Modeling the atmospheric life cycle and radiative impact of mineral dust in the Hadley Centre climate model

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Abstract. A parameterization of mineral dust within the Hadley Centre atmospheric general circulation model (AGCM) is described, modeled dust distributions are compared with observations, and estimates of the radiative forcing due to the inclusion of dust in the model are obtained. The parameterization uses six particle size divisions in the range 0.3–30 μm radius, and all calculations are performed on each division independently, using the GCM’s prognostic variables. The dust production scheme works within the GCM and includes dependencies on particle size distribution, soil moisture, vegetation, and friction velocity. Dust transport is carried out by the GCM’s tracer advection scheme and includes vertical motion due to convection, gravitational settling, and turbulent mixing in the boundary layer. Wet and dry deposition processes are included within the GCM’s precipitation schemes. Representative dust radiative parameters are incorporated into the GCM’s two stream radiation code. Modeled monthly average near-surface dust concentrations are compared with observations; good agreement is seen in most locations, though the dust scheme tends to produce too little dust from China and too much from Australia in the southern spring. Global annual mean direct forcing due to the inclusion of dust in the GCM is +0.07 W m⁻² at the top of the atmosphere (TOA) and −0.82 W m⁻² at the surface. The geographical distributions of annual mean forcings are very inhomogeneous, with peak values exceeding the global means by a factor of approximately 2 orders of magnitude.

1. Introduction

As confidence in the estimates of the radiative forcing of climate due to the major greenhouse gases has increased, more attention has been focused on the radiative effects of atmospheric aerosols: first sulfate and then others, including mineral dust. Early models of mineral dust within GCMs involved the prescription of the aerosol field within the climate model [Coakley and Cess, 1985; Joseph, 1977; Tanné et al., 1984]. More recently, the complete atmospheric life cycle of mineral dust has been modeled [Joussaume, 1990; Tegen and Fung, 1994; Mahowald et al., 1999], and an estimate of the direct radiative forcing of climate due to dust from disturbed sources has been made [Tegen et al., 1996]. However, neither the magnitude nor the sign of the global mean direct radiative forcing can be considered to be determined conclusively, because of the many uncertainties inherent in any current parameterization of dust (the lack of detailed understanding of processes such as deflation and dry deposition and the wide variation in measurements of many properties from spectral refractive index to atmospheric distribution) and the heterogeneous nature of the global forcing field, as well as the strong model dependence of dust. This work describes an attempt to create a realistic simulation of atmospheric mineral dust and to provide an estimate of the direct radiative forcing of model climate due to the inclusion of this dust, as a step toward an improved understanding of the radiative effect of dust on climate.

2. Parameterization

The atmospheric GCM used for this work was a version of HadAM3 [Pope et al., 2000]. It is a finite difference gridpoint model with 2.5 × 3.75° horizontal resolution, 19 vertical levels and a 30-minute physics timestep. The model parameterizations include, among others, a two-stream radiation code [Edwards and Slingo, 1996], a stability-dependent variable depth boundary layer scheme, a mass-flux convection scheme [Gregory and Rowntree, 1990; Gregory and Allen, 1991], a multilayer hydrology scheme, and a new land-surface scheme [Cox et al., 1999]. Climatological sea surface temperatures are prescribed.

The dust scheme is an integral part of the model and uses the model’s prognostic variables throughout. The scheme may conveniently be divided into three sections, which will be described in turn: dust production, transport and deposition, and radiative effects.

Throughout the code the dust is divided into six size divisions in the range 0.03–30 μm radius, which are treated independently (Figure 1). The use of a set of size bins avoids the assumption of any particular form of the size distribution curve. The number of bins was limited by computational considerations. Little investigation has been made into the optimal dust particle size distribution for models such as this. Schultz et al. [1998] have suggested that a lognormal size distribution with mass mean diameter of 2.5 μm and standard deviation of 2.0 is dominant in the long-range transport of mineral dust, but as this work is restricted to a quite limited source area and is verified against observations over a single 7 day period, it would be rather incautious to assume this size distribution for all dust sources in all seasons. Instead, the size range was chosen to go somewhat beyond the widely used 0.1–10 μm range [Joussaume, 1990; Tegen and Fung, 1994; Lohman et al.,...
1999] as there is some evidence that nontrivial quantities of both smaller and larger particles are found [Patterson and Gillette, 1977; D’Almeida and Schütze, 1983; Betzer et al., 1988; Li-Jones and Prospero, 1998]. As dust radiative parameters are highly size dependent, the exclusion of these particles could deleteriously affect the calculated radiative forcings. It is also desirable to cover as near the full natural size spectrum as possible for the evaluation of model results against observational data. A particle density of 2.65 kg m\(^{-3}\) is assumed.

