Model simulations of dust sources and transport in the global troposphere. Effects of soil erodibility and wind speed variability

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Abstract. Global atmospheric dust is simulated using DEAD (Dust Entrainment and Deposition model) in combination with the chemical transport model Oslo CTM2 with meteorological data for 1996. Dust sources are calculated both using wind speeds in model resolution (1.9 x 1.9 degrees) and using different assumptions on soil erodibility and on wind speeds. Some aspects of the annual dust cycle (such as the east Asian dust emissions) are largely dependent on the data used to determine soil erodibility. Other aspects (such as the timing of the maximum in the African plume at Northern hemisphere summer) is well modeled with all datasets applied here. We show that the daily variation in optical depth at Cap Verde at the west coast of Africa is very well simulated assuming that erodibility is correlated with surface reflectivity from MODIS satellite data. Using a sub-grid probability density function of wind speed to drive the dust sources facilitates dust emissions in areas with low wind speeds. Dust concentrations in remote areas are sensitive to the parameterization of wet deposition. Our results point out the need for a detailed soil erodibility dataset for global dust modeling, and they suggest that MODIS surface reflectivity data is potentially valuable for evaluating such datasets.

1. Introduction

Atmospheric mineral dust plays a role in regulation of the earth’s climate.

The dust aerosols modify the atmospheric radiation balance by scattering and absorbing radiation and in this way alter the Earth’s energy budget [Tegen and Lacis, 1996; Myhre and Stordal, 2001]. Haywood et al. [1999] showed the importance of aerosols for the global radiative balance when calculated fluxes of reflected sunlight from a General Circulation Model (GCM) and found that they could not match Earth radiative budget Experiment (ERBE) measurements unless aerosols were taken into account. The minerals can interact both with solar and terrestrial radiation.

Tegen et al. [1996] proposed that mineral emissions from disturbed soils could cause a local top of the atmosphere forcing of -2.1 to +5.5 W m⁻² giving a globally averaged forcing of +0.09 W m⁻² for anthropogenic dust and +0.14 W m⁻² for total dust. Later studies have proposed that Saharan dust has a rather high Single Scattering Albedo (SSA) around 0.95 or higher at 550 nm [Kaufman et al., 2001] leading mostly to a cooling effect of dust aerosols. Measurements during the SHADE experiments confirmed this high value for [Tanre et al., 2003].

The minerals can also interact with atmospheric photochemistry. Dentener et al. [1996] proposed an important depletion of NOx on dust. Bian and Zender [2003] showed that dust can suppress atmospheric OH significantly in source areas due to lower incoming UV radiation. Heterogeneous chemistry and photolysis change can have opposite effects. For example will dust in remote areas lead to less photochemical destruction of Ozone. On the same time it will lead to loss of Ozone due to heterogeneous uptake on the dust aerosols.

Aerosols can serve as cloud condensation nuclei [Pruppacher and Klett, 1997]. Mineral aerosols are also supposed
to be important as ice nuclei, and they are therefore important for ice cloud formation [Lohmann, 2002]. Sherwood [2002] argued that dust or biomass can influence stratospheric water vapor because a decrease in the size of ice nuclei (and an increase in their number concentration) would increase evaporation near the tropical tropopause.

Dust contributes to transport of nutrients to ocean areas [Prospero, 1996]. Iron can regulate production of phytoplankton in the oceans. The phytoplankton can be one of the factors regulating the atmospheric CO2 flux to the oceans. Falkowski et al. [1998] proposed that in the contemporary ocean, photosynthetic carbon fixation by marine phytoplankton leads to formation of 45 gigatons of organic carbon per annum of which 16 gigatons are exported to the ocean interior. It has also been proposed that dust can bring nutrients such as phosphate to the Amazon rain forest [Swap et al., 1992].

All the factors described above make it important to understand what factors regulate the production, transport and loss of mineral dust in the global atmosphere.

Several studies have tried to quantify the global production and transport of atmospheric mineral dust [Tegen and Fung, 1994; Claquin, 1999; Ginoux et al., 2001; Woodward, 2001; Zender et al., 2003a]. Early estimates for global production lie in the range 500 - 5000 Tg/yr. This number is largely dependent on the size distribution of the dust which is produced. Prospero [2003] evaluated dust concentration data at Barbados together with Saharan rainfall data and proposed that dust production is closely coupled to rainfall in the prior year. The uncertainties are large with respect to production, loss and transport of dust, and with respect to physical and chemical characteristics of dust (e.g. Myhre and Stordal [2001]).

Early modeling studies (e.g. Tegen and Fung [1994]) calculated dust production and transport using as hypothesis that all desert was equally available for soil erosion. Supposing that all deserts are erodible and changing vegetation data results in large change in dust production. Those changes were attributed to human influence by Tegen and Lacis [1996] and Sokolik and Toon [1996]. The weakness of those studies is that they give a large human influence on dust production without taking into account that soils might be non erodible.

