the 1990s global annual mean and variability of $D < 10 \mu\text{m}$ dust to be the following: emissions, $1490 \pm 160 \text{ Tg yr}^{-1}$; burden, $17 \pm 2 \text{ Tg}$; and optical depth at $0.63 \mu\text{m}$, 0.030 ± 0.004 . This emission, burden, and optical depth are significantly lower than some recent estimates. The model underestimates transport and deposition of East Asian and Australian dust to some regions of the Pacific Ocean. An underestimate of long-range transport of particles larger than $3 \mu\text{m}$ contributes to this bias. Our experiments support the hypothesis that dust emission “hot spots” exist in regions where alluvial sediments have accumulated and may be disturbed.

INDEX

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1. Introduction

[2] Mineral dust aerosol is involved in many important processes in Earth's climate system. These processes include direct radiative forcing [Tegen *et al.*, 1996], nutrient transport [Martin, 1990; Swap *et al.*, 1992], land-use change [Nicholson *et al.*, 1998], and ecosystem health [Prospero, 1999; Shinn *et al.*, 2000]. Simulations of the distribution of mineral dust help improve our understanding of the role of dust in these processes, and the behavior of these processes in past, present, and future climates. In this paper we describe the Dust Entrainment and Deposition (DEAD) model, designed for studying dust-related processes at both local and global scales. We present the underlying physical assumptions of DEAD and evaluate predictions of a 1990–1999 global natural dust simulation.

[3] Models of the global distribution of mineral dust must successfully couple the microphysical processes of entrainment and deposition. In global models, soil and meteorological conditions are only available at coarse spatiotemporal resolution and with great uncertainties attached. Dust models are often differentiated by their representation of mobilization. At least two distinct classes of mobilization schemes exist. The simpler class parameterize mobilization in terms of the third or fourth power of the wind speed or wind friction speed and then impose an empirical size distribution upon the emitted dust. We call these bulk mobilization schemes, and they include those of Tegen and Fung [1994], Mahowald *et al.* [1999], and Perlwitz *et al.* [2001]. The more complex class use complete microphysical specification of the erodible environment to predict the size-resolved saltation mass flux and resulting sandblasted dust emissions [Marticorena and Bergametti, 1995; Shao *et al.*, 1996; Shao, 2001]. Many of the input parameters for these fully microphysical

schemes are not known globally, but these schemes have shown promising results in regional simulations [Shao and Leslie, 1997; Marticorena et al., 1997]. Models intermediate in complexity between these two extremes, including the present model, are now in use at the global scale [Ginoux et al., 2001; Woodward, 2001; C. Luo et al., A 22-year climatology of mineral aerosols, submitted to *Journal of Geophysical Research*, 2003] (hereinafter referred to as Luo et al., submitted manuscript, 2003). Generally, these intermediate complexity dust models use microphysical parameterizations where possible, but make simplifying assumptions to produce adequate global simulations.

[4] Herein we describe the physics and the mean climatology of the Dust Entrainment and Deposition (DEAD) model in the present climate. In this study, DEAD is implemented as a component of the Model for Atmospheric Chemistry and Transport (MATCH) Chemical Transport Model (CTM) [Rasch et al., 1997], which is driven by National Center for Environment Prediction (NCEP) analyzed meteorology for the period 1990–1999 [Kalnay, 1996]. Using observational analyses (rather than predicted wind) minimizes meteorology-induced biases in the resulting dust distribution, and highlights deficiencies in the model representation of the mobilization and deposition processes. Quantifying these biases helps to reveal where more study is necessary.

[5] Earlier versions of DEAD were used in aerosol assimilation studies of INDOEX [Rasch et al., 2001; Collins et al., 2001, 2002]. A modified version of DEAD has recently been used to study decadal variations in the 30-year dust record at Barbados [Mahowald et al., 2002], the mean climatology of a 22-year time series (Luo et al., submitted manuscript, 2003), and interannual variability in current climate dust production [Mahowald et al., 2003]. Studies of the PRIDE and ACE-Asia field experiments are ongoing.

[6] Section 2 describes the physical basis and formulation of the dust entrainment and deposition model. Section 3 presents the 1990–1999 simulated dust climatology and global budgets. Section 4 evaluates the simulated dust climatology with station observations of dust concentration, deposition, and optical depth. Section 5 summarizes our results and their implications for future research.

2. Dust Prediction

[7] Predicting the distribution of mineral dust in the atmosphere requires representation of source, sink, and transport processes. The source process for dust is mobilization by wind. Sink processes are dry deposition (gravitational sedimentation and turbulent mix-out), and wet deposition (in-cloud and below-cloud scavenging). These processes are described in this section.

[8] Dust is advected and diffused as a passive scalar quantity by the transport processes and methods used in the host model. For MATCH, these processes include mixing within the planetary boundary layer [Holtslag and Boville, 1993], shallow convective transport [Hack, 1994], a plume ensemble method of deep convection [Zhang and McFarlane, 1995], and advection by large-scale winds [Rasch et al., 2001]. These transport mechanisms are nearly

identical in both the MATCH and the Community Climate Model (CCM) versions of DEAD.

2.1. Mobilization

[9] Adequate representation of the long-lived dust burden requires careful attention to the process of dust mobilization [Schulz et al., 1998]. A large number of factors, many of them poorly known on a global scale, determine soil erodibility and dust emissions. The most important factors include wind friction speed, vegetation cover, and surface soil moisture content. We generally follow the microphysical and micrometeorological approach to dust mobilization developed by Marticorena and Bergametti [1995] (hereinafter referred to as MaB95).

[10] Dust entrained into the atmosphere originates in source soils which contain the clay-sized ($D_p < 2.5 \mu\text{m}$) and silt-sized ($2.5 < D_p < 60 \mu\text{m}$) particles whose atmospheric residence time exceeds about 20 min (i.e., the temporal resolution of the host model). However, clay and silt-sized particles are not directly mobilized by the wind because cohesive forces (e.g., capillary and electrostatic) bind these particles tightly to the soil. Laboratory [Iversen and White, 1982] and field [Shao et al., 1996] wind tunnel studies show that dust is primarily injected into the atmosphere during the sandblasting caused by saltation bombardment [Alfaro and Gomes, 2001; Grini et al., 2002]. Sandblasting refers to the disaggregation and ejection of clay and silt particles by saltating sand-sized particles ($D_p > 60 \mu\text{m}$).

[11] The first steps of the source scheme are therefore the prediction of saltation events and their intensity. The condition for saltation initiation is that the turbulent drag of the surface atmosphere dissipates enough momentum to overcome the gravitational inertia of sand-sized particles. The solution of the boundary layer turbulence problem provides required surface drag properties in the form of the modeled wind friction speed u_* . To obtain a u_* suitable for dust mobilization, DEAD re-solves the boundary layer turbulence problem independently of the host model (i.e., MATCH) using boundary conditions more suitable for dust producing regions. The surface roughness length in dust producing regions is set to the globally uniform value of $z_{0,m} = 100 \mu\text{m}$, a value more typical of erodible soil beds [Gillette et al., 1997] than the large-scale roughness lengths for bare ground ($z_{0,m} \sim 5 \text{cm}$) used in general circulation models [e.g., Bonan, 1996]. We obtain the kinematic and thermodynamic properties of the boundary layer by assuming that the surface and atmosphere constantly adjust surface heat, vapor, and momentum exchanges in order to maintain thermal equilibrium with the radiation field [Bonan, 1996].

2.2. Threshold Friction Velocity

[12] The mass flux of saltating particles Q_s (see equation (10), below) depends on the excess of the wind friction speed u_* over the threshold wind friction speed for saltation, u_{*t} . The relation between soil particle size and u_{*t} has been measured in wind tunnels [Bagnold, 1941; Greeley and Iversen, 1985; Shao et al., 1996; Batt and Peabody, 1999]. Iversen and White [1982] developed a semi-empirical parameterization of this relationship in terms of the threshold friction Reynolds number $\text{Re}_{*t} \equiv u_{*t}D/\nu$ where ν

is the kinematic viscosity of air. DEAD employs this parameterization in the computationally amenable form

$$u_{*t}(D) = \begin{cases} \left[\frac{0.1666681\rho_p g D}{-1 + 1.928 \text{Re}_{*t}^{0.0922}} \left(1 + \frac{6 \times 10^{-7}}{\rho_p g D^{2.5}} \right) \right]^{1/2} \rho^{-1/2} & : 0.03 \leq \text{Re}_{*t} \leq 10 \\ \left[0.0144\rho_p g D (1 - 0.0858e^{-0.0617(\text{Re}_{*t}-10)}) \left(1 + \frac{6 \times 10^{-7}}{\rho_p g D^{2.5}} \right) \right]^{1/2} \rho^{-1/2} & : \text{Re}_{*t} > 10 \end{cases} \quad (1)$$

Equation (1) is arranged so that all the microphysical properties are in the first term on the right hand side. This term contains all properties determined by soil size and density and thus needs to be evaluated only once (at model start) for a given saltating particle size. The $\rho^{-1/2}$ term is the same for all particle sizes, but depends on ambient, time-varying environmental conditions. Since Re_{*t} is defined in terms of u_{*t} , equation (1) is an implicit definition of u_{*t} which must be solved iteratively.

[13] The optimal particle size for saltation D_0 occurs where u_{*t} is at a minimum. Solutions to equation (1) show that, for typical conditions on Earth, $D_0 \approx 75 \mu\text{m}$ [Iversen and White, 1982]. Following MaB95, we assume all soils in erodible regions contain particles of size D_0 , so that saltation is initiated whenever $u_* > u_{*t}(D_0)$. To accelerate solution of equation (1), we compute $u_{*t}(D_0)$ using the noniterative parameterization for Re_{*t} introduced in MaB95.

[14] Three processes modify u_* and u_{*t} : drag partitioning, the Owen effect, and moisture inhibition. First, a drag partition parameterization is applied to represent the sink of atmospheric momentum into nonerodible roughness elements [Raupach, 1992]. We consider two roughness lengths pertinent to dust emissions from erodible surfaces. The first is the aerodynamic roughness length of the bare ground including the nonerodible elements such as pebbles, rocks, and vegetation. This is traditionally known as the roughness length for momentum transfer, $z_{0,m}$. The second roughness length is the so-called ‘‘smooth’’ roughness length, $z_{0,m}^s$ [Marticorena and Bergametti, 1995]. $z_{0,m}^s$ is the roughness length of a bed of potentially erodible particles without any nonerodible elements. Wind tunnel experiments over uniform beds comprised of known particle sizes show that

$$z_{0,m}^s \approx D/30 \quad (2)$$

We use globally uniform values for $z_{0,m}$ and $z_{0,m}^s$ of $100.0 \mu\text{m}$ and $33.3 \mu\text{m}$, respectively. From equation (2), this corresponds to particle beds of particle area-mean size of $D \sim 1 \text{mm}$.

[15] The efficiency with which drag is partitioned between erodible and nonerodible soils is expressed as an increase f_d in the threshold friction speed for saltation u_{*t} [Marticorena and Bergametti, 1995]

$$f_d = \left[1.0 - \left(\frac{\ln(z_{0,m}/z_{0,m}^s)}{\ln\left\{0.35 \left[(0.1/z_{0,m}^s)^{0.8} \right]\right\}} \right) \right]^{-1} \quad (3)$$

[16] The second process which modifies u_* and u_{*t} is the Owen effect, which refers to the positive feedback of saltation upon surface roughness length and friction speeds [Owen, 1964]. On the basis of field measurements at Owens Dry Lake, California, Gillette *et al.* [1998] found that the increase in wind friction speed due to saltation varies quadratically with the difference between the 10 m wind speed U_{10} and the threshold wind speed at 10 m $U_{10,t}$ as

$$u_{*,s} = u_* + 0.003(U_{10} - U_{10,t})^2 \quad (4)$$

The Owen effect, $\Delta u_* = u_{*,s} - u_*$, affects only the saltation fluxes and does not affect heat, moisture, and momentum exchange in the large-scale host model. This approximation will be removed in future versions of DEAD.