2.1. Production

The parameterization of dust emission is based on the scheme of Marticorena and Bergametti [1995], with some modifications. This has been shown to produce realistic results in a simulation of Saharan dust emissions [Marticorena and Bergametti, 1997]. The magnitude of the dust flux \(G\) is given by

\[ G = H \times 10^{(13.6F_s - 6)}, \]

where \(F_s\) is the clay fraction of the soil and \(H\) is the horizontal flux. All terms in this and subsequent equations are in SI units. This relation between horizontal and vertical fluxes was based on data from Gillette’s [1979] measurements for soils with a clay content below 20\%. The one other measurement in that data set, from a soil with 52\% clay, suggests that the relationship is not valid for high clay content soils; however, it is not clear at what level of clay content it ceases to be valid. As 90\% of land grid boxes in the model have a clay content below 23\% and none have a clay content as high as 52\%, equation (1) is used here generally, in the absence of any more widely applicable relationship. The value of \(H\) is given by

\[ H = 2.61p(1 - \nu)U^*\left(1 + \frac{U^n}{U^*}\right)^{1 - \frac{U^n}{U^*}}\frac{M_{el}}{g}, \]

where \(\nu\) is the vegetation fraction in the grid box; \(p\) is air density; \(g\) is acceleration due to gravity; \(U^*\) is surface layer friction velocity; \(U^n\) is threshold friction velocity, and \(M_{el}\) is the ratio of mass of dust in the size division to total mass. This is slightly changed from the algorithm employed by Marticorena and Bergametti, which used relative surface area rather than relative mass \(M_{el}\). The modification was necessary to avoid overproduction of submicron particles in clayey soils (which are rare in the Sahara, where the original algorithm was tested). It is based on the assumption that the size distribution

<table>
<thead>
<tr>
<th>Division Number</th>
<th>Size Range, (\mu m)</th>
<th>Scavenging Coefficient</th>
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<tbody>
<tr>
<td>1</td>
<td>0.0316–0.1</td>
<td>(2.0 \times 10^{-5})</td>
</tr>
<tr>
<td>2</td>
<td>0.1–0.316</td>
<td>(2.0 \times 10^{-5})</td>
</tr>
<tr>
<td>3</td>
<td>0.316–1.0</td>
<td>(3.0 \times 10^{-5})</td>
</tr>
<tr>
<td>4</td>
<td>1.0–3.16</td>
<td>(6.0 \times 10^{-5})</td>
</tr>
<tr>
<td>5</td>
<td>3.16–10.0</td>
<td>(4.0 \times 10^{-4})</td>
</tr>
<tr>
<td>6</td>
<td>10.0–31.6</td>
<td>(4.0 \times 10^{-4})</td>
</tr>
</tbody>
</table>

of the dust flux is, to a first approximation, similar to that of the parent soil. This is supported by observational data [Westphal et al., 1987; Sørensen, 1985].

The terms in (2) were derived using different algorithms from those employed by Marticorena and Bergametti, because of the constraints of the GCM: the vegetation fraction is prescribed annual average data obtained from the IGGBP land use data set [Loveland and Belward, 1997]; air density and friction velocity are calculated interactively from model fields, and \(M_{el}\) is derived from the GCM’s soil data set [Wilson and Henderson-Sellers, 1985]. This 1° × 1° data set contains soil clay, silt, and sand fractions, from which the fractional mass of dust in each division is obtained by assuming a simple size distribution as shown in Figure 1. Threshold friction velocity is defined as