Later studies [Ginoux et al., 2001; Prospero et al., 2002] have focused on the existence of dust “hot spots” which are dry lakes where sediments earlier have accumulated and are now released. The main conclusion from these works is that dust production is closely tied to the water cycle and river flow. If there is a human influence on dust production, this influence must be examined in connection with drying of rivers and lakes, and not only with changed vegetation. Applying these theories, Tegen et al. [2002] simulated soil erodibility using a water routing and storage model. Zender et al. [2003b] calculated the erodibility by assuming it was proportional to the upstream area from which sediments may have accumulated locally through all climate regimes.

Schaaf et al. [2002] showed how land types can be observed retrieving surface reflection with the MODerate Imaging Spetroradiometer (MODIS) satellite. Even though surface reflection is largely dependent on soil types, Tsvetkskaya et al. [2002] point out that sand dunes have the highest surface reflectivity of all surface types.

We suppose that sand dunes are easily erodible. Therefore we would like to find out whether areas with high reflectivity also are areas with high erodibility. To examine the validity of this hypothesis, we run a 3D global dust production and transport model drive by meteorological data from 1996 to calculate atmospheric dust production and transport.

Since our goal is to explore the connection between soil reflectivity and soil erodibility, we introduce two new erodibility factors based on MODIS surface reflection and compare them to model results using the earlier published erodibility factors.

We also explore the importance of using a probability density function of wind speeds [Justus et al., 1978] to drive dust production rather than using mean winds at model resolution.

We confirm that global dust budgets and concentrations from the DEAD/Oslo CTM2 model are within reasonable range. We compare our results both to earlier published results and to measurements.

2. Modeling

2.1. Dust production

Dust emissions are modeled using the Dust Entrainment and Deposition model (DEAD) [Zender et al., 2003a]. This model is based on the work of Marticorena and Bergametti [1995]. Emissions start when wind friction speeds reach a threshold wind friction speed of approximately 0.2 m s−1 [Iversen and White, 1982]. A horizontal saltation (soil) flux and a vertical (dust) flux are calculated. The size distribution of the vertical dust flux is distributed to three modes according to [D’Almeida, 1987] The most important mode (96 % of the mass) has a mass median diameter (MMD) of 4.82 μm.

We do not explicitly take into account production of dust by saltation and sandblasting (e.g. Shao [2001]). The size distributed production of dust is a complex interplay between wind speed and soil properties. This interplay can cause differences in dust fluxes of several orders of magnitude [Alfaro and Gomes, 2001; Grini et al., 2002b]. Grini and Zender [2004] conclude that input data needed to drive dust production with these mechanisms are not available on global scale.

The dust production calculated by the physical equations are modified by two factors:

1. A global Tuning factor, T: This factor is determined “a posteriori” and ensures that global emissions in all simulations are the same. T is globally constant.
2. An erodibility factor, RDBFCT: This factor is described in detail below. The factor is meant to take into
Simulations of dust sources and transport

account that some desert surfaces are easier to erode than others [Prospero et al., 2002].

The total emissions in a given grid, is thus

\[ EM = EM_{\text{phys}} \times RDBFCT \times T \]

where \( EM_{\text{phys}} \) corresponds to the emissions modeled by physical equations and EM to total emissions.

2.2. Soil moisture

Soil moisture inhibits dust production [Fecan et al., 1999]. Soil moisture in desert areas is very low. The parameterization of evaporation used in the ECMWF model [Viterbo and Beljaars, 1995] does not capture variations in soil moisture at the scale described by Fecan et al. [1999]. Evaporation from soils in the ECMWF model stops at a globally constant permanent wilting point of 0.171 m^3 m^{-3}.

Instead of tuning the Fecan et al. [1999] parameterizations to fit the ECMWF soil moisture, we have chosen to develop a simplified approach based on rainfall. Our approach takes into account two factors:

1. After a rainfall, the soil has to dry before it can start producing dust.
2. The soil needs longer time to dry after a large rainfall than after a small rainfall.

The time required by the soil to dry depends on air temperature and humidity, surface winds and soil texture. To implement these processes, we would need to build a whole new soil moisture model which is beyond the scope of this study.

To get around the problem, we made the following simple assumptions:

1. the production of dust is stopped if precipitation during the last 24 hours is larger than 0.50 mm.
2. The length of the period without emissions (in days) is equal to the amount of rain during the last 24 hours (in mm)
3. If no rain has fallen the last 5 days, the soil is assumed to be dry no matter the size of the last rainfall

A similar approach has been used by Claquin [1999] and Myhre et al. [2003]. Claquin [1999] used a more sophisticated way to calculated the time needed for the soil to dry (taking into account soil temperature).