[17] Finally, the inhibition of saltation by soil moisture is accounted for by increasing u_{*t} in moist soils. A number of investigators have created simple parameterizations which account for the increase of u_{*t} with soil water [Belly, 1964; Pye, 1987; Gillette, 1988; Selah and Fryrear, 1995; Shao *et al.*, 1996; Fécan *et al.*, 1999]. We adopt the parameterization of Fécan *et al.* [1999] who used measurements to specify free parameters in the adsorptive theory of McKenna-Neuman and Nickling [1989]. In our model, the capillary force is allowed to suppress dust deflation when the near-surface soil gravimetric water content w exceeds a threshold w_t determined by

$$w_t = a \left(0.17M_{\text{clay}} + 0.14M_{\text{clay}}^2 \right) \quad (5)$$

where the parenthetical factor is directly from Fécan *et al.* [1999] and a is an ad hoc factor chosen to improve model simulations. The increase f_w in threshold friction velocity for saltation u_{*t} due to soil water is [Fécan *et al.*, 1999]

$$f_w = \begin{cases} 1 & : w \leq w_t \\ \sqrt{1 + 1.21[100(w - w_t)]^{0.68}} & : w > w_t \end{cases} \quad (6)$$

In these simulations, the CTM uses upper layer volumetric soil water content $\theta \text{ m}^3 \text{ m}^{-3}$ interpolated from the 6-hourly NCEP/NCAR reanalysis data. It is necessary to convert the NCEP volumetric water content θ to the gravimetric water content (dry mass basis) $w \text{ kg kg}^{-1}$ used in equations (5) and (6). This requires knowledge of the volume of air (pores) per unit volume soil. We assume that the volumetric water content of the soil at saturation, $\theta_s \text{ m}^3 \text{ m}^{-3}$ equals the porous volume of air in dry soil, i.e., soil is saturated when all interparticle pores are filled with water. Under this assumption, the equivalence between θ and w is

$$\theta_s = 0.489 - 0.126M_{\text{sand}} \quad (7)$$

$$\rho_{b,d} = \rho_p(1.0 - \theta_s) \quad (8)$$

$$w = \theta_{pl}/\rho_{p,d} \quad (9)$$

where $M_{\text{sand}} \text{ kg kg}^{-1}$ is the mass fraction of sand in the soil [Bonan, 1996; Global Soil Data Task, 1999], $\rho_p = 2500 \text{ kg m}^{-3}$

is the mean soil particle density, $\rho_l = 1000 \text{ kg m}^{-3}$ is the density of liquid water, and $\rho_{b,d} \text{ kg m}^{-3}$ is the bulk density of dry soil. Thus sandier soils have smaller w for a given θ .

[18] The NCEP soil moisture w usually exceeds $0.1 \text{ m}^3 \text{ m}^{-3}$ over active dust emission regions all year long. We have found empirically that setting $a = 5$ in equation (5) gives reasonable results in global-scale models driven by NCEP w . The NCEP θ is only available at two model layers, every six hours, and is based on a soil texture, hydrologic model, and precipitation fields that differ from those employed in DEAD and MATCH, respectively. This inconsistency is a potential cause of bias in our results. Embedding DEAD in a host model with a consistent land surface representation removes this inconsistency.

2.3. Saltation and Dust Production

[19] Primary dust production, i.e., the direct mobilization of small particles by wind, is negligible on Earth because $u_{*t(D)}$ (equation (1)) is usually too large [Greeley and Iversen, 1985]. Instead, the vertical flux of fine dust is dominated by a secondary process called sandblasting, which refers to the release of small particles by large particles in saltation [Gomes *et al.*, 1990]. Sandblasting and disaggregation of small clay and silt-sized particles from the surface and from larger particles during saltation is strongly sensitive to the size distribution of saltating particles [Shao *et al.*, 1996; Grini *et al.*, 2002].

[20] DEAD simulates the total horizontally saltating mass flux of large particles Q_s according to the theory of White [1979]

$$Q_s = \frac{c_s \rho u_*^3}{g} \left(1 - \frac{u_{*t}}{u_*}\right) \left(1 + \frac{u_{*t}}{u_*}\right) \quad (10)$$

where, $c_s = 2.61$, ρ is the atmospheric density, u_* is the wind friction speed, g is acceleration of gravity, and u_{*t} is the threshold wind friction speed. This relationship assumes linearity between the vertical mass flux of saltators and the wind friction excess $u_* - u_{*t}$, an assumption borne out by microphysical saltation models as well as wind tunnel studies [White, 1979]. DEAD assumes that total Q_s (equation (10)) is determined solely by the threshold friction speed for the optimal particle size $u_{*t(D_0)}$ (equation (1)). More detailed models predict a size-resolved Q_s [Shao and Leslie, 1997; Marticorena *et al.*, 1997], but this requires global knowledge of the parent soil texture.

[21] The horizontal (saltation) mass flux Q_s (equation (10)) is converted to a vertical dust mass flux F_d with an efficiency α , called the sandblasting mass efficiency [Alfaro *et al.*, 1997], i.e., $F_d = \alpha Q_s$. Observations reveal that F_d exhibits high sensitivity to parent soil texture [Shao *et al.*, 1993] and wide scatter with increasing u_* [Alfaro and Gomes, 2001; Grini *et al.*, 2002]. DEAD adopts the size- and drag-independent α parameterization of MaB95

$$\alpha = 100 \exp[(13.4 M_{\text{clay}} - 6.0) \ln 10] \quad (11)$$

where the mass fraction of clay particles in the parent soil is restricted to $M_{\text{clay}} < 0.20$. The sandblasting mass efficiency α (equation (11)) increases by nearly 3 orders of magnitude

as the parent soil texture changes from $M_{\text{clay}} = 0.0$ (sand) to 0.20 (sandy loam). This parameterization yields reasonable results when applied in regional models [Marticorena *et al.*, 1997] where soil characterization is reliable, but proves to be overly sensitive to M_{clay} in our global model. Therefore we use a globally uniform value of $M_{\text{clay}} = 0.2$ to determine α in equation (11). Without this assumption, dust emissions, burden, and optical depth would be from 2–15 times higher in East Asia relative to North Africa, and would degrade the fidelity of the simulation. This assumption will be relaxed when more reliable global erodible soil data sets become available.

[22] Currently we use the IGBP soil texture data set [Global Soil Data Task, 1999]. This data set has less clay in the Saharan region of Africa than the soil data set of Webb *et al.* [1993]. The sandier IGBP soils have a reduced threshold for evapotranspiration, and consequently allow drier surfaces.

[23] DEAD assumes that saltation leads to dust production whenever $u_* > u_{*t(D_0)}$ over bare ground. This assumption means that soils depleted in particles of size D_0 will begin saltation at unrealistically low wind friction speeds.

[24] The physics of entrainment in global dust models differ significantly. It is helpful to summarize the basic differences so that the origin of inter-model differences may be understood. Many earlier studies [Genthon, 1992; Tegen and Fung, 1994; Mahowald *et al.*, 1999] based $F_d(D)$ on a cubic or quartic relationship with wind speed U . A uniform threshold wind speed U_t was commonly used. Woodward [2001] predicts a saltation flux $Q_s(D)$ (equation (10)) for each dust size class by assuming the saltation layer mass distribution equals the parent soil mass distribution. These horizontal fluxes $Q_s(D)$ are converted to vertical fluxes using the local soil clay content as in equation (11). Ginoux *et al.* [2001] do not explicitly model saltation. Instead, they apply a size-dependent threshold velocity u_{*t} for dust particles (rather than saltation-sized particles) which decreases with decreasing particle size. In effect, the wind directly entrains dust-sized particles into the atmosphere. This allows small dust particles to be entrained by weak wind events, relative to wind tunnel observations [Iversen and White, 1982].

2.4. Particle Size Distributions

[25] DEAD uses a bin-method to independently transport discrete, noninteracting, size (mass) classes. Each bin has an independently configurable subbin distribution which allows more accurate treatment of particle number and optical properties when the number of transport bins is small. The number of transport bins is arbitrary, limited only by the computational requirements of the model. Although the entrainment of mineral dust aerosol is initiated by the saltation of sand-sized particles, only particles with sizes $D_p \lesssim 10 \mu\text{m}$ reside in the atmosphere long enough to travel significant distances downwind. The vertical dust flux F_d obtained from equations (10) and (11) is assumed to be size-distributed in an analytic, trimodal lognormal probability density function (PDF) which is globally uniform. This PDF comprises the “background” modes of dust suggested by D’Almeida [1987]. These three modes are hereafter called the source modes. Table 1 lists the parameters of dust in the source regions. Shown are the number median diameter \bar{D}_n ,

Table 1. Tri-modal Size Distribution in Source Regions^a

\tilde{D}_s , μm	\tilde{D}_v , μm	σ_g Fraction	M Fraction
0.16	0.832	2.10	0.036
3.19	4.82	1.9	0.957
10.0	19.38	1.6	0.007

^aBackground dust mode of *D'Almeida* [1987].

mass median diameter \tilde{D}_v , geometric standard deviation σ_g , and mass fraction M of each mode.

[26] Note that the middle mode, which dominates long-range transport, has $\sigma_g = 1.9$, which is close to the $\sigma_g = 2.0$ found to produce good agreement in visible optical depth with satellite observations [*Schulz et al.*, 1998].

[27] The host model carries dust in J transport bins. The number J is empirically chosen to balance computational resources against desired accuracy in the prediction of dust particle number, surface area, and volume (mass). As a bin method, this procedure converges to an exact representation of transport of the source modes as J increases. The mass fraction of each source mode i that is carried in each sink (transport) bin j is the mass overlap $M_{i,j}$. Since the mass in the source modes is assumed to be lognormally distributed, the $M_{i,j}$ are [*Schulz et al.*, 1998]

$$M_{i,j} = \frac{1}{2} \left[\operatorname{erf} \left(\frac{\ln(D_{\max,j}/\tilde{D}_{v,i})}{\sqrt{2} \ln \sigma_{g,i}} \right) - \operatorname{erf} \left(\frac{\ln(D_{\min,j}/\tilde{D}_{v,i})}{\sqrt{2} \ln \sigma_{g,i}} \right) \right] \quad (12)$$

where erf is the standard error function, $D_{\min,j}$ and $D_{\max,j}$ are the minimum and maximum diameters of bin j , and $\tilde{D}_{v,i}$ and $\sigma_{g,i}$ are the mass median diameter and geometric standard deviation of the source modes (Table 1). The total transported mass fraction of the source modes is $\sum \sum M_{i,j} = M$. If all of the mass in the I source modes is contained in the size range $[D_{\min,0}, D_{\max,J}]$ then $M = 1$. In the present study, $0.1 < D < 10.0 \mu\text{m}$ and $M = 0.87$. The residual, nontransported mass (i.e., 13%) is almost all in the coarse mode centered at $19 \mu\text{m}$ (Table 1) which is not important for long-range transport [*Schulz et al.*, 1998].

[28] Within each transport bin, particles are assumed to have an analytic, time-invariant, subbin distribution. While the absolute mass in each bin changes every time step because of size-dependent source and sink processes (e.g., sedimentation), the assumed subbin distribution within each bin never changes. Figure 1 shows the simulated transported size distribution at Barbados for July. Approximately 80% of the transported mass is in sizes $D > 2.5 \mu\text{m}$ in African source regions. Sedimentation and preferential scavenging of large particles (see Table 3) has reduced the mass in the silt bins relative to the clay bins during transport, but the assumed subbin distributions of number, surface area, and mass have not changed. Only 50% of the mass remains in silt sizes at the surface in Barbados. As discussed in section 5, the transport mode of African dust has recently been estimated as about $\tilde{D}_v = 3.5 \mu\text{m}$ or larger [*Reid et al.*, 2003].