\[ U^*_t = A \log_{10}(D_{eq}) + BW + C, \]

where \(D_{eq}\) is a representative particle diameter for each division (the logarithmic mean particle diameter), and \(W\) is the soil moisture in the model’s 10 cm deep top soil layer (in kg m\(^{-3}\)). The value of \(A\) was obtained from a straight line fit to the data of Bagnold [1941] over the 10–30 \(\mu m\) range assuming dry soil. These data were the only measurements available which extended into the particle size region of interest. The use of a linear dependence on soil moisture was taken from the work of Hotta et al. [1984]. Constants \(B\) and \(C\) were determined empirically. The necessity for empirical determination of these values is unsurprising, as the soil moisture required is that in a surface “skin” of the soil, whereas the only available model variable is the average moisture in the top 10 cm of the soil. The values used were \(A = -0.2, B = 0.5, C = -1.2\). The formula can produce negative values of threshold friction velocity for the two largest size divisions at points where the soil is exceptionally dry: less than 4% land points are affected in a year. In this case, the threshold friction velocity is set to zero. No dust is produced when the soil surface is frozen.

2.2. Transport and Deposition

The dust that has been freed from the surface is transported in the GCM as six tracers using the model’s positive definite tracer advection scheme. Vertical motion through convection, gravitational settling, and turbulent mixing in the boundary layer are included.

Wet deposition due to precipitation scavenging within and below cloud is included using a first-order removal rate:

\[ \frac{dD}{dt} = -kRD, \]

where \(D\) is dust mass, \(t\) is time, \(R\) is the precipitation rate, and \(k\) is a particle-size-dependent scavenging coefficient. The values of \(k\) (given in Table 1) were derived from experimental measurements [Volken and Schumann, 1993]. Equation (4) is
solved using the backward Euler scheme [Press et al., 1992]. Scavenging by cloud droplets is ignored, because it is a much smaller effect for the largely insoluble dust particles.

Dry deposition due to gravitational settling throughout the atmosphere is included in the scheme. In the boundary layer it is combined with turbulent mixing, using a resistance analogy method where deposition velocities are treated as inverse resistances [Seinfeld, 1986]. The resultant deposition velocity \( V_D \) is given by

\[
V_D = (R_A + R_S + R_i R_V V_S)^{-1} + V_S, \tag{5}
\]

where \( R_A \) is the aerodynamic resistance, \( R_S \) is the surface layer resistance, and \( V_S \) is the Stokes deposition velocity. The aerodynamic resistance represents the turbulent diffusion of dust from the free atmosphere to the quasi-laminar sublayer, and the surface layer resistance represents transport across the quasi-laminar sublayer: in the GCM these are combined in the simulation of transport from the center of the lowest layer to the surface. The surface layer resistance is calculated as described by Seinfeld:

\[
R_i = \frac{1}{U^* (Sc^{0.3} + 10^{-3.5St})}, \tag{6}
\]

where \( Sc \) is the Schmidt number and \( St \) is the Stokes number [Seinfeld, 1986]. The calculation of the gravitational settling velocity includes the Cunningham correction factor to account for slip flow [Pruppacher and Klett, 1997].

2.3. Radiative Properties

The radiative properties of dust have been incorporated into the model's 2-stream Edwards-Slingo radiation code [Edwards and Slingo, 1996]. The extinction coefficient, single-scattering albedo, and asymmetry parameter were calculated from refractive index data using Mie theory, assuming spherical particles. Representative values of refractive index data were used, estimated from a range of measurements made in various locations [Carlson and Benjamin, 1980; Sokolik et al., 1993, 1998; WMO, 1983]. These are listed in Table 2 and plotted in Figure 2. A high-resolution version of this radiation code has been used in a study of Saharan dust outbreaks by Haywood et al. [this issue].

3. Modeled Dust Distributions

All the results described in this section are means from runs of 5 years following 1 year spin-ups.