2.3. Soil erodibility

In this study, we use 4 erodibility factors to simulate dust production. Two of them are already published ([Ginoux et al., 2001] and Zender et al. [2003b]). Ginoux et al. [2001] applied the idea that wherever there are large basins in the world, rivers and lakes would have accumulated loess and sand to give large erodibility. This reasoning led to the use of the simple formula

\[ TOPO = \left( \frac{z - z_{\text{min}}}{z_{\max} - z_{\text{min}}} \right)^5 \]

where TOPO is the erodibility factor with which all dust emissions are multiplied.

Zender et al. [2003b] calculated the erodibility assuming it was proportional to the upstream area from which sediments may have accumulated locally during different climate regimes. (The details of the algorithms used, can be found in Jenson and Domingue [1988]). Using a global transport model, they found that correlation with measurements improved when changing the Ginoux et al. [2001] erodibility factor with the more advanced one. The erodibility factor proposed by Zender et al. [2003b] is called GEO in the rest of this work.

Two new erodibility factors are based on the assumption that erodibility is correlated with surface reflectance. We used the data set MOD09 [Schaaf et al., 2002] from the MODIS satellite and produced two new erodibility factors which are used in runs labeled MDSLNR and MDSSQR. The factor MDSLNR is calculated according to Equation 3 and MDSSQR is calculated according to Equation 4. The erodibility described by the MDSLNR will be our base case in this work.

\[ MDSLNR(i, j) = \frac{SR(i, j)}{SR_{\text{max}}} \]

\[ MDSSQR(i, j) = \frac{SR(i, j)^2}{SR_{\text{max}}^2} \]

where SR means surface reflectance.

We compare the erodibility factors in Figure 1.

Figure 1 shows that in the Sahara, all three datasets propose high erodibility in the West (Mali/Mauritania/Algeria) area, south-east (Lake Chad) and east (Egypt/Libya/Sudan/Chad). Even though the placements show similarities, both Ginoux et al. [2001] and Zender et al. [2003b] propose high erodibility in Mauritania further west than MODIS which has highest reflectance in a square between (12°W 17°N) and (3°W 22°N). In eastern Sahara, Ginoux et al. [2001] does not predict the same area as MODIS and Zender et al. [2003b]. Ginoux et al. [2001] predicts high erodibility in a small area in North-Eastern Libya.

In East Asia, MDSLNR and MDSSQR give low erodibility in both Taklamakan and Gobi deserts compared to Zender et al. [2003b] and Ginoux et al. [2001]. However, the geographical placements of the maximum is approximately equal in all datasets.

In Arabia, all datasets agree on high erodibility in Southern Saudi Arabia. MODIS gives high reflectivity in northern Saudi Arabia too, whereas both Ginoux et al. [2001] and Zender et al. [2003a] gives a maximum in Iraq along the Euphrat and Tigris rivers which is not indicated by MODIS.

In Australia, all datasets propose higher erodibility in the Lake Eyre basin (South East) compared to the great Sandy desert (North west).
Figure 1. Erodibility factors obtained by four different methods. Upper left is the method of Ginoux et al. [2001], upper right is the method of Zender et al. [2003b], lower left is the factor described in Equation 3 and lower right is the factor described in Equation 4. Note that the images have different color scale.
2.4. Probability density function of wind speeds

We use a probability density function (PDF) of winds speeds proposed by Justus et al. [1978] to drive dust emissions. The Weibull distribution is described by a shape factor \( k \) and a scale factor \( c \). The factors are calculated from wind speed at reference height (10 m) from Equation 5 and 6.

\[
k_{ref} = 0.94 \sqrt{U_{ref}} \tag{5}
\]

where \( k_{ref} \) is the Weibull distribution shape factor, and \( U_{ref} \) is the wind speed at reference height.

\[
c_{ref} = \frac{U_{ref}}{\Gamma(1 + \frac{1}{k_{ref}})} \tag{6}
\]

where \( c_{ref} \) is the Weibull distribution scale factor and \( \Gamma \) is the gamma function. Further discussion on the Weibull distribution or the gamma function can be found in any textbook on probability, e.g. Dougherty [1990].

To get the shape of the probability density function at the midpoint height, we use the formula proposed by Justus et al. [1978]:

\[
k_{mdp} = k_{ref}[1-0.088 \ln(z_{fr}/10)]/[1-0.088 \ln(z_{mdp}/10)] \tag{7}
\]

\( c_{mdp} \) is found using (6) with values for midpoint height instead of reference height.

We use a 95 % percentile in the distribution, meaning that the lowest wind speed taken into account is the one where 95 % of the winds are lower. This is given using the cumulative form of the Weibull distribution to describe the winds used to drive dust production.

\[
p(U < U_x) = 1 - \exp\left(-\left(U_x/c\right)^k\right) \tag{8}
\]

where \( p \) is the probability of wind being lower than the wind \( U_x \).