[29] We employ the long-range transport mode observed by *Shettle* [1984] ($\tilde{D}_v = 2.524 \mu\text{m}$), as modified by *Schulz et al.* [1998] ($\sigma_g = 2.0$), for the subbin distribution of all transport bins. Thus the modeled dust transport is the most realistic for the long-range transport mode, exactly where it

is most important. Table 2 lists the size grid, subbin distribution parameters, and bin-mean physical properties for the $J = 4$ transport bins used in the model. Shown are minimum size D_{\min} , maximum size D_{\max} , volume median \tilde{D}_v and geometric standard deviation σ_g of subbin distribution, specific (i.e., per unit mass) number \underline{N} , surface area \underline{S} , scattering ψ_s , extinction ψ_e , and mass fraction of entrained and transported dust M_j , in each bin.

[30] For large J the subbin distribution is not important, since physical parameters do not vary greatly across the bin width. Subbin distribution is important when J is chosen to be small for computational efficiency ($J = 4$ in the present study). The analytic subbin distribution allows accurate prescription of physical properties known to vary significantly across the bin width. For example, the visible mass extinction coefficient ψ_e varies between $125\text{--}3600 \text{ m}^2 \text{ kg}^{-1}$ for dust particle sizes $0.1 < D_p < 1.0 \mu\text{m}$. The value used for the entire bin is $\psi_e = 2893 \text{ m}^2 \text{ kg}^{-1}$, corresponding to extinction by monodisperse particles of size $D = 0.37$ or $0.84 \mu\text{m}$.

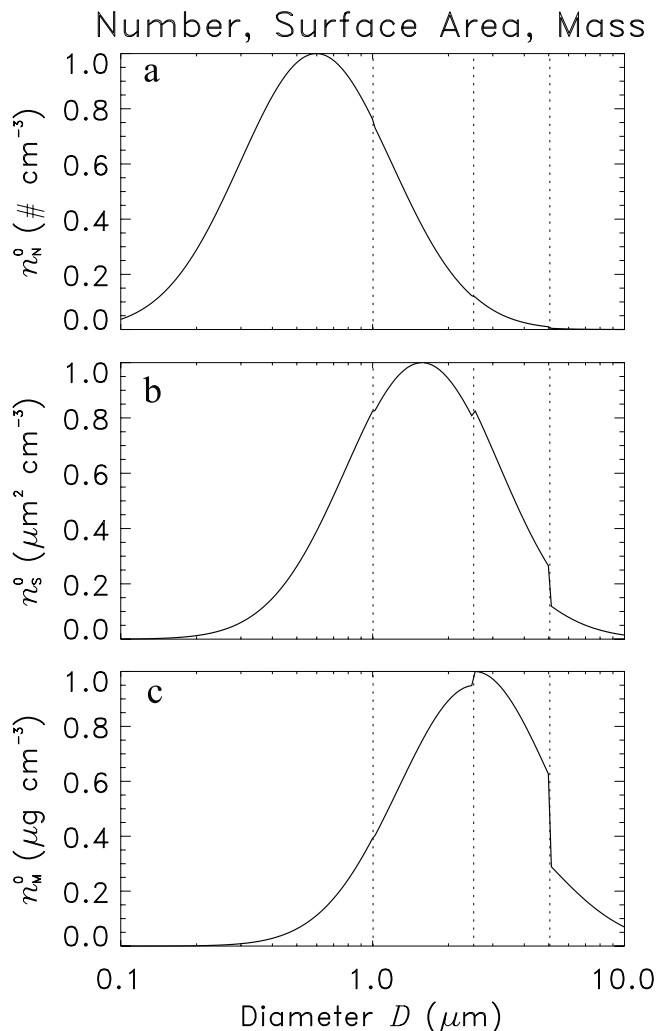


Figure 1. Predicted normalized size distribution at Barbados in July. Transport bins are separated by dashed vertical lines. Solid curve shows prescribed (a) subbin number, (b) surface area, and (c) mass distribution.

Table 2. Transport Bins and Subbin Distribution Parameters^a

Bin	D_{\min} , μm	D_{\max} , μm	\tilde{D}_v , μm	σ_g	N_i , # kg^{-1}	S_i , $\text{m}^2 \text{kg}^{-1}$	ψ_s , $\text{m}^2 \text{kg}^{-1}$	ψ_e , $\text{m}^2 \text{kg}^{-1}$	M_i , %
1	0.1	1.0	2.524	2.0	$3.484 + 15$	$3.464 + 03$	$2.834 + 03$	$2.893 + 03$	3.2
2	1.0	2.5	2.524	2.0	$2.138 + 14$	$1.471 + 03$	$7.779 + 02$	$8.350 + 02$	17
3	2.5	5.0	2.524	2.0	$2.205 + 13$	$7.107 + 02$	$3.343 + 02$	$3.825 + 02$	41
4	5.0	10.0	2.524	2.0	$3.165 + 12$	$3.741 + 02$	$1.705 + 02$	$1.961 + 02$	38

[31] All time-independent size-varying properties (extinction, scavenging cross-sections, sedimentation velocities) are computed on a high-resolution size grid, then weighted by the appropriate subbin distribution (surface area for extinction, volume for sedimentation), and then integrated to bin-mean values. The bin-mean value shown in Table 2 is used in all successive time steps. The bin boundaries at 2.5 and 10.0 μm facilitate comparison with U.S. PM2.5 and PM10 standards.

2.5. Optical Properties

[32] Optical properties of mineral dust are required to compare modeled dust distributions with remote sensing data, and to predict radiative feedbacks from dust in coupled models. There is great uncertainty in mineral dust optical properties for many reasons [Sokolik *et al.*, 2001]. For instance, long-range-transported mineral dust is generally not a pure crystal (e.g., quartz), but rather an internal mixture of many minerals [e.g., Pye, 1987] which is regionally dependent. Moreover, transported dust undergoes chemical processing by gases and other species leading to coated, multicomponent aerosols in the long-range mode [Dentener *et al.*, 1996]. As a result, treating dust measured in different regions with different techniques leads to large uncertainties in radiative forcing [Sokolik and Toon, 1996]. Fortunately, the total extinction coefficient of dust is much less uncertain than the individual scattering and absorbing components of dust, especially at visible wavelengths. In this study we use the visible indices of refraction for dust measured by Patterson [1981], $n = n_r + in_i = 1.56 + 0.0038i$ at 0.63 μm . This is very close to $n = 1.56 + 0.0033i$ at 0.625 μm measured in Afghanistan [Sokolik *et al.*, 1993]. It is, however, generally more refractive and absorptive than annual mean values at 0.67 μm inferred from AERONET measurements [Holben *et al.*, 1998] at dusty stations (Bahrain, Saudi Arabia, Cape Verde, Banizoumbou, Ougadougou, Barbados, and Mongolia) where $1.43 < n_r < 1.56$ and $0.001 < n_i < 0.0027$ (O. Dubovik, personal communication, 2001). The AERONET values appear to be influenced by aged dust with a significant water or organic component, and do not necessarily represent dry dust itself, the focus of this study. The specific scattering and extinction coefficients that result for each size bin of dust are shown in Table 2.

[33] The adequacy of these optical properties for the present purposes is confirmed by comparison to in situ estimates of the scattering properties of long-range-transported African dust. Li *et al.* [1996] measured the specific scattering efficiency of ashed Saharan dust at Barbados to be $\psi_s = 770 \text{ m}^2 \text{kg}^{-1}$ at 530 nm. Since dust is about 130% of ash by weight, this is equivalent to $\psi_s \sim 590 \text{ m}^2 \text{kg}^{-1}$ for dust. Maring *et al.* [2000] estimate $\psi_s = 610 \pm 100 \text{ m}^2 \text{kg}^{-1}$ at 550 nm for dust in Barbados. For comparison, the July mean

dust distribution shown in Figure 1 has $\psi_s = 720 \text{ m}^2 \text{kg}^{-1}$ at 550 nm. The modeled scattering efficiency of long-range Saharan dust is thus about 20% too high, but close to the measurement uncertainty. The inefficient long-range transport of large dust particles, previously mentioned, contributes to this bias since the visible specific scattering efficiency decreases with increasing particle size for $D > 1 \mu\text{m}$.

2.6. Land Surface and Geographic Constraints

[34] Many significant dust plumes emanate from completely barren regions, e.g., the Sahara desert or Saudi Arabian peninsula. There are, however, significant sources of dust in semi-arid regions where vegetation may act as a primary dust constraint, e.g., the Sahel [Teegen and Fung, 1995; Mbourou *et al.*, 1997]. Moreover, significant amounts of bare ground exist in temperate regions where dust emissions are minimal. It is therefore important to account for emission constraints by vegetation and by snow. The fraction of bare soil exposed in a grid cell A_m and thus suitable for mobilization is the maximally overlapped product of the fractions of dry ground, ground not covered with snow, and nonvegetated ground

$$A_m = (1 - A_l - A_w)(1 - A_s)(1 - A_V) \quad (13)$$

where A_l and A_w are the fractions of land covered by lakes and wetlands, respectively [Cogley, 1991; Bonan, 1996], A_s is the snow covered fraction of ground, and A_V is the fraction of ground covered by vegetation.

[35] Vegetation acts to constrain mineral dust emissions in multiple ways, which is why vegetation is a primary strategy to reduce soil erosion [e.g., Nicholson *et al.*, 1998]. First, standing vegetation and litter compete with bare ground as sinks for atmospheric momentum. This drag partitioning results in less drag on the erodible component of the surface [Raupach, 1994], because the canopy acts as a windbreak for the surface. Second, plant shade and root systems are effective at trapping soil moisture [e.g., Hillel, 1982]. This plant-induced soil moisture constraint is implicitly accounted for by equation (6). We assume that vegetation acts to constrain dust by linearly reducing the fraction of bare soil A exposed in a grid cell.

$$A_V = \min[1.0, \min(V, V_t)/V_t] \quad (14)$$

where V is the vegetation area index, the sum of the (one-sided) leaf plus stem area index. Thus $A_V = 0$ for $V = 0 \text{ m}^2 \text{m}^{-2}$ and $A_V = 1$ for $V \geq V_t$. The threshold for complete suppression of dust emissions is set to $V_t = 0.3 \text{ m}^2 \text{m}^{-2}$, a reasonable value based on previous studies [Mahowald *et al.*, 1999]. We use a global vegetation data set which supplies monthly leaf area index derived from 1 km satellite data [Kergoat *et al.*, 1999; Bonan *et al.*, 2002].

[36] A similar constraint is imposed for the horizontal fraction of ground covered by snow, A_s , and thus unavailable for dust entrainment. Snow coverage is derived from the liquid water equivalent snow depth h_l provided by the analyses (or model). We assume the bulk density of snow is $\rho_s = 100 \text{ kg m}^{-3}$, a reasonable value for fresh, nonpacked snow [Wiscombe and Warren, 1980], which is most relevant to dust emissions constraints.

$$h = h_l \rho_l / \rho_s \quad (15)$$

$$A_s = \min(h/h_t, 1.0) \quad (16)$$

where h is the geometric snow thickness and $h_t = 0.05 \text{ m}$ is the thickness for 100% snow coverage [Bonan, 1996] and dust suppression (equation (13)).

[37] Satellite-based retrievals of proxies for absorbing aerosol over land have revolutionized our understanding of mineral dust emission constraints. As suggested by J. M. Prospero, the TOMS absorbing aerosol index [Herman et al., 1997] provides strong evidence that climatologically strong dust emission regions, so-called “hot spots”, coincide with topographic depressions where alluvial sediments have accumulated [Gillette, 1999; Prospero et al., 2002]. Ginoux et al. [2001] parameterized this hypothesis in terms of a source erodibility factor S , defined as the fifth power of the ratio of the local height above a regional minimum height to the total elevation range of the surrounding $10^\circ \times 10^\circ$ region. This definition of S does not replicate the hydrological definition of an alluvial basin because it is based on a discontinuous function of elevation rather than on hydrological channels for surface runoff.