3.1. Qualitative Assessments

The simulated annual mean dust production is shown in Figure 3 and total column dust mass in Figure 4. The two are, of course, closely related. Qualitatively, the model appears to represent the major sources in North Africa, the Arabian Peninsula, and the Aral Sea area quite well, but other Northern Hemisphere sources, such as the Chinese and North American deserts, seem to be underestimated. The relatively large amount of dust produced at very high latitudes in the Northern Hemisphere is ascribable to the use of annual mean vegetation fields. The seasonal variation in vegetation fraction at high latitudes is much greater than at low latitudes, and moreover, because many high-latitude regions are snow covered for a large part of the year, the annual mean vegetation there is much lower than the current value for those periods when there is no lying snow and dust could, potentially, be produced. It should be noted, however, that this dust produced in the northern high latitudes makes only a very small contribution to dust loading, even on a regional scale, because most atmospheric dust is transported from the major desert sources. The geographical distribution of deposition (Figure 5) is, again, similar to the pattern of loading. This is not unexpected, but the effect is exaggerated here because the model dry deposition field includes all dust which is produced and then deposited within the same model time step; that is, it includes dust that is never lifted out of the lowest model layer. Quantitative verification of the fields is discussed below.

All dust processes are highly size dependent, as can be seen in Figure 6. The strong dependence of production on threshold friction velocity and hence on the logarithm of particle diameter is evident, as is the high order of the dependence of gravitational settling on diameter. Wet deposition also increases with particle size, through the size dependence of scavenging coefficients, but falls off in the largest size division because much of this dust has already been removed by the highly efficient dry deposition process. The resultant mass size distribution shows a clear peak in the fourth size division, corresponding to a 1–3 \( \mu m \) radius range, which agrees well with measured data for locations remote from sources [Buat-Ménard et al., 1983; Dulac et al., 1989, 1992], which might be expected to be reasonably representative of the global mean value.

<table>
<thead>
<tr>
<th>Wavelength, m</th>
<th>Real Part</th>
<th>Imaginary Part</th>
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<tr>
<td>1.000e–01</td>
<td>1.000e+00</td>
<td>0.000e+00</td>
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</tbody>
</table>

Intermediate values are obtained by linear interpolation.
3.2. Quantitative Verification

Quantitative verification of global dust distribution is somewhat problematic, due to the lack of reliable, global-scale measurements, particularly over the long timescale that is required for the verification of such a highly temporally variable distribution. Ground-based measurements of dust concentration allow the most direct verification of model fields, but they are limited to a relatively small number of stations. Airborne measurements may give very detailed data on individual dust events but cannot provide long-term climatological data and are, perhaps, better suited to use in process studies than in validation of mean model climates. Satellite data provide global-scale spatial coverage and in some cases also a long-term data record. However, the retrieval algorithm for the calculation of aerosol optical depth from the measured radiances relies on assumptions about aerosol properties that may well be different from those made in a GCM dust scheme, in which case the comparison between the modeled and the satellite fields may become effectively a comparison of two different dust models. The use of radiances calculated by the climate model for direct comparison with satellite-measured values should circumvent this problem, though distinguishing between the many processes affecting radiances would not be trivial, and the effects of dust would have to be distinguished from those of other aerosols. Indirect evidence of dust loading might be obtained from ocean sediments, which have the advantage of providing long-term mean data; however, they can only give direct validation of deposition rates and are affected by riverine input, particularly near coasts.

3.3. Comparison With Surface Observations

Model lowest-level dust concentrations have been compared with observations made at stations of the University of Miami Aerosol Network. Data from those stations with at least 4 years of observations were used to compare with model data. Figure 2 shows the spectral refractive index of mineral dust as used in the model: (top) real and (bottom) imaginary parts. Figure 3 shows the annual mean dust production (kg m\textsuperscript{-2} s\textsuperscript{-1}). Figure 4 shows the annual mean total column dust mass (kg m\textsuperscript{-2}).
of monthly mean dust concentration measurements have been used. It should be noted, however, that measurements were not continuous at all stations, and in many cases, the monthly mean values are the average of less than 10 observations. This is of particular concern because of the high variability of dust fields. The locations of the stations are shown in Figure 7. Model concentrations have been calculated using mixing ratios from the lowest model level, with surface pressure and temperature fields (surface level mixing ratio fields are not available from the model). The exception to this is the value at Izana where model data from the level nearest to the station height of 2 km have been used. All model values are grid box means: this is unavoidable but may introduce systematic errors. The observed and modeled concentrations for each station are plotted in Figure 8, which also shows the range of one standard deviation of the observational values. The stations may conveniently be divided into four geographical areas: southern high

Figure 5. Annual mean total deposition (kg m$^{-2}$ s$^{-1}$).