2.5. Atmospheric transport

We use the Oslo CTM2 for atmospheric transport. The model is a chemical transport model which is described in Sundet [1997]. It has been used for several chemistry and transport studies [Grini et al., 2002a; Myhre et al., 2003; Berglen et al., 2003; Endresen et al., 2003; Gauss et al., 2003]. It is an off line model driven by ECMWF forecast data. Advection is done with second order moment method [Prather, 1986], convection is done with the mass flux scheme of Tiedtke [1989].

2.6. Wet deposition

Wet deposition is a very difficult process to model. The hygroscopical properties of dust are not well known. We expect the hygroscopical properties of dust to change during transport. The dust can for example get coated with water soluble organics or sulfate which will make it easier to lose during rain events. To calculate the right wet deposition, one would need a detailed description of cloud and aerosol microphysics (see e.g. Ghan et al. [1998]; Nenes et al. [2001]).

Because of the complexity of wet deposition, we chose to include a simple non size dependent wet deposition scheme. To test sensitivity to different washout efficiencies, we use a parameter \( \epsilon \) to describe the efficiency with which dust removed by rain (Equation 9) and we test sensitivity to \( \epsilon \) using values equal to 1.0 and 0.3.

Our model includes two different types of wet deposition:

2.6.1. Large scale Wet deposition is done using three-dimensional rainfall data and assuming that dust washout is proportional to rainout, cloud liquid water and cloud fraction. Re-evaporation is taken into account only if all the rain evaporates.

\[
loss = \epsilon \cdot C_{dust} \cdot CLDFRC \cdot \frac{RAIN}{CLDLWC} \tag{9}
\]

where \( loss \) is the loss in kg, \( \epsilon \) is a factor between 0 and 1, \( C_{dust} \) is dust concentration in kg, CLDFRC is the fraction of the grid cell covered by cloud, RAIN is rainfall in the timestep (kg), and CLDLWC is the cloud liquid water (kg).

2.6.2. Convective This process removes dust whenever the air rising in a convective tower gets super saturated. It is important to couple removal directly to the convective transport so that dust is not first transported to high altitudes by convection before any removal algorithm is applied. After transport, the dust would no longer be available in the clouds to be removed by rain. The convective rain removes dust with an efficiency \( \epsilon \) equal to the one described for large scale wet deposition. The convective removal is described in detail by Berglen et al. [2003].

2.7. Dry deposition

The dry deposition uses a resistance method as described in Zender et al. [2003a] where dry deposition velocity is dependent on wind friction speed and surface stability. Dry deposition is largest for the large aerosols and the areas with high wind friction speed [Seinfeld and Pandis, 1998].

3. Description of model runs

We did 6 model runs. The runs where normalized using different tuning factors (T) so that all gave the same annual emissions of 1500 Tg yr\(^{-1}\). The runs are described in Table 1.

We use four different descriptions of soil erodibility: The soil erodibility (Equation 1) is varied in simulation 1-4. In Table 1, the different erodibility data used are labeled MD-SLNR (Equation 3), MDSSQR (Equation 4), GEO [Zender et al., 2003a] and TOPO (Equation 2).

All simulations except simulation 5 use the Weibull distribution to describe the winds used to drive dust production. Simulation 5 use mean wind speeds at model resolution. The differences between simulation 5 and 1 show sensitivity to
ignoring wind speed variability in dust production. The simulations using Weibull winds are labeled “Yes” in Table 1, and the simulation using mean wind is labeled with “No” in Table 1.

The efficiency of wet deposition is changed in simulation 6. The parameter $\epsilon$ applied in Equation 9 was varied between 0.3 (simulation 6) and 1.0 (simulation 1). Differences in simulation 1 and 6 show sensitivity to washout efficiency. (Section 2.6).

4. Results and sensitivity tests

4.1. Global Budgets and fluxes

Table 2 shows the global fluxes in the different runs. All runs are prescribed a total of 1500 Tg yr$^{-1}$ emissions. Approximately 50-60% of the dust mass is lost by dry deposition in all simulations. 30-40% is lost by large scale wet deposition, and 10% is lost by convective rainout. In the simulation where rainout is less effective, the dry deposition fraction is highest.

Decreased rainout gives increased lifetime of dust. The global and yearly averaged burden increases by 55% when only 30% of the dust is assumed to fall out as rain. Figure 2 shows where the different loss processes are important. Dust close to the sources is mainly lost by dry deposition. Wet deposition is more important in remote ocean regions. This difference reflects change in dust size distribution during transport. The largest aerosols (which contain most mass) fall out close to the source areas. Another reason is that close to the sources (in the deserts) there is not much rain which can give wet deposition. The convective washout is (as expected) most efficient close to the equator, in the Inter Tropical Convergence Zone (ITCZ).