[38] Our time-independent erodibility factor is also based on the alluvial sediment/disturbance hypothesis, but represents the hydrologic processes involved more realistically. We quantify S as the upstream area from which sediment transported in surface runoff may have accumulated. We computed this upstream area from a Digital Elevation Map (NGDC TerrainBase 5 minute resolution) using the two step procedure described by Jenson and Domingue [1988]. Since estimates of surface runoff are explicitly not included in S , we call this S a geomorphic, rather than topographic or hydrologic, erodibility factor. A complete description and intercomparison of the geomorphic and other erodibility factors is presented by Zender et al. [2003]. Use of a spatially heterogeneous S , whether ours or that of Ginoux et al. [2001], dramatically improves the spatial correlation of simulated emissions with TOMS satellite indices of absorbing aerosols relative to simulations with no erodibility factor [Zender et al., 2003].

[39] Without correction DEAD predicts some small, geographically disparate sources where all the criteria are met (equation (17)) but where satellite data do not show climatologically strong dust burdens [Prospero et al., 2002] and surface data are lacking. These questionable “hot spots” are concentrated along or near the coasts of Greenland, Hudson’s Bay, Northern Siberia, and Northern Europe and are highly dependent on the land cover (vegetation) data set employed. These polar sources would not contribute much to dust loading, as they are generally washed out very quickly, but they can contribute up to 5–10% of surface fluxes. Rather

than allowing these dubious sources to influence our global simulations, we removed them by zeroing the erodibility factor S along coasts and north of 60°N . Ruling out apparently spurious sources in a model using ad hoc criteria (e.g., location or mean relative humidity), does conceal weaknesses in the model physics and/or boundary data sets. These questionable sources may be actual dust sources that do not show up in TOMS imagery because of weak strength, low absorption index, or thick boundary layers [Herman et al., 1997; Torres et al., 1998; N. M. Mahowald et al., Sensitivity of the TOMS aerosol index to boundary layer height: Implications for detection of mineral aerosol sources, submitted to *Geophysical Research Letters*, 2002]. They may also be dormant or potential sources which are close to activation or are anthropogenically constrained in the present climate.

2.7. Dust Entrainment

[40] Summarizing the physical processes described in the preceding sections, the total vertical mass flux of dust $F_{d,j}$ into transport bin j is

$$F_{d,j} = T A_m S \alpha Q_s \sum_{i=1}^I M_{i,j} \quad (17)$$

where the summation is over the $I = 3$ source modes. A global tuning factor, T , is chosen to give a reasonable climatological simulation. The global mean $F_{d,j}$ depends on the horizontal and temporal resolution of the model primarily because of the nonlinearity of Q_s with wind speed. These simulations set $T = 7.0 \times 10^{-4}$, resulting in a global annual entrainment of 1500 Tg yr^{-1} of fine dust ($D < 10 \mu\text{m}$) into the atmosphere.

2.8. Dry Deposition

[41] The simulation of dry deposition of mineral aerosol particles requires accurate specification of the gravitational settling and turbulent mix-out of particles ranging from $0.1 < D_p < 10.0 \mu\text{m}$.

2.8.1. Gravitational Settling

[42] Particles are assumed to settle gravitationally at their terminal velocity v_g and to reach this velocity instantaneously.

$$v_g = \left(\frac{4gDC_c\rho_p}{3C_D\rho} \right)^{1/2} \quad (18)$$

where the slip correction factor C_c and the drag coefficient C_D are given by Seinfeld and Pandis [1997]. In general C_D is a function of v_g so equation (18) is an implicit equation for v_g . An iterative solution to equation (18) is straightforward but too time consuming for large-scale atmospheric models. The following Stokes approximation provides a solution.

[43] For Reynolds numbers $\text{Re} < 0.1$ particles obey the Stokes settling velocity u_{St}

$$u_{\text{St}} = \frac{D^2 \rho_p g C_c}{18\mu} \quad (19)$$

where μ is the dynamic viscosity of air. The present study is limited to the size range 0.1 to $10.0 \mu\text{m}$, where the Stokes

velocity u_{St} is an excellent approximation to the terminal velocity v_g . However, the Stokes velocity overestimates sedimentation by more than 10% for $D > 20 \mu\text{m}$. The model is capable of simulating saltation and dust production of particle sizes up to about $1000 \mu\text{m}$. For large particle sizes, we define a time-invariant Stokes correction factor C_{St} as

$$C_{St} = v_g/u_{St} \quad (20)$$

The C_{St} correction exceeds 10% for mineral particles larger than $45 \mu\text{m}$. At startup, the model computes $C_{St}(D, p_0, T_0)$ (equation (20)) for each size at a temperature and pressure representative of arid erosion regions on Earth (currently $p_0 = 1000 \text{ mb}$ and $T_0 = 295 \text{ K}$). The model computes u_{St} (equation (19)) every time step and applies the time-invariant correction factor $C_{St}(D, p_0, T_0)$ to obtain v_g . In this manner the iterative solution to (equation (18)) at every time step is avoided.

2.8.2. Turbulent Mix-Out

[44] Turbulent deposition is treated using the resistance-in-series method [e.g., *Wesely*, 1989]. The turbulent deposition velocity v_t

$$v_t = \frac{1}{r_a + r_b + r_a r_b v_g} \quad (21)$$

where the aerodynamic resistance through the constant flux layer r_a accounts for local stability effects using similarity theory as given by *Bonan* [1996]. While r_a is independent of particle size, the quasi-laminar layer resistance r_b depends on a particle's microphysical characteristics

$$r_b = \frac{1}{u_* (\text{Sc}^{-2/3} + 10^{-3}/\text{St})} \quad (22)$$

The Schmidt number Sc in the denominator accounts for Brownian diffusion and is dominant for $D \lesssim 0.7 \mu\text{m}$. The Stokes number St accounts for inertial impaction and becomes important for $D \gtrsim 5 \mu\text{m}$. The resistance to particle or gaseous diffusion across the quasi-laminar layer to a solid surface is proportional to $\text{Sc}^{-2/3}$ [*Slinn et al.*, 1978]. The rate-limiting resistance for turbulent mix-out of accumulation mode particles is r_b for all wind speeds. For coarser particles ($D \gtrsim 2 \mu\text{m}$), r_b is the limiting resistance for weak wind speeds, but r_a becomes the limiting resistance as U increases above about 5 m s^{-1} .

2.9. Wet Deposition

[45] Wet deposition or scavenging of particles by water is the dominant deposition process for small dust particles because dry deposition processes are very inefficient for $D < 2 \mu\text{m}$, i.e., the accumulation mode [e.g., *Seinfeld and Pandis*, 1997]. The model distinctly treats four types of wet deposition: Particles in precipitating clouds may be removed by nucleation scavenging, which accounts for particles becoming embedded in raindrops. Particles beneath precipitating clouds are susceptible to subcloud scavenging (collision scavenging), which accounts for collection of particles by precipitation. Nucleation and subcloud scavenging are each treated separately for convective and for stratiform precipitation.

[46] Evaporation of stratiform precipitation releases scavenged dust in direct proportion to the mixing ratio of dust in the precipitation and the local rain evaporation rate. The evaporation tendency is consistent with the hydrologic cycle of the CCM3 [*Hack et al.*, 1998]. In reality, scavenged dust remains scavenged within hydrometeors until complete evaporation occurs. Thus our evaporative tendency is expected to overestimate the source of dust at lower atmospheric layers in cases where not all precipitation evaporates, i.e., virga. Evaporation is a significant source of dust in some regions (see Figure 4c).

[47] Cloud aerosol interaction and scavenging are treated using the method of *Rasch et al.* [2000] with modifications. Wet scavenging processes are treated with mass mean, path-normalized scavenging coefficients $\bar{\Lambda}_M \text{ m}^2 \text{ kg}^{-1}$ defined such that

$$\frac{dM_p}{dt} = AP_M \bar{\Lambda}_M M_p \quad (23)$$

where $M_p \text{ kg m}^{-2}$ is the grid cell aerosol mass path, A is the horizontal fraction of the grid cell over which the scavenging process occurs, and $P_M \text{ kg m}^{-2} \text{ s}^{-1}$ is the rate of interaction with the scavenging droplets. Nucleation scavenging occurs through the entire vertical domain of the fraction of the grid cell occupied by the appropriate (convective or stratiform) cloud type. For nucleation scavenging, P_M is the local rate of conversion of water vapor to droplets (condensation) and $\bar{\Lambda}_M = 0.10$ for all size classes for both stratiform and convective rain. This is identical to the sulfate nucleation scavenging representation of *Rasch et al.* [2000], except that we allow dust nucleation scavenging in frozen clouds. These modifications to the sulfate scavenging procedure in MATCH are intended to account for the hygrophobic nature of mineral aerosol, as well as its potential role in forming ice nuclei [*Pruppacher and Klett*, 1978]. The present simple formulation of nucleation scavenging must be viewed as a temporary procedure while more microphysically based methods involving aerosol composition and CCN activation suitable for large-scale models are investigated [e.g., *Zhang et al.*, 2000].

[48] The treatment of subcloud scavenging partially accounts for effects due to vertical cloud structure and to aerosol size and hydrometeor size distribution. For subcloud scavenging, A in equation (23) is the maximally overlapped fraction of the product of cloud fraction and precipitation rate of all grid points above the given grid point, while P_M is the appropriate (convective or stratiform) precipitation rate [*Rasch et al.*, 2000]. The model accounts for the strong dependence of subcloud scavenging on both the aerosol and precipitation size distributions which arises from the interaction of the constituent processes of Brownian diffusion, interception, and impaction [e.g., *Dana and Hales*, 1976]. The contributions of diffusion, interception, and impaction to the precipitation collection efficiency is modeled following *Seinfeld and Pandis* [1997]. The subbin distribution of particles within each transport bin is as described in section 2.1. As mentioned above, the transport model separately diagnoses stratiform and convective precipitation. For the purposes of scavenging, stratiform and convective precipitation droplets are assumed to obey lognormal size distri-

Table 3. Subcloud Scavenging Efficiencies

Diameter, μm	Convective		Stratiform	
	Theory, ^a $\text{m}^2 \text{kg}^{-1}$	DEAD, ^b $\text{m}^2 \text{kg}^{-1}$	Theory, ^a $\text{m}^2 \text{kg}^{-1}$	DEAD, ^b $\text{m}^2 \text{kg}^{-1}$
0.1–1.0	7.61–5	2.00–2	2.47–4	3.00–2
1.0–2.5	4.10–3	5.00–2	8.53–3	1.00–1
2.5–5.0	1.05–1	1.05–1	1.97–1	1.97–1
5.0–10.0	2.68–1	2.68–1	4.78–1	4.78–1

^aTheoretical values computed as in the work of *Seinfeld and Pandis* [1997].

^bValue used in DEAD.

butions with number median diameters of 400 and 1000 μm , respectively.

[49] Table 3 shows the wet deposition scavenging efficiencies employed in the model. For each type of precipitation (convective and stratiform are diagnosed separately by MATCH) two below-cloud scavenging efficiencies are shown for each transport bin. The first is a theoretical mass mean scavenging coefficient $\bar{\Lambda}_M$ and the second is the value DEAD uses.

[50] The theoretical $\bar{\Lambda}_M$ values are obtained using the method of *Dana and Hales* [1976]. The raindrop-aerosol collision efficiency between droplets of size D_p and dust particles of size D_p is set to

$$E(D_p, D_p) = E_{BD} + E_{NTC} + E_{MPC} \quad (24)$$

where E_{BD} , E_{NTC} , and E_{MPC} are the individual collision efficiencies for the processes of Brownian diffusion, interception, and inertial impaction. The meanings and analytic form of each term are given by *Seinfeld and Pandis* [1997]. The collision efficiencies are integrated over the raindrop size distribution and the subbin dust distribution. For these calculations, convective and stratiform rain droplets are assumed to adhere to globally uniform lognormal size distributions with number median diameters of 1000 and 400 μm , respectively, and geometric standard deviations of $\sigma_g = 1.86$, as given by *Dana and Hales* [1976]. The overbar in the $\bar{\Lambda}_M$ symbol indicates that all of the tabulated efficiencies have been normalized by the precipitation intensity $P_M \text{ kg m}^{-2} \text{ s}^{-1}$. This procedure results $\bar{\Lambda}_M < 0.01$ for clay-sized dust particles. These small $\bar{\Lambda}_M$ may be contrasted with empirically determined values of $\bar{\Lambda}_M$ closer to 0.1 employed in some global models [*Balkanski et al.*, 1993; *Rasch et al.*, 2000].