Figure 6. Size dependence of dust production, dry and wet deposition, and total column mass. Values plotted are the mass fractions per division of global annual mean totals of each variable.

Figure 7. Approximate locations of stations in the University of Miami Aerosol Network used here for verification. Exact locations are as follows: Cape Grim (cgr) 144.68°E, 40.68°S; King George Island (kgi) 58.30°W, 62.18°S; Mawson Station (mas) 62.50°E, 76.00°S; Palmer Station (pal) 166.05°W, 64.77°S; Funafuti (fun) 179.20°W, 8.50°S; Nauru (nar) 166.95°E, 0.53°S; Norfolk Island (nor) 170.98°E, 29.08°S; American Samoa (asm) 170.58°W, 14.25°S; Midway Island (mid) 177.35°W, 28.22°S; Oahu (ohu) 157.70°W, 21.33°N; Cheju (che) 126.48°E, 33.52°N; Fanning Island (fan) 159.33°W, 3.92°N; Ewenetak Atoll (ewt) 162.33°E, 11.33°N; Barbados (bar) 59.43°W, 13.17°N; Bermuda (ber) 64.87°W, 32.27°N; Mace Head (mah) 9.85°W, 53.32°N; Miami (rma) 80.25°W, 25.75°N; Izana (izo) 16.50°W, 28.30°N.

At Mawson Station the model generally agrees well with observations. This station is probably the least likely to be directly affected by any dust event, so this result suggests that the model's background level of dust for the Southern Hemisphere is realistic. At King George Island and Palmer Station, model concentrations are smaller than observations (though at Palmer Station, they are within one standard deviation of the
Figure 8. Monthly mean dust concentrations at each station. Asterisks represent model data for the nearest grid box. Squares represent observational means and diamonds represent the observational means plus and minus one standard deviation. The range of one standard deviation on either side of the observational mean is shaded.
mean for all but two months of the year). As the measured concentrations at these stations are higher than at Mawson Station and also the values at the more northerly King George Island are higher than at Palmer Station, it seems most likely that they are being affected by Patagonian dust. The model may be producing too little dust from Patagonia, there may be too much deposition in the region, or the circulation there may be advecting the dust in too eastward a direction rather than southward toward the pole. The second of these seems most likely, as the circulation in that area might be expected to favor eastward transport, and there is some evidence of anomalously high precipitation in the region [Pope et al., 2000]. Moreover, a reduction in the atmospheric dust produced from Patagonia might balance the overproduction from Australia in spring, thus accounting for the realistic background levels at Mawson station.

Five stations lie in the South Pacific where the dominant dust source is the Australian continent. Model concentrations at Cape Grim in the south of Australia show anomalously large values, particularly in the austral spring but also, to a lesser extent, in February and May. The main peak is most probably due to excessive dust production, as a peak is seen at another station near Australia at that season, though it may be exacerbated by anomalies in the model circulation. Norfolk Island is the island station nearest to Australia and the model agrees reasonably well with the observations here, except in October and November, when the model’s peak dust production appears to be a month later than observed. At Nauru and Funafuti, which lie somewhat farther north, outside the region most strongly affected by Australian dust, agreement is generally good. American Samoa lies farther east: here model results agree with observations, except in the austral spring, where the problem of excess Australian dust is again evident.