4.2. Nutrient budgets

As mentioned in Section 1, dust can participate in biological processes. For example, dust is an important nutrient for phytoplankton in the oceans. Modifying oceanic primary production can influence the ocean carbon cycle.

Table 3 shows the yearly deposition to different ocean and forest regions. Biological processes in the ocean and forests are limited by trace metals such as iron and phosphate. All the simulations yield approximately 400 Tg yr$^{-1}$ deposition to the oceans. Zender et al. [2003a] calculated a total flux of 315 Tg yr$^{-1}$ and Prospero [1996] gives values between 358 Tg yr$^{-1}$ and 910 Tg yr$^{-1}$.

The Amazon forest receives 3-4 Tg yr$^{-1}$. Simulation 6 (with reduced rainout) yields 9 Tg yr$^{-1}$ deposition to the Amazon forest. The increased lifetime of dust in simulation 6 allows more dust to be transported to the Amazon forest before being deposited. For the same reason, simulation 6 gives higher deposition to the Northern Pacific. When reducing wet deposition rate, Asian dust is transported longer distances before it is lost by rainout. Swap et al. [1992] propose that 14 Tg yr$^{-1}$ of dust are deposited in the Amazon forest.

4.3. Global production

4.3.1. Effect of erodibility

In Figure 3, we compare the yearly averaged dust production for simulation 1-4. These simulations all use different erodibility factors. We show total production in kg m$^2$ s$^{-1}$ for simulation 1 and the absolute difference from simulation 1 to simulation 2-4. The results show that production is extremely dependent on erodibility. Figure 3 leads to the following main conclusions:

1. Simulation 3 and 4 yield higher emissions than simulation 1 and 2 in East Asia (in both the Taklamakan and Gobi deserts).
2. In the Western Sahara, simulation 3 and 4 give emissions further west than simulation 1 and 2. Simulation 3 and 4 also gives high emissions in an area close to (0E, 25N) which is slightly further north than the most active areas in simulation 1 or 2.
3. Simulation 4 gives higher emissions in Somalia than the other simulations.
4. Simulation 3 and 4 give higher emissions in India than simulation 1 and 2.
5. Simulation 3 and 4 give higher emissions around the Caspian sea than simulation 1 and 2.
6. Simulation 1 and 2 have high emissions in Northern Saudi Arabia instead of in Iraq (simulation 3 and 4).
7. In Australia, simulation 3 and 4 give higher emissions in the lake Eyre basin rather than in the Great Sandy desert which is the opposite of simulation 1 and 2. Simulation 3 and 4 also give higher total Australian production.
8. Simulation 3 and 4 give higher emissions in North America than simulation 1 and 2.

4.3.2. Effect of wind speed variability

Figure 4 shows the effect of ignoring wind speed variability. The figure is difficult to interpret since all emissions are fixed to 1500 Tg. Ignoring wind speed variability reduces total emissions. We need a tuning factor twice as high for the emissions when wind speed variability is ignored. Figure 4 actually shows the areas where the increase in dust production due to increasing the tuning factor (Equation 1) is larger than the “natural” increase when including wind speed variability.

In our simulations, ignoring wind speed variability (and keeping total emissions fixed) increases yearly Australian emissions. It increases emissions around the Caspian sea and in the Gobi desert and decreases emissions from Taklamakan in Asia. In Western Sahara, the increase in emissions when ignoring wind speed variability is co-located with preferential source areas in Mali, Mauritania and Algeria. In eastern Sahara, the Egyptian and Libyan emissions are increased. Ignoring wind speed variability increases Somali emissions.

Areas with high wind (e.g. coastal areas) produce more dust when ignoring wind speed variability. This is explain-
Table 1. Description of model runs. Dust production is changed using different assumptions on soil erodibility and wind variability. Wet deposition loss of dust is changed in simulation 6

<table>
<thead>
<tr>
<th>Run Number</th>
<th>Erodibility</th>
<th>PDF Winds</th>
<th>Washout</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>MDSLNR</td>
<td>Yes</td>
<td>100 %</td>
</tr>
<tr>
<td>2</td>
<td>MDSSQR</td>
<td>Yes</td>
<td>100 %</td>
</tr>
<tr>
<td>3</td>
<td>GEO</td>
<td>Yes</td>
<td>100 %</td>
</tr>
<tr>
<td>4</td>
<td>TOPO</td>
<td>Yes</td>
<td>100 %</td>
</tr>
<tr>
<td>5</td>
<td>MDSLNR</td>
<td>No</td>
<td>100 %</td>
</tr>
<tr>
<td>6</td>
<td>MDSLNR</td>
<td>Yes</td>
<td>30 %</td>
</tr>
</tbody>
</table>

Table 2. Global budget for loss fluxes and burden. The fluxes are expressed as percentage of mass. The annual total production flux is 1500 Tg. The burden is given in Tg. DRYDEP is dry deposition flux (Section 2.7), LS WETDEP is large scale wet deposition flux (Section 2.6.1), CNV WETDEP is convective wet deposition flux (Section 2.6.2)