[51] The $\bar{\Lambda}_M$ used by DEAD for $D > 2.5$ were picked to provide more realistic simulations of long-range-transported dust. These $\bar{\Lambda}_M$ are larger than the theoretical values obtained from equation (24) by a factor of about 200 (for bin 1) and 10 (for bin 2). The extremely small subcloud scavenging efficiency of accumulation mode aerosol (see discussion given by *Seinfeld and Pandis* [1997]), when applied without modification to dry-sized dust particles in global models, appears to lead to unrealistically long lifetimes. It is possible that dust particles in humid regions are swelled internally (e.g., smectites) [*Pruppacher and Klett*, 1978], coated with water [*Hänel*, 1976], coagulated with other aerosols, or aspherical in shape so that their effective size during transport is larger than accounted for by the model. These mechanisms would increase particle interception

and impaction efficiencies in equation (24) and hence help to reconcile the discrepancy between theoretical and ad hoc values of subcloud scavenging efficiencies for $D < 2.5 \mu\text{m}$ in Table 3. However, all these mechanisms except the last also reduce the particle's density which tends to decrease interception and impaction.

3. Global Simulations

[52] To evaluate the model predictions against observations, we have simulated the dust distribution for the ten year period 1990–1999 using NCEP/NCAR reanalyses [*Kalnay*, 1996] to drive the DEAD module in the MATCH chemical transport model [*Rasch et al.*, 1997]. The results shown below were obtained from a MATCH run from 1 December 1989 to 31 December 1999, using 6-hourly NCEP meteorology interpolated to the nearest 30 min model time step. Results were archived as daily mean values, then aggregated into monthly, annual, and climatological means. The first month was discarded to remove spin-up effects. We present only the gross climatological simulation of DEAD. We omit discussion of the interannual variability in all but one figure, pending longer model simulations. More extensive comparisons between DEAD and observed dust climatology are shown by Luo et al. (submitted manuscript, 2003) and *Mahowald et al.* [2003].

[53] Figure 2 shows the predicted annual mean mobilization and deposition tendencies. In general, entrainment occurs in regions with a combination of high winds, low vegetation, no snow cover, and large upstream area. The strongest sources are in North Africa and Asia, consistent with satellite observations and previous studies [*Herman et al.*, 1997; *Husar et al.*, 1997; *Ginoux et al.*, 2001; *Prospero et al.*, 2002]. These sources include the Bodele depression, Mali and Southern Algeria, the Takli Makan, and Gobi deserts, Patagonia, and the Lake Eyre basin in Australia. North American sources center on Texas and the Great Plains regions, with additional sources in the Sonoran Desert of Mexico and the southwestern United States.

[54] The predicted dust deposition in Figure 2b shows that mass transport affects extensive areas downwind of the source regions. Deposition to the tropical North Atlantic from the North African plume is especially significant, averaging $750 \text{ mg cm}^{-2} \text{ ka}^{-1}$ in the ocean region bounded by $0\text{--}30^\circ\text{N}$, $0\text{--}60^\circ\text{W}$. We estimate an annual mean aeolian deposition flux at the mouth of the Amazon (0°N , 50°W) of $75 \text{ kg ha}^{-1} \text{ yr}^{-1}$. This is consistent with the estimate by *Swap et al.* [1992] of less than $190 \text{ kg ha}^{-1} \text{ yr}^{-1}$.

[55] Figures 2c and 2d decompose the total deposition into dry and wet deposition fluxes. As expected, dry deposition dominates source regions because of large particle sedimentation. Dry deposition operates continuously, so the dry deposition patterns are an excellent proxy for atmospheric dust loading (Figure 3). Wet deposition removes most dust far from source regions. It accounts for 41% of global deposition, and for 89% of oceanic deposition.

[56] The important role of wet deposition in our study is consistent with some previous modeling studies [*Tegen and Fung*, 1994; *Woodward*, 2001] but inconsistent with *Ginoux et al.* [2001], who find that wet deposition accounts for only 13% of global deposition. These models represent wet

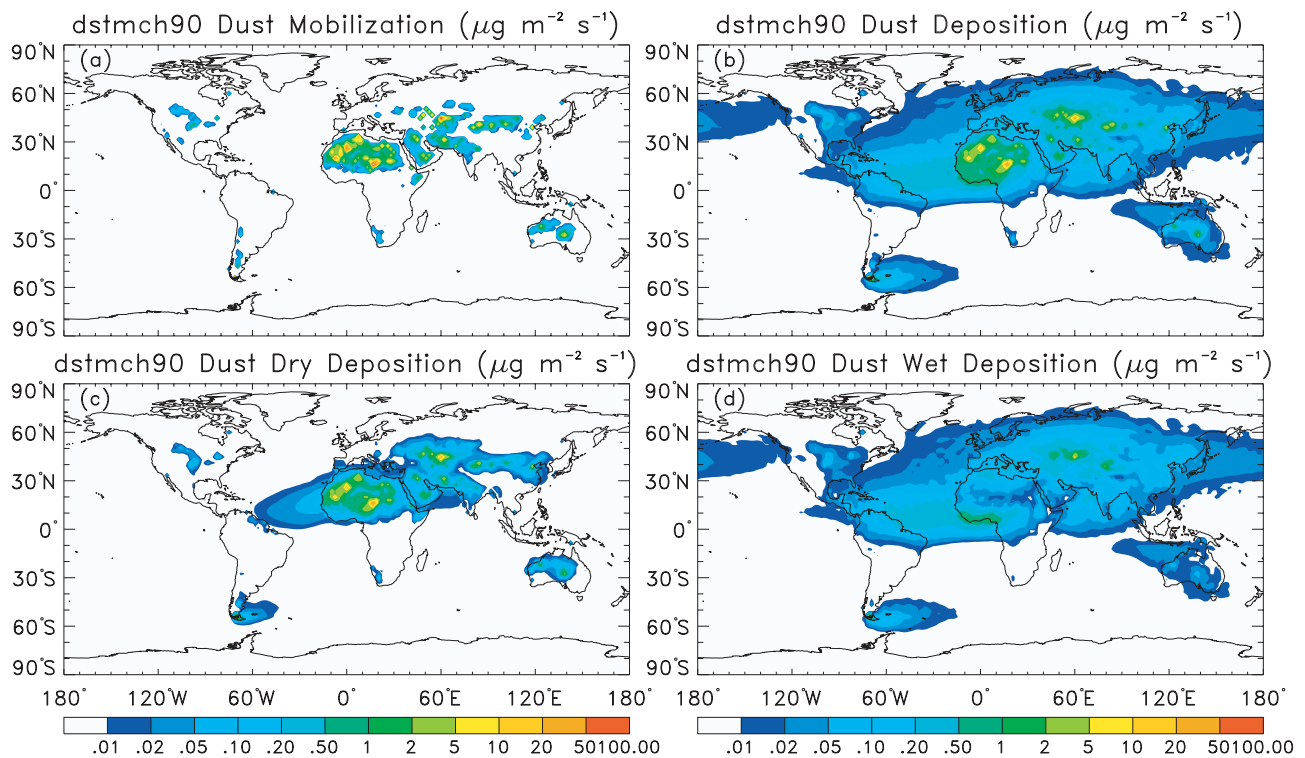


Figure 2. Predicted annual mean dust source and sink fluxes in $\mu\text{g m}^{-2} \text{s}^{-1}$ for (a) mobilization, (b) total deposition, (c) dry deposition, (d) wet deposition. Scale is logarithmic.

scavenging differently and use different effective scavenging coefficients. We (Table 3) and Woodward [2001] use size-dependent washout coefficients, while Tegen and Fung [1994], Ginoux *et al.* [2001], and Luo *et al.* (submitted manuscript, 2003) use uniform scavenging coefficients which vary from 250 to 750. Observations of the ratio of dry to wet deposition are difficult [Duce *et al.*, 1991] and are available at too few locations [e.g., Arimoto *et al.*, 1985; Guieu *et al.*, 1997] to provide robust global constraints. However, spatiotemporal variability in deposition processes would not reconcile the above modeling differences. This emphasizes the need for coordinated, long term measurements of the spatial and temporal distribution of wet and dry particle deposition.

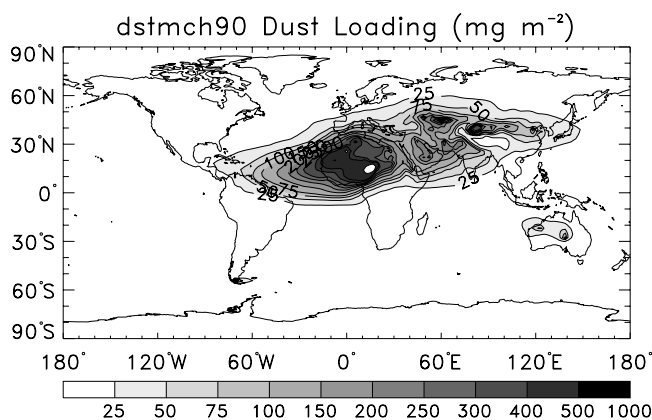


Figure 3. Predicted annual mean dust mass burden in mg m^{-2} . Scale is nonlinear.

[57] In descending order, the annual mean emissions from the relevant continents in Tg yr^{-1} are Africa, 980; Asia, 415; Australia, 37; South America, 35; North America, 8. Table 4 compares the predicted annual mean deposition to ocean regions to previous observation-based [Duce *et al.*, 1991; Prospero, 1996b] and model [Ginoux *et al.*, 2001] estimates.

[58] DEAD and GOCART agree on deposition to the Atlantic and South Indian Oceans, but strongly disagree in the Pacific and North Indian Oceans. The lower estimates of DEAD are much closer than GOCART to Prospero [1996b] in the Indian and South Pacific Oceans, but much farther from Prospero [1996b] in the North Pacific. Predicted and observed ocean sediment fluxes from a slightly different version of DEAD, as well as source/sink mappings, are compared by Luo *et al.* (submitted manuscript, 2003).

[59] Table 5 compares the annual mean deposition of trace metals to each ocean. These estimates assume that trace metals constitute a fraction of the dust mass equal to

Table 4. Oceanic Deposition^a

Region	Duce <i>et al.</i> [1991]	Prospero [1996b]	GOCART	DEAD
N. Pacific	480	96	92	31
S. Pacific	39	8	28	8
N. Atlantic	220	220	184	178
S. Atlantic	24	5	20	29
N. Indian	100	20	138	36
S. Indian	44	9	16	12
Global	910	358	478	314

^aUnits are Tg yr^{-1} . Sources are Duce *et al.* [1991], Prospero [1996b], and Ginoux *et al.* [2001].

Table 5. Trace Metal Deposition to Oceans^a

Region	Al	Fe	Si	P
N. Pacific	2.5	1.1	9.7	0.03
S. Pacific	0.7	0.3	2.6	0.01
N. Atlantic	14.3	6.2	55	0.19
S. Atlantic	2.3	1.0	8.9	0.03
N. Indian	2.9	1.3	11.1	0.04
S. Indian	0.9	0.4	3.6	0.01
Global	25.3	11.0	96.8	0.33

^aUnits are Tg yr⁻¹. Assumes dust is 8.01% Al, 3.5% Fe, 30.8% Si, and 1050 ppm P as given by *Duce et al.* [1991].

their estimated abundance in the upper continental crust, following *Duce et al.* [1991].

[60] Figure 3 shows the predicted annual mean mass burden in mg m⁻². Strong dust burdens are apparent over source regions, where large particles have not yet sedimented. Source regions in North and South America and South Africa show no persistent dust burden, contrary to what might be expected on the basis of Figure 2a. Emissions in these midlatitude source regions are very susceptible to wet scavenging, and so have a relatively short lifetime.