One station in the North Pacific (Cheju) is close to the Asian continent and therefore relatively close to the major dust sources. Results at this location show a problem with model concentrations: the tendency for the model to underestimate dust production from the Chinese desert regions, which is particularly evident in Northern spring, when the model fails to represent the seasonal maximum in concentration. Enewetak is the station next closest to the continent, but here agreement is much better, though the modeled concentration is too high in April. This may suggest that the modeled size distribution of the Chinese dust may be wrong, with too many small particles that are transported long distances but too few of the larger particles that give high concentrations nearer the source; it may suggest a certain amount of overproduction from other Northern Hemisphere sources, such as those in the Aral Sea area, to balance the lack of Chinese dust; or it may indicate an unrealistic model circulation in this area. Midway and Oahu are located in the central North Pacific, remote from any sources. Here agreement between model and observations is good, apart from a slight excess of dust in the model in April, implying that in the Northern Hemisphere the background level of dust is fairly realistic for most of the year. At Fanning Island, which is close to the equator, the model agrees well with the observations for the first part of the year but shows excessive dust levels between September and January: this may be due to small errors in the position of the model’s Intertropical Convergence Zone (ITCZ), because the high levels of modeled dust at this time are most likely to be due to Australian dust advected into this zone.

At the one North Atlantic station that is close to the Sahara (Izana), there is generally good agreement between the model and the observations, except in December and January when the model seems to produce slightly too much dust from North Africa. Three stations are near the western end of the Saharan plume. At Bermuda and Miami the modeled concentrations agree quite well with observations. At Barbados the model overestimates dust concentrations for the first three months of the year, though the agreement with observations is reasonable for the remaining months; this overestimation suggests a fault with the model’s representation of the Atlantic ITCZ. At Mace Head the agreement is generally good.

Overall, the results confirm the qualitative impression that the model simulates the life cycle of dust from many sources reasonably well and produces generally realistic seasonal cycles and background concentrations.

The major shortcomings appear to be in the underestimation of Chinese dust and overestimation of dust from Australia. Such shortcomings are common in global dust models [Lohmann et al., 1999; Tegen and Fung, 1995; Mahowald, 1999]. Though their origins are not clear, they are likely to be related to some of the uncertainties described in section 5.

3.4. Comparison With Ocean Sediment Data

The usefulness of ocean sediment data for verification is somewhat limited but is probably greatest in areas remote from river outflows, in areas where no direct measurements of dust concentration exist, and in terms of major trends rather than details.

Modeled deposition rates have been compared with ocean sediment data [Duce et al., 1991]. The results broadly agree with those of the previous section. Anomalously low model deposition rates in the Northwest Pacific is further evidence of underproduction of dust from China, but deposition rates in the rest of the North Pacific agree well with measured values. Modeled deposition rates around Australia are a little higher than those from the sediment data, but modeled deposition in the Southern Ocean is a little lower, supporting the previous evidence of problems with the production or transport of dust from Australia in the model. Model and observations tend to agree in the Indian Ocean, though there are differences in the Arabian Sea, which may be due to the model overproducing dust from the Aral Sea area. Deposition from the Saharan plume appears to be a little higher in the model than in the measurements at the southern edge of its range, which is consistent with the results of the comparison of atmospheric concentrations at Barbados. There appears to be less modeled deposition in the Northwest Atlantic, which is consistent with underproduction from the deserts of North America (see also Figure 3), otherwise values in the North Atlantic agree quite well. There is also good agreement over most of the South Atlantic, with the exception that Patagonian dust is being carried eastward in the model, and the deposition data suggest a westward transport; the reasons for this are not clear.

4. Modeled Radiative Forcings

The results described in this section were obtained with a “double radiation call” method, in which the radiation scheme was called twice at each radiation time step, but dust radiative effects were calculated only in the first call. The second call was used to progress the model, ensuring that there was no feedback of dust effects in the model’s evolution. The direct radi-
ative forcings due to dust were obtained from the differences in fluxes from the two calls.

The annual mean net surface forcing due to the inclusion of mineral dust in the model is shown in Plate 1 (top). The global mean value is \(-0.82\) W m\(^{-2}\), made up of a shortwave component of \(-1.22\) W m\(^{-2}\) and a longwave component of \(+0.40\) W m\(^{-2}\). Each has a similar distribution, which broadly follows the distribution of dust loading. Peak values in excess of \(-10\) W m\(^{-2}\) in the annual mean are located off the North African coast.