<table>
<thead>
<tr>
<th>Run number</th>
<th>DRYDEP [%]</th>
<th>LS WETDEP [%]</th>
<th>CNV WETDEP [%]</th>
<th>BURDEN [Tg]</th>
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Figure 2. Importance of the different loss processes for simulation 1. The figures show annual mean production, dry deposition, large scale rainout and convective rainout. All fluxes are in kg m\(^{-2}\) s\(^{-1}\)
Table 3. Yearly deposition to oceans and forests (Tg). The following abbreviations are used: GOC: Global ocean, NATL: Northern Atlantic ocean, SATL: Southern Atlantic ocean, NPAC: Northern Pacific, SPAC: Southern Pacific, IND: Indian ocean, Persian Gulf and the Red sea, MED: Mediterranean and AMZ: Amazon forest

<table>
<thead>
<tr>
<th>Run number</th>
<th>GOC</th>
<th>NATL</th>
<th>SATL</th>
<th>NPAC</th>
<th>SPAC</th>
<th>IND</th>
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<td>9</td>
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</tbody>
</table>

Figure 3. Global and yearly averaged dust production simulated by Oslo CTM2. Upper left shows production in simulation 1. The other three figures are differences.
able: Using wind speed PDF facilitates emission from areas with low winds. Keeping the total emissions constant, the emission from high wind areas must decrease when wind PDF is taken into account.

4.4. Global loading

4.4.1. Yearly variations Figure 5 shows global mass column burdens. There is a maximum in the northern hemisphere summer over Sahara, and a maximum in the northern hemisphere spring in Asia. The Asian “spring dust” is a well known phenomenon. The plume out of Western Sahara peaks in July giving most transport across the Atlantic ocean in the northern hemisphere summer. Loadings over Australia are low, but peak in the southern hemisphere summer.

In figure 6 we give zonal mean mass mixing ratios of dust. The level which the dust is lifted to, is higher in the northern hemisphere summer. During the northern hemisphere summer, the deserts at 20° north are heated making air rise higher. In July, significant amounts of dust are lifted above the 200 hPa level, whereas in January, the dust is only lifted to approximately 600 hPa. The higher lifting is important for the interaction with terrestrial radiation. When dust is lifted high, it absorbs and re-emits long wave radiation at lower temperatures, giving a stronger long wave radiative effect than it would have at low levels.

4.4.2. Effect of erodibility In Figure 7 we show the effect of changing the soil erodibility on the annual burden. The results are summarized below. Not all of these observations can be evaluated because we do not have measurement data for all the regions, but we try to evaluate some of the observations in Section 4.6:

1. Both simulation 3 and 4 give higher dust loading over Asia (Taklamakan/Gobi and over the Caspian sea) than simulation 1 and 2.
2. Simulation 3 gives higher burdens in Southern Sahara due to very strong emissions close to Lake Chad. These emissions are transported south-west in our model.
3. In simulation 3 and 4, both the Lake Chad area and the Mali/Mauritania/Algeria have very high loadings.
4. Simulation 4 generally gives low burdens over Sahara.
5. Simulation 3 gives low Saudi-Arabian burdens
6. Simulation 4 gives high Arabian burdens, mainly originating from Somalia and South-western Saudi Arabia.
7. In Australia, simulation 4 gives more dust transported East of Australia instead of North-West as in simulation 1 and 2. This is due to higher emissions in the Lake Eyre basin than in the Great Sandy Desert area.

4.4.3. Effect of wind speed variability Figure 8 shows the effect of ignoring wind speed variability on the annual burden. This figure must be seen in connection with Figure 4. When ignoring wind speed variability, the increased emissions in Somalia and Australia increase burden over all the Indian ocean. Both these areas contribute to the increased deposition to the Indian ocean in simulation 5 (Table 3).

In the Saharan desert, the burden increases over coastal areas when ignoring wind speed variability. This is due to winds being high near the coast (section 4.3.2).

4.5. Effect of reducing washout

In figure 9 we show the annual mean absolute difference in burden for the runs 1 and 6. (In simulation 6 washout efficiency has been reduced). The results show that only washing out 30% of the dust increases dust burden significantly. As shown in Table 2, the annual mass loading increases by 55% (from 18.9 Tg to 29.3 Tg) when only 30% of the dust is washed out. Dust washout is hard to model since the dust can change hygroscopic properties during transport (Section 2.6).

Figure 9 shows that the effect of reducing washout is largest in the Atlantic ocean in the ITCZ. It can also be seen that the dust emitted from the Taklamakan Chinese desert is increased when washout is less efficient. Table 3 shows that deposition to the Pacific goes up when rainout is less effective. This higher deposition in the Pacific is due to increased transport from Asia to the Pacific due to the increased dust lifetime in simulation 6.