[61] Table 6 compares the climatological (1990–1999) mean budget of the DEAD simulation against previous studies and estimates. DEAD predicts emissions of about 1490 Tg yr⁻¹ for particles smaller than 10 μm. This is about 84% of the GOCART estimate of 1782 Tg yr⁻¹ for $D < 12$ μm [*Ginoux et al.*, 2001]. The reduced emissions predicted by DEAD relative to GOCART are consistent with the relative dust concentrations and deposition fluxes (and their biases) predicted by the models. DEAD is consistent with the IPCC estimate of the range of $D < 20$ μm dust emissions in the year 2000 of 1000–3000 Tg yr⁻¹ with a best guess value of 2150 Tg yr⁻¹ [*Penner et al.*, 2001].

[62] Because of the relatively short length of our base simulation (1990–1999) we have not until now discussed the modeled interannual dust variability. To illustrate this we computed the modeled mean \bar{x} and interannual variability for dust emissions, burden and optical depth from the 10 total years (1990–1999). We quantify the interannual variability using the standard deviation σ_x of the global annual mean time series of each field. The resulting estimates of $\bar{x} \pm \sigma_x$ for the 1990s are as follows: emissions, 1490 ± 160 Tg yr⁻¹; burden, 17 ± 2 Tg; and optical depth, 0.030 ± 0.004 . The normalized variability (σ_x/\bar{x}) is 10–15% for all three fields. The normalized variability in emission is slightly less than that for burden and optical depth because emission is sensitive to all particle sizes (see Table 8), whereas burden and optical depth are not sensitive to the largest particle sizes. Interestingly, the 1996–1998 means were all about 20% less dusty than 1990–1995.

[63] Figure 4 shows the predicted annual mean mixing ratio and source and sink tendencies. Figure 4a shows maxima in the subtropical belts which contain the Earth's major desert sources. Dust from North Africa and the Saudi Arabian peninsula dominates the northern hemisphere zonal mean. The broad half-width of the maxima is due to the inclusion of central and East Asian dust sources which are 5–10° north of the African sources. Australia and Patagonian sources create the peak in the southern hemisphere. Figure 4b shows that northern hemispheric wet deposition

occurs in two distinct meridional locations: scavenging over the Atlantic and Indian Oceans accounts for the subtropical maxima, while scavenging over the Pacific determines the extratropical maxima.

[64] As mentioned in section 2.9, stratiform rain continually releases scavenged dust in proportion to its local evaporation rate. Figure 4c shows that gravitational settling is about 5–10 times less efficient than wet processes in removing dust from the free atmosphere in the subtropics. Settling is a small source of dust at low levels near the equator and in the extratropics and in the polar regions. Turbulent deposition operates only in the lowest model layer and thus is difficult to portray in vertical cross-section (see Figure 2c for turbulent deposition). Figure 4d shows that evaporation of stratiform rain vertically redistributes a significant fraction of the scavenged dust. Evaporation is comparable to wet scavenging over the tropical north Atlantic. A more complex but physically consistent model representation of evaporation would release scavenged dust discretely as each rain droplet completely evaporated, i.e., in virga. Thus it is likely our model overestimates the importance of evaporation.

[65] Table 7 shows the size dependence of all deposition timescales in days. Dry deposition is the sum of gravitational and turbulent deposition. The turnover time for particles in the accumulation mode ($D < 2.5$ μm) is on the order of months when considering only dry processes. The relative importance of dry and wet deposition changes at 2.5 μm, near the boundary between clay and silt-sized particles. Wet scavenging accounts for 41% of global mean deposition and dry processes account for 59%. The size-dependent global annual mean fluxes are summarized in Table 8. Wet deposition is the only effective removal process for small particles in our model. This is consistent with some [*Teegen and Fung*, 1994], but not all [*Ginoux et al.*, 2001] previous studies.

[66] We compare the predicted Aerosol Optical Depth (AOD) at 0.63 μm against AOD inferred from AVHRR channel 1 observations [*Stowe et al.*, 1997] with a nominal band center of 0.63 μm. Table 2 lists the specific extinction coefficients used for each bin. Figure 5 shows the predicted seasonal mean dust optical depth at 0.63 μm. Figure 5 may be compared to Plate 1 of *Husar et al.* [1997]. Note the scale is nonlinear for $\tau > 0.5$. The seasonal migration of the North African dust plume with the ITCZ seen in AVHRR and TOMS observations [*Husar et al.*, 1997; *Herman et al.*, 1997], and in other models [*Teegen and Fung*, 1995; *Ginoux et al.*, 2001; Luo et al., submitted manuscript, 2003], is

Table 6. Climatological Budget Statistics^a

Quantity	<i>Andreae</i> [1996]	<i>Teegen and</i> <i>Fung</i> [1994]	GOCART ^b	DEAD
Emission	1500	1222	1814	1490
Lifetime	4	4	7.1	4.3
Burden	8.4	18.8	35.9	17.4
τ Ocean	–	–	–	0.020
τ Land	–	–	–	0.051
τ Global	0.023	0.033	–	0.030

^aEmission in Tg yr⁻¹, Lifetime in days, Burden in Tg, Optical depths τ at 0.63 μm. Sources: *Andreae* [1996], *Teegen and Fung* [1994], and *Ginoux et al.* [2001].

^bFor size range $0.2 \leq D \leq 12$ μm.

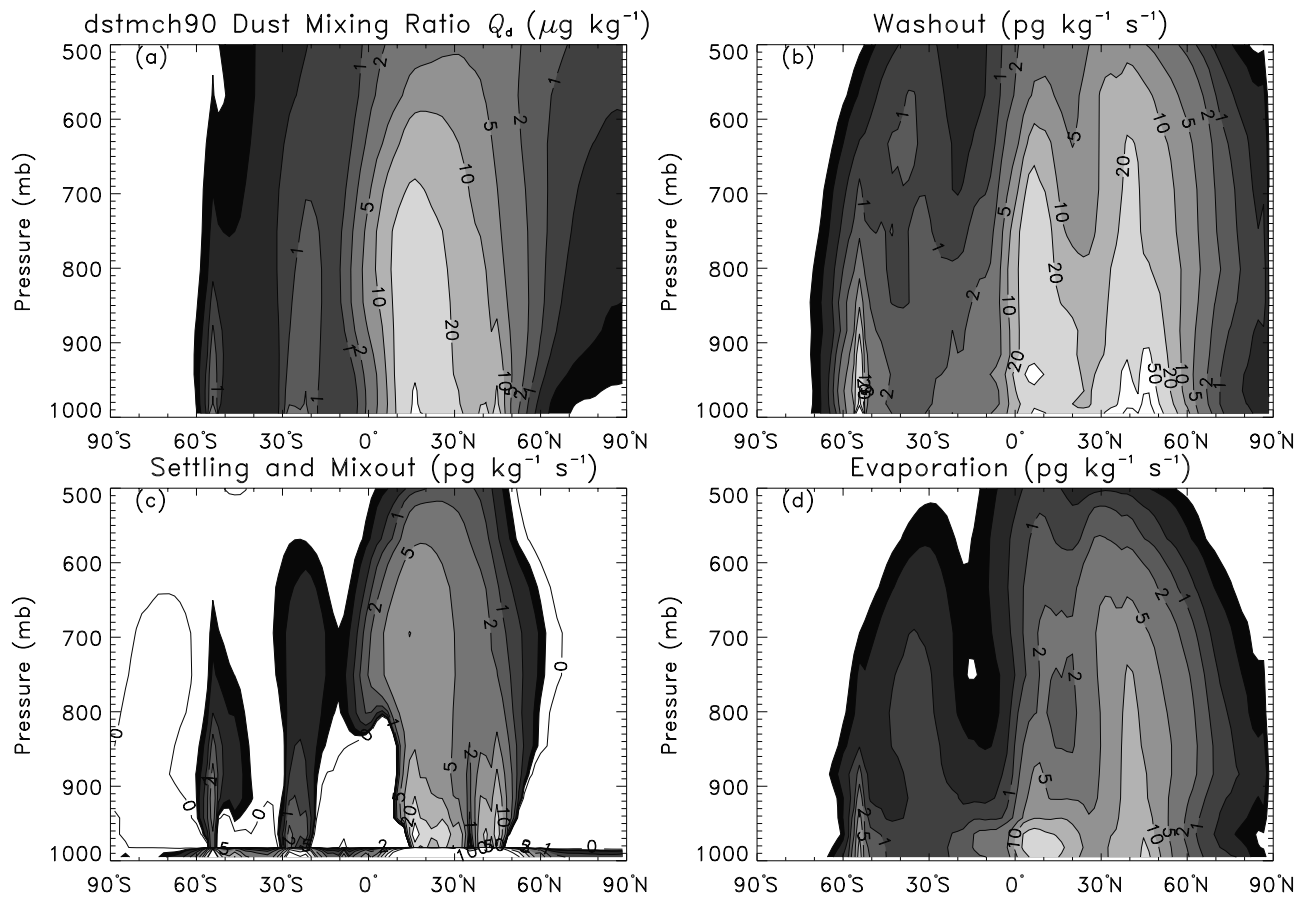


Figure 4. Predicted climatological mean dust mixing ratio and mixing ratio tendencies. Shown are (a) mixing ratio in $\mu\text{g kg}^{-1}$, and, in $\text{pg kg}^{-1} \text{s}^{-1}$, (b) wet deposition, (c) dry deposition, and (d) evaporation. Contour intervals are logarithmic.

clearly present. This dust plume is thought to dominate the AVHRR observed total aerosol plume in July, and to contribute about half of the observed optical depth in January, when it is complemented by carbonaceous aerosol from biomass burning [Chin *et al.*, 2002]. The dust plume reaches its farthest westward extent in July, when it is observed to be responsible for the majority of aerosol scattering as far north as Miami [Prospero, 1999]. The peak extent of the Asian-Pacific dust plume occurs in northern Spring. Vertical cross-sections of the seasonal dust distribution (not shown) reveal that East Asian dust may travel 200 mb higher in the atmosphere than African dust.

4. Station-Based Evaluations

[67] Surface observations at sites managed by the University of Miami provide a unique long-term record and

climatology of surface dust concentrations. The dust concentration data (which are provided by J. Prospero and D. Savoie) for many University of Miami stations are calculated from Al concentrations provided by R. Arimoto. Interannual and decadal trends in the African dust plume in the North Atlantic have been analyzed using these data [e.g., Prospero and Nees, 1986; Prospero, 1996a]. Previous global dust simulations have been evaluated against these University of Miami station observations [Tegen and Fung, 1994; Tegen *et al.*, 1996; Mahowald *et al.*, 1999; Ginoux *et al.*, 2001; Perlwitz *et al.*, 2001; Luo *et al.*, submitted manuscript, 2003; Penner *et al.*, 2001]. Station locations [Ginoux *et al.*, 2001; Woodward, 2001] and sampling techniques [Savoie *et al.*, 1992; Maring *et al.*, 2000] are discussed in the literature. Figure 6 compares predicted to observed monthly mean dust concentrations in $\mu\text{g m}^{-3}$. All stations are compared to 1990–1999 model climatology

Table 7. Global Mean Turnover Time by Removal Process^a

Diameter	Grav	Turb	Dry	Wet	Total
0.1–1.0 μm	1059	928	494	18	17
1.0–2.5 μm	171	198	92	12	11
2.5–5.0 μm	34	7	6	9	3.5
5.0–10.0 μm	7	1.6	1.3	6.1	1.1
0.1–10.0 μm	40	9.1	7.4	10.5	4.3

^aValues are given in days.