The corresponding modeled net TOA forcing is shown in Plate 1 (bottom). The global mean annual mean value is \(+0.07\) W m\(^{-2}\), made up of \(-0.16\) W m\(^{-2}\) from the shortwave and \(+0.23\) W m\(^{-2}\) from the longwave. This is in fairly good agreement with the value of \(+0.14\) W m\(^{-2}\) obtained by Tegen et al. [1996], bearing in mind that a completely different GCM and a completely different dust parametrization were used.

The pattern of longwave forcing is again similar to that of the dust loading, but the shortwave forcing field is much more heterogeneous. The clear-sky TOA shortwave forcing (Plate 2 (top)) is positive over bright surfaces, such as ice, snow, and bright desert, and negative elsewhere. The forcing is also positive over cloud (Plate 2 (bottom)). The areas of largest forcing, exceeding \(\pm1\) W m\(^{-2}\) in the annual mean, can be seen in the regions of highest dust loading over and downwind from the North African and Australian deserts.

Including the full range of dust particle sizes is important for obtaining an accurate estimate of dust forcing. A set of three double radiation call experiments were performed, each with dust in only two of the six size divisions, covering small (0.03–0.3 \(\mu\)m), medium (0.3–3.0 \(\mu\)m) and large (3.0–30.0 \(\mu\)m) particle radius ranges. The quantities of dust in each division in each experiment were the same as in the corresponding divisions in the same period of the full experiment. The forcings estimated from these experiments are listed in Table 3. In summary, at the surface the global annual mean forcing is always negative and is dominated by the medium-size particles; the small particles contributing 15% and the large particles only 3%. At the TOA the global annual mean forcing from medium-size particles is again similar to the total forcing, but the situation is complicated by the fact that the forcings from all three size groups are similar in magnitude, though not in sign. The TOA annual mean forcing due to small particles is negative in most places, that due to medium-size particles is negative in some areas and positive in others, and that due to large particles is positive everywhere. These results suggest that had the small particles been omitted from the main experiment, the estimate of global annual mean TOA forcing would be almost doubled to \(-0.15\) W m\(^{-2}\), and had the large particles been omitted, it would be greatly reduced to \(-0.004\) W m\(^{-2}\). It is conceivable that omission of the large particles could even change the sign of the TOA forcing in another experiment.

The global annual mean net TOA forcing is the residual of the sum of much larger positive and negative values and is strongly dependent on dust and cloud distributions, neither of which can be modeled very accurately at present. It has been quoted here because it is widely used in the climate modeling community, but its usefulness is perhaps questionable when the field is so heterogeneous and has such large uncertainties associated with it.

### 5. Discussion

There are numerous uncertainties associated with the relatively new field of GCM dust modeling, some of which are outlined below. The great majority of them are applicable to any such current modeling endeavor, rather than confined to this scheme.

There are a number of limitations associated with the dust production algorithms: the Marticorena and Bergametti scheme, upon which the dust flux algorithm is based, is obtained using Saharan data and might be expected to apply to similar deserts, however it may not be equally valid for soils in other areas which are only beginning to experience the effects of wind erosion and have very different particle size distributions. The reentrainment of settled dust has been ignored for the sake of simplicity as has the effect of sub-grid-scale orography, though both processes affect dust production [Pye, 1987]. The effects of surface crusting, soil salt and organic content, and surface geomorphology have been omitted due to lack of data. The effect of vegetation on roughness length and hence friction velocity has also been excluded: the vegetation parameters available from global vegetation data sets do not allow this to be calculated easily; for example, “vegetation fraction” refers to canopy fraction, rather than fractional areal coverage at ground level.

The use of a threshold for friction velocity, which includes a soil moisture term, renders the dust production very sensitive to any errors in these two model fields. Also, the linearity of the dependence of threshold friction velocity on soil moisture must be regarded as somewhat uncertain. Other workers have noted a logarithmic relationship [Belly, 1964], but it was found to be impossible to create even a vaguely realistic distribution of dust production from this model when a logarithmic relation was used. A dependence on soil moisture is required, however, if a temporally and spatially realistic distribution of dust production, related to local conditions, is to be achieved.