4.6. Observed Aerosol Index

Figure 10 shows the annual average Aerosol Index from the Total Ozone Mapping Spectrometer (TOMS) satellite [Torres et al., 1998] for the year 1980. Even though 1980 is not the year simulated, the picture reproduces many of the features from our different simulations.

TOMS data should be interpreted with care, because it does not only observe dust. TOMS observe all absorbing aerosols, and it detects black carbon and biomass aerosols in addition to dust aerosols.

Several observations can be made when comparing Figure 10 with Figure 7 and Figure 8:

1. TOMS indicates very high aerosol index in the Lake Chad area consistent with simulation 3.
2. TOMS indicates very high aerosol index in the Chinese Taklamakan desert, consistent with simulation 3 and 4. The Gobi desert seems like a much weaker dust source (inconsistent with simulation 3 and 4).
3. TOMS indicates low Somalian emission, consistent with simulation 1, 2 and 3.
4. TOMS indicates that Australian emission should mainly come from the Lake Eyre area, and not from the Great Sandy desert. This is consistent with simulation 3 and 4.
5. The high Caspian sea dust loading is found in TOMS and in all simulations.
Figure 4. Global and yearly averaged dust production simulated by Oslo CTM2. Left shows production in simulation 1. Right shows difference in production between simulation 1 and 5. The difference is due to neglecting wind speed variability.

Figure 5. Column burden mass in January (Upper left), April (Upper right), July (Lower left) and October (lower right) simulated by Oslo CTM2
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Figure 6. Zonal mean mass mixing ratio in January (Upper left), April (Upper right), July (Lower left) and October (lower right) simulated by Oslo CTM2

Figure 7. Effect of changing soil erodibility on column burden. Upper left is reference simulation (1). The other three are absolute difference in simulation 2-4
Figure 8. Effect of neglecting wind speed variability on column burden. Left is reference simulation. Right is difference in simulation 5.

Figure 9. Absolute difference in dust burden (kg/m²) for 1996. Left is Yearly averaged burden in simulation one, and right is increase in burden when washout is decreased.

Figure 10. Yearly average TOMS aerosol index for the year 1980.
4.7. Station comparisons

4.7.1. Monthly Mass concentrations Model output from CTM2 has been compared to mean mass concentrations at several stations. For some stations, we have concentrations from 1996 (Figure 11), and for the others, we only have climatological means (Figure 12).

Figure 11 shows that all the stations in the Atlantic plume are simulated reasonably. For the remote stations (Barbados, Bermuda and Miami), all the simulation with 100% rainout show similar results. When rainout is decreased, simulated concentrations in these stations are too high compared to measurements.

For the stations close to the source (Izana, Sal Island), the parameterization of rainout does not change the results much. This is logical since there is not much rainfall close to the desert sources. Dry deposition is the most important loss mechanism close to the deserts. Therefore, the different wet loss mechanisms are not so important when evaluating stations near to sources.

The only Asian station we have is Cheju. All the simulations show too low mixing ratios at this station. Simulations 3 and 4 is closest to reproducing the observed concentrations. Being situated south of Korea it should be dominated by dust from Gobi (and Taklamakan). This results suggests that our Asian sources are too weak in all simulations. The simulation with decreased washout (6) gives significantly higher concentrations than the simulation (1) with full washout, but not high enough according to the measurements.

For the stations where we only have climatological means, all simulations (except simulation 6) represent well the concentration in the remote Mace Head stations. For the remote Oahu station, the measurements are in between the model results with full washout (1-5) and the simulation with decreased washout (6). We suspect that the washout in simulation 6 is too low. Too low washout compensates for too low Asian emissions to give reasonable values for Oahu in simulation 6.

The Kaasidogo station is situated in the Ocean to the south of India. All simulations give reasonable concentration levels at this station. The fact that simulation 6 does not diverge much from simulation 1-5 indicates that this station is not far from a source area (the deserts in North western India).

4.7.2. Optical depths in the Western Saharan plume To check if the choice of soil erodibility had a large effect on the simulation of specific dust episodes, the optical depths observed at Cape Verde (in the western Saharan plume) have been compared to model derived optical depth on a daily basis. The choice of erodibility factors have a large effect on the output. Cape Verde was chosen because it is situated in the middle of the dust plume leaving the Saharan desert.

Table 4 gives correlation coefficients for the different model runs and the daily aerosol optical depth at Cape Verde. All the MODIS based erodibility datasets yield high correlation coefficients (larger than 0.70).

Figure 13 (left) shows the aerosol optical depth at Cape Verde for model run 1 and 6 which have the highest correlation coefficients at this location. It can be seen that both the magnitude and the timing of the events is good. It is surprising that simulation 6 gives the best correlation with measurements. Globally, this simulation gives a rather high dust loading compared to loadings generally proposed by other authors. It is possible that for a station so close to the source area, the low wet deposition efficiency is realistic because the dust has not had time to get coated with materials making it more hydrophilic (Section 2.6).