Table 8. Global Emissions and Deposition Fluxes

Diameter, μm	Source, Tg yr^{-1}	Dry, Tg yr^{-1}	Wet, Tg yr^{-1}	Load, Tg	Life, days
0.1–1.0	48	2	47	2.2	17
1.0–2.5	260	31	229	7.7	11
2.5–5.0	609	371	236	5.8	3.5
5.0–10.0	573	463	96	1.6	1.1
0.1–10.0	1490	866	607	17.4	4.3

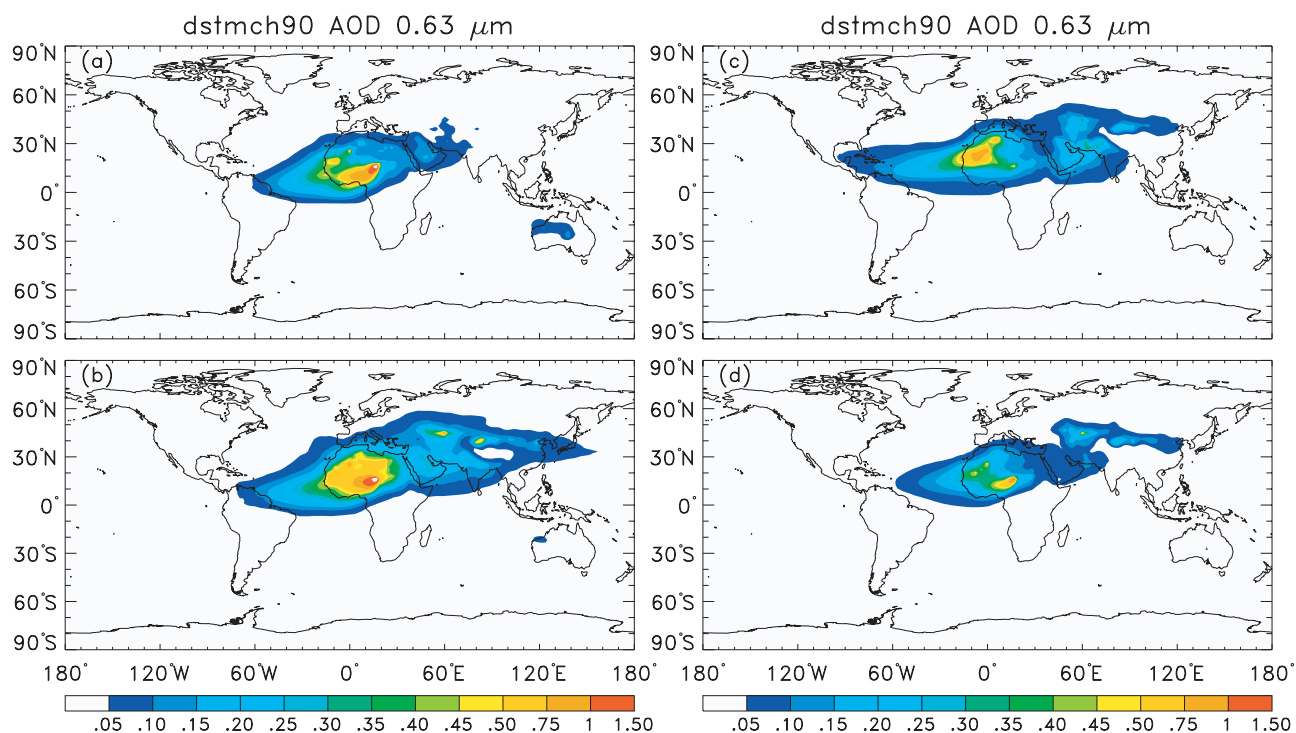


Figure 5. Predicted seasonal mean dust optical depth at 0.63 μm for (a) DJF, (b) MAM, (c) JJA, and (d) SON.

except Sal Island, Jeju (formerly known as Cheju), Okinawa, and Mace Head. Model data for these stations have been subsampled to include only months during which observations were recorded.

[68] The first six stations are sensitive to the North African dust plume. The modeled and observed surface dust seasonal cycle generally overlap at Barbados, Miami, Bermuda, Izaña, Sal Island and Mace Head. At Barbados, the observed seasonal cycle peaks in June but DEAD peaks in May and is too active in winter. In Miami and Bermuda the peak occurs in July and August in both observed and model results, but the modeled amplitude appears too weak in summer. In general the phasing and amplitude of the surface concentration is within the monthly variance at these stations. We are currently producing longer model climatologies to better quantify the model variance. The greatest absolute disparity with the North Atlantic measurements occurs in the winter at Sal Island, about 700 km from the African continent, when the model overpredicts dust concentration by a factor of 2. This disparity indicates a problem with source regions near the coast or the vertical location of the plume. Izaña, which lies only 350 km from Western Sahara coast, and is located at 2360 m in the free troposphere, is also overpredicted in amplitude, but only by 20%. At Izaña, however, the observational sampling is sectorized to the west, and so may miss some large dust events which could help to reconcile the model and measurements. The simulation at Mace Head has a low annual-mean concentration with strong variability in springtime as observed.

[69] Evaluating the dust distribution over the Indian Ocean is difficult because of the paucity of data there

[Prospero, 1996b]. The seasonal cycle at Kaashidhoo appears reasonable, but anthropogenic aerosol such as fly ash may account for a significant fraction of the aerosol reported as dust (D. Savoie, personal communication, 1999). Thus DEAD's underestimate of concentration at Kaashidhoo may be more representative of true desert dust in the central Indian Ocean region.

[70] Station observations in the North Pacific generally show a springtime concentration maximum [Prospero, 1996b]. DEAD captures this seasonal phasing at Jeju, Okinawa, Enewetak, and Midway. The observed and modeled climatology at Oahu is very similar to Midway and has been omitted. DEAD underestimates the mean concentrations closer to the Asian sources at Jeju and Okinawa (where complex terrain may have biased sampling) by about 40%, with most of the bias occurring during the winter and spring. On the other hand, DEAD overestimates concentrations farther from the continent at Enewetak, Midway and Oahu by up to 100%, most noticeably in spring. Excessive springtime transport of African dust contributes to this high bias.

[71] The South Pacific stations farthest from source regions show weak seasonal cycles and very low mean concentrations. The overlap between University of Miami observations and model predictions is generally acceptable at Nauru, American Samoa, and New Caledonia. Cape Grim, however, which is closest to a source region (Australia), shows a strong austral summer peak in the observations that is absent in the model. Closer examination (see, e.g., Figures 2a, 2b, and 5a) shows that DEAD captures the phasing of summertime Australian emissions, but that the circulation moves this dust more

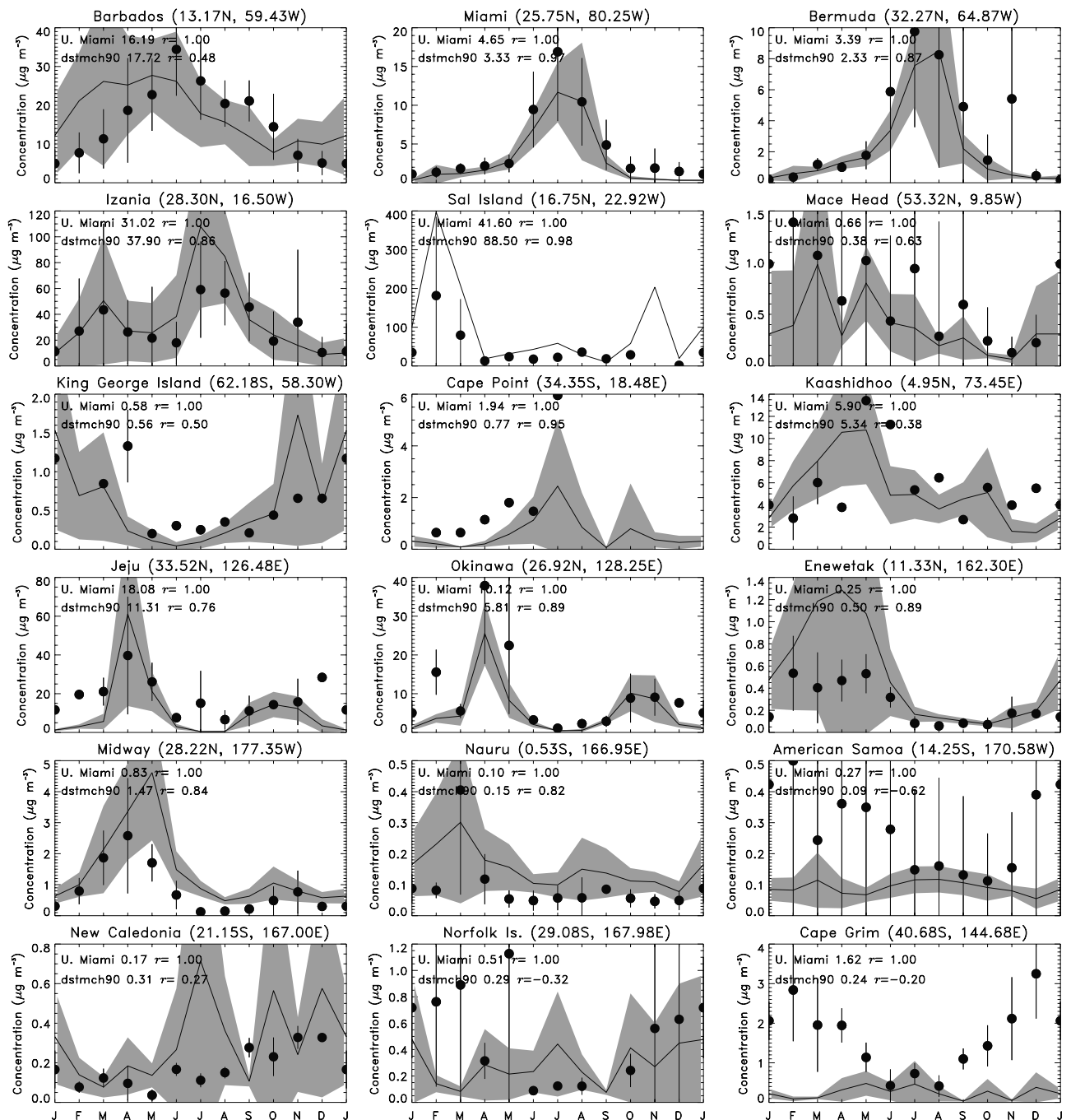


Figure 6. Predicted (line) and observed (circles) monthly mean surface dust concentration ($\mu\text{g m}^{-3}$) at University of Miami stations. Two standard deviations are shown for predicted (shading) and observed (whiskers) values. No whisker indicates data only available for same month in $N < 2$ years. Station and model names are followed by the climatological mean concentration ($\mu\text{g m}^{-3}$) and the linear correlation coefficient r with the observed monthly mean data.

efficiently toward the northwest than the southeast to Cape Grim.

[72] Intercomparison of model performance among all the University of Miami–operated stations shows the great dynamic range of dust concentration in the atmosphere. The climatological mean observed and modeled concentrations at each station in Figure 6 are given as a scatterplot in Figure 7. Both observed and modeled concentrations span 3

orders of magnitude. In general, stations downwind of the African plume are better simulated than Asian and North Pacific stations which are in turn better simulated than southern hemisphere stations. Disparities at stations with the highest and lowest mean dust concentrations are generally less than a factor of 2, and the model is never more than a factor of 2 too high. Dust concentration appears significantly underpredicted at four stations in the southern hemi-

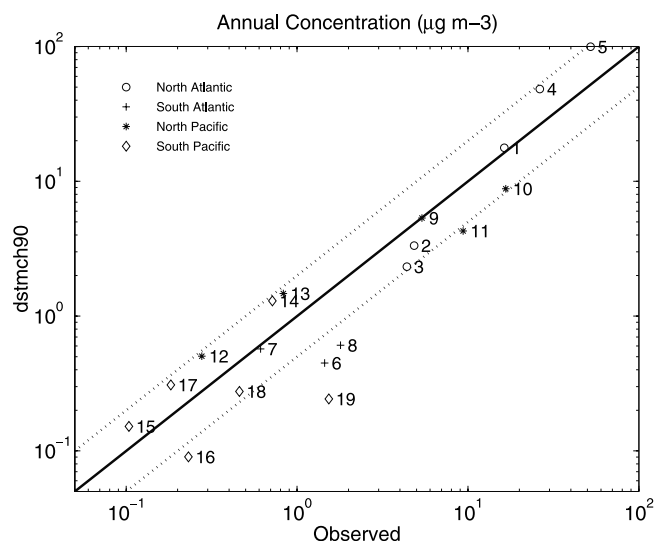


Figure 7. Predicted and observed climatological mean surface concentration of dust in $\mu\text{g m}^{-3}$ at University of Miami stations. Dashed lines indicate factor of 2 disparity. Site numbers are 1. Barbados, 2. Miami, 3. Bermuda, 4. Izaña, 5. Sal Island, 6. Mace Head, 7. King George Island, 8. Cape Point, 9. Kaashidhoo, 10. Jeju, 11. Okinawa, 12. Enewetak, 13. Midway, 14. Oahu, 15. Nauru, 16. American Samoa, 17. New Caledonia, 18. Norfolk Island, 19. Cape Grim.

sphere: American Samoa, Cape Point, Cape Grim, and Norfolk Island. Simulations at the latter three stations suggest inadequate emissions and transport from South Africa and from Australia, respectively. However, Cape Point and Cape Grim measurements may be biased by the complex terrain surrounding the stations. In general, local sources cause relatively more severe contamination of samples at stations with very low dust concentration, such as those in the southern hemisphere (J. Prospero, personal communication, 2002).