Both transport and deposition may be affected by any GCM systematic errors, particularly in the circulation, precipitation, and cloud fields: such errors will, of course, vary between GCMs. Lack of knowledge about dry deposition processes is also a problem. The parameterization that has been chosen is fairly simple; several more complex formulae exist and give a wide range of deposition velocities which vary by more than an order of magnitude over the dust particle size range [Gallagher et al., 1997].

Considerable uncertainties are associated with any calculation of dust radiative effects based on mean radiative parameters, yet there are currently insufficient data available for dust from each major source to be represented separately. Dust radiative properties are highly dependent on particle size and mineralogical composition. These are interrelated, as minerals have characteristic particle sizes, but particle size distribution is associated with atmospheric residence time and hence location, as large particles are preferentially removed by gravitational settling and wet deposition. The use of a single spectrum of refractive indices for all atmospheric dust involves considerable approximation, illustrated by the wide range of measured values of the imaginary refractive index, which spans nearly an order of magnitude over large sections of both the shortwave and the longwave spectral regions [Sokolik et al., 1993]. A systematic error is also present due to the lack of data for dusts from some major sources, such as the Chinese and Australian deserts. The use of a single set of radiative proper-
Plate 1. Annual mean net forcing due to dust (W m\(^{-2}\)) at (a) surface and (b) TOA.
Plate 2. Annual mean TOA shortwave forcing due to dust (W m$^{-2}$). (a) Clear-sky conditions. (b) Clear and cloudy conditions.
ties instead of a separate set for each of the major mineral constituents may lead to errors in estimated forcing [Claquin et al., 1999; Sokolik and Toon, 1999; Claquin et al., 1998]. The modeled forcing may also be affected by the surface albedo field, by the assumption of spherical particle shape (which may introduce uncertainties of up to 15% [Mishchenko et al., 1997]) and by the vertical distribution of dust: the strongest effect of changes in the dust distribution being due to the relative altitudes of the dust layer and of cloud, as this can change the sign of the TOA shortwave forcing.

Finally, errors and approximations in the particle size distribution will also affect the modeled radiative forcings, because of the strong dependence of dust single-scattering parameters on particle size [Tegen and Lacis, 1996]. As each stage of the dust scheme is highly particle-size dependent and, due to computational restrictions, only a few particle size bins can be used to approximate the spectrum, this is potentially a major source of error.

The presence of so many uncertainties suggests that there is a large error range associated with the results but does not invalidate them. The verification of dust distributions shows good agreement between model fields and observations in most locations, suggesting that the results are potentially useful over large areas of the world.

6. Summary

A parameterization of dust deflation, transport, deposition, and radiative effects has been developed for the Hadley Centre climate model. The simulated dust distributions have been compared with observations of surface level concentrations and deposition rates and found to be in reasonable agreement with them in most areas, and the seasonal cycle generally seems to be well represented. The model’s underproduction of dust in China and overproduction in Australia are problems that remain to be addressed.

The estimated global annual mean TOA direct radiative forcing of model climate due to the introduction of mineral dust is +0.07 W m\(^{-2}\). This is distributed very inhomogeneously, with generally negative forcings over low-latitude oceans and dark vegetation and positive forcings elsewhere. Local maxima and minima are approximately 2 orders of magnitude larger than the global mean. The pronounced latitudinal variation of the forcing suggests that were dust radiative effects allowed to feed back into the model, the circulation might be affected. The corresponding global annual mean surface forcing is ~0.82 W m\(^{-2}\).

A simulation with less Australian and more Chinese dust would tend to give a more positive global mean annual mean TOA forcing: in the annual mean, there is more cloud over China and the Northwest Pacific than over Australia and the Southeast Indian Ocean, which results in a positive annual mean TOA forcing over most of the range of Chinese dust, whereas the presence of Australian dust tends to result in a positive forcing over land and a negative forcing over sea. This effect would be exaggerated in the HadAM3 model, which tends to produce too much cloud in the North Pacific and too little in the Australian dust region.

The simulated forcings must be regarded as somewhat tentative, given the uncertainties associated with the parameterization. However, the majority of these uncertainties are not specific to the Hadley Centre model, and so the estimated forcings may be useful, together with results from other models, in giving an idea of the possible range of values.

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