[73] Low concentrations at Jeju and Okinawa suggest a too weak source upwind in the Takli Makan and Gobi deserts. We appear to underpredict dust at Mace Head, similar to the study by Woodward [2001], but opposite to the study by Ginoux *et al.* [2001]. This suggests inadequate transport from North African sources, but the high variability emphasizes the importance of obtaining lengthier simulations and measurements.

[74] Long term land-based measurements of deposition fluxes are highly valuable for model evaluation since, unlike marine sediments, they reflect only atmospheric transport and can be precisely dated. Ginoux *et al.* [2001] compiled observations previously reported in the literature from 11 locations which span both hemispheres and approximately 90-station-years of measurements from the 1950s to the present. The majority of the stations were most active in the 1980s, before our simulation period. Comparisons to deposition at Shemya, Nauru, New Caledonia, Rarotonga, and Norfolk Island are not possible because only concentration was measured at these stations. The deposition measurements used by Ginoux *et al.* [2001] for these 5 stations were based on a misinterpretation of this point. Figure 8 compares

our 1990–1999 mean predictions with the 11 stations in this observed climatology. A model low bias is readily apparent at land locations in Asia and at most ocean locations. No consistent trend is apparent at the stations which sample the North African plume. Mean deposition appears high at the French Alps, somewhat low at Spain, and good in Miami. Mean deposition is underpredicted by more than a factor of 2 at both Asian land stations, Taklimakan and Tel Aviv. Since these stations are near active source regions, underpredictions there, similar to Ginoux *et al.* [2001], could result from neglecting transport of $D > 10 \mu\text{m}$ particles or missing small but nearby sources.

[75] Comparisons at Pacific stations are mixed but a low bias is evident. Mean deposition appears significantly low at Midway and Oahu in the North Pacific, although Shemya in the western Aleutians is well simulated. Simulated deposition is also low in the equatorial Pacific at Enewetak, Nauru, and Samoa, but not at Fanning. Finally in the South Pacific simulated deposition appears too low at Rarotonga, New Caledonia, and Norfolk, but not at New Zealand. The Pacific stations are generally far from emissions and so integrate deposition fluxes over large regions. Low model deposition fluxes there indicate a bias in transport and removal of smaller particles. This low bias is consistent with excessive wet scavenging upstream and/or insufficient turbulent mix-out at the surface. However, the mismatch between simulation and observation periods may contribute significantly to the apparent magnitude of the low bias.

[76] In summary, the model performs much better in terms of surface concentrations than deposition fluxes. The model tendency to underestimate the full amplitude

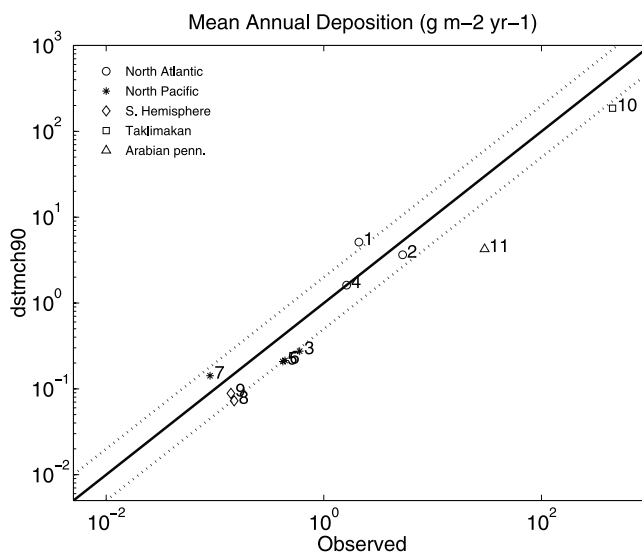


Figure 8. Predicted and observed climatological mean dust deposition flux dust in $\text{g m}^{-2} \text{yr}^{-1}$ at 11 stations compiled by Ginoux *et al.* [2001]. Site numbers are 1. French Alps (45.5N, 6.5E), 2. Spain (41.8N, 2.3E), 3. Midway (28.2N, 177.35W), 4. Miami (25.75N, 80.25W), 5. Oahu (21.3N, 157.6W), 6. Enewetak (11.3N, 162.3E), 7. Fanning (3.9N, 159.3W), 8. Samoa (14.25S, 170.6W), 9. New Zealand (34.5S, 172.75E), 10. Taklimakan (40.0N, 85.0E), 11. Tel Aviv (32.0N, 34.5E).

of the seasonal cycle of dust concentration at key stations downwind of North Africa (Barbados), East Asia (Jeju), and Australia (Cape Grim) suggests biases in model emissions. The inclusion of anthropogenic dust sources could help remediate these biases, if such sources were known. The model tendency to underestimate deposition fluxes in remote regions of the Pacific is more indicative of biases in small particle transport and scavenging mechanisms.

5. Discussion and Summary

[77] The basis of a physical model for global dust entrainment and deposition, DEAD, has been described. Entrainment is based on saltation theory adjusted to fit wind tunnel measurements. The soil erodibility is proportional to the runoff area upstream of each source region. Deposition is determined by size resolved particle and precipitation aerodynamics. We prescribe only a minimal set of ad hoc corrections to physical processes and do not invoke any anthropogenic emissions. The main features of a model climatology generated from a 1990–1999 simulation were presented. The spatial and temporal patterns of the atmospheric dust are consistent with station observations and satellite imagery on this timescale. The credibility of the simulations produced by this factor support the hypothesis that dust emission “hot spots” exist in regions where alluvial sediments accumulate and may be disturbed [Herman *et al.*, 1997; Gillette, 1999; Prospero *et al.*, 2002]. This naturally determined erodibility factor helps to constrain the role anthropogenic dust may play in present climate.

[78] On the basis of simulations of 1990–1999, we estimate the global annual mean and variability of $D < 10 \mu\text{m}$ dust to be as follows: emissions, $1490 \pm 160 \text{ Tg yr}^{-1}$; burden, $17 \pm 2 \text{ Tg}$; and optical depth at $0.63 \mu\text{m}$, 0.030 ± 0.004 . For comparison, DEAD predicts about 80% of the emissions, 60% of the lifetime, and only 45% of the burden predicted by GOCART [Ginoux *et al.*, 2001], which uses a distinct set of physical parameterizations and analyzed winds. Clearly the sensitivity of model predictions to analyzed meteorologies and model resolution should be investigated before drawing firm conclusions regarding the contribution of model physics to climatological biases. However, the differences between DEAD and observations and other models do suggest possible causes of the model biases and their potential improvements.

[79] As mentioned earlier, recent measurements from the PRIDE experiment [Reid *et al.*, 2003] show that the transport mode of African dust is about $\tilde{D}_v = 3.5 \mu\text{m}$ or larger. This is significantly greater than the $\tilde{D}_v = 2.5 \mu\text{m}$ predicted by DEAD in the Caribbean (Figure 1) and used for our subbin distribution (Table 2). A larger transport mode could help to reconcile some of the disparities between DEAD and observations. In particular, a larger \tilde{D}_v would allow predicted optical depths to remain constant or decrease at the same time as mass concentrations and thus deposition fluxes increase. Understanding the physical processes which permit efficient transport of particles coarser than the accumulation mode is a key problem for future studies. These processes may include particle size at the source, particle shape, particle density, and numerical representation of advection.

[80] DEAD treats convective and stratiform nucleation and collection scavenging using a relatively sophisticated size-resolved technique. Collection by convective precipitation and nucleation scavenging by stratiform precipitation are extremely important as these processes determine the lifetime of long-range-transported dust in our model. Although successful in many regions, this treatment does not yet capture the full phasing and amplitude of the seasonal cycle of dust over the subtropical north Atlantic, so further improvements are required. The relative roles of wet and dry deposition vary dramatically among models [Ginoux *et al.*, 2001]. However, the physical parameters which control wet scavenging are not amenable to measurement. Near-term improvements of in-cloud and below-cloud scavenging therefore may continue to depend on iteratively refining models against observables such as concentration and optical depth.

[81] The DEAD model has many limitations. The mobilization parameterization is detailed yet still omits some important processes. Factors known to have a significant influence on dust emissions which are not directly accounted for in the present model include: geographic variation of surface size distributions [Marticorena *et al.*, 1997], soil modulus of rupture [Gillette, 1988], binding by plant root matter, spacing and aspect ratio of nonerrodible roughness elements [Raupach, 1994], size-dependent energy thresholds for particle release during sandblasting [Lu and Shao, 1999; Alfaro and Gomes, 2001]. Because of the absence of these processes, the present model is one of intermediate complexity. Where warranted, these limitations may be addressed by more realistic (and expensive) physical representations. We are now testing a refined saltation-sandblasting model [Grini *et al.*, 2002], which predicts rather than prescribes the source size distribution. We are also using DEAD in CTMs to understand aerosol composition, transport, and radiative forcing during the PRIDE and ACE-Asia experiments.

[82] Some current constraints in DEAD are assumptions which will be re-examined when adequate or improved boundary data become available in the host model. Examples include accounting for spatial heterogeneity in the smooth roughness length (equation (2)), tightening the relatively weak moisture inhibition (equation (6)), and relaxing ad hoc constraints on emissions north of 60°N . However, our initial results show that an intermediate complexity model is capable of reproducing many features of the dust distribution observed in the current climate.

[83] The DEAD model and this climatology are available to all interested researchers. Visit the model homepage (<http://dust.ess.uci.edu/dead>) or contact the authors for details. DEAD also runs as a standalone box model for convenient calibration against station or wind-tunnel measurements of erosion. DEAD has been integrated into a general circulation model (GCM), the National Center for Atmospheric Research Community Climate Model version 3. Future studies employing DEAD in GCMs will evaluate the radiative forcing due to dust, and associated climate feedback of both the atmosphere and coupled climate systems.

[84] The DEAD model was designed to be physically based as much as possible. The wide range of spatial and temporal scales involved in global dust production necessi-

tated some tuning of the parametric relations which were mainly derived from wind tunnel measurements. Nevertheless, the response of dust production and transport to many key climate and landscape parameters has been included. Thus DEAD should be suitable for studies of the interaction between dust and climate change in past, present, and future climates. Currently DEAD is being used to investigate dust radiative forcing and feedback in the present climate, oxidant uptake and photochemistry, the roles of atmospheric iron and silica in marine productivity. DEAD has been built upon the physics of natural erosion so that anthropogenic dust sources are not yet explicitly represented. In the future we hope to identify and explicitly include anthropogenic dust sources.

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H. Bian, D. Newman, and C. S. Zender, Department of Earth System Science, University of California at Irvine, 205 Physical Sciences Research Facility, Irvine, CA 92697-3100, USA. (zender@uci.edu)