Figure 13 (right) shows the same plot with simulation 3 and 4. It can be seen that whenever there are peaks in simulation 3, the peaks are too large. In particular simulation 3 gives peaks which are much higher than the measured ones. Simulation 4 generally gives too low optical depth in periods when the optical depth is not peaking. Note that the scale in the right panel of Figure 13 is different from the scale in the left panel due to the large values of the peaks in simulation 3.

Simulation 3 shows significantly lower correlation than the other simulations at Cape Verde. This indicates that the method applied by Zender et al. [2003a] does not give the most appropriate erodibility dataset for western Sahara. The amplitude of the peaks in simulation 3, indicate that the erodibility datasets proposed by Zender et al. [2003b] is too heterogeneous, and that the differences in soil erodibility between the desert areas in the world are not as large as proposed by the GEO dataset.

In the different data sets, the erodibility factor can vary by an order of magnitude for a given area. We have compared several Saharan dust outbreaks to TOMS aerosol index (not shown). The results show that using different erodibility factors can replace the outbreaks. (That is: The outbreaks will come from different areas depending on the erodibility dataset applied). This means that even though on average, the plume will originate from approximately the right area (e.g. Western Sahara), each individual outbreak can originate from an area only close to the correct source area. Campaign studying individual Saharan dust outbreaks can hopefully help determine the exact areas with the most erodible soils.

The good correlations obtained by the MDSLNR and
Figure 11. Mass concentrations from U. Miami network (for 1996) compared to CTM2 concentrations (for 1996). The figures represent Barbados (301E, 13N), Bermuda (295E, 32N), Miami (280N, 25E), Izana (17W, 28N) and Sal Island (25W, 15N). Black solid line is measurements. Red dashed is simulation 1, blue dashed is simulation 2, yellow dashed is simulation 3, cyan dashed is simulation 4, green stars is simulation 5, and blue stars is simulation 6.
Figure 12. Mass concentrations form U. Miami network (climatological means) compared to CTM2 concentrations (for 1996). The figures represent Cheju (126E, 33N), Kaasidogo (73E, 5N), Mace Head (10E, 53N) and Oahu (202E, 21N). Black solid line is measurements. Red dashed is simulation 1, blue dashed is simulation 2, yellow dashed is simulation 3, cyan dashed is simulation 4, green stars is simulation 5, and blue stars is simulation 6.

Figure 13. Aerosol optical depth at Cape Verde. Left figure is simulation 1 and 6 compared to measurements. Right figure is simulation 3 and 4 compared to measurements. Note the difference in scale between left and right panel.
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MDSSQR are encouraging given the simplicity of these datasets. We believe there is connection between soil erodibility and soil reflectivity. Both MDSSQR and MDSLNR are only simple concepts which can probably be refined to give better datasets for soil erodibility.

5. Summary

Several different parameterizations of dust production and transport have been tested and evaluated. We have focused on the parameters “soil erodibility” and wind speed variability. Both these influence the magnitude and geographical location of the dust emissions.

We show that the geographical location and magnitude of atmospheric dust can vary significantly when changing soil erodibility datasets. The emissions are highest in the areas where the erodibility is highest.

When ignoring wind speed variability, the geographical pattern of dust emissions is changed. Ignoring wind speed variability (and fixing total dust emissions) gives high production in areas which have high wind (e.g. areas close to the coast).

The datasets used here to describe soil erodibility have different weaknesses and strengths. We show that using the soil erodibility dataset of Zender et al. [2003b] reproduces several typical aspects of the global dust cycle (high Chinese dust, high loading in Lake Chad area, higher emissions in Taklamakan than in Gobi). However it fails reproduce other aspects such as Lake Eyre (and not the great sandy desert) being the dominant Australian source. The dataset proposed by Ginoux et al. [2001] gives high Somalian emissions. This is not reproduced by any other model run.

The reflectivity based erodibility factors give low emissions in the Chinese deserts, however they capture the main Saharan regions such as north eastern Sahara (Egypt/Libya), western Sahara (Mali/Mauritania/Algeria) and Lake Chad.

Some aspects of the annual mineral dust cycle, such as maximum dust transport from Sahara to the Caribbean in the northern hemisphere summer, maximum dust production in Australia in the southern hemisphere summer and maximum dust production in the Chinese deserts in spring is reproduced by all model simulations. Monthly average concentrations at remote stations in the Saharan plume are similar in all simulations except when wet loss of dust is significantly reduced.

When comparing aerosol optical depths at Cape Verde day by day we show that assuming that soil reflectivity can represent soil erodibility is a good assumption which represents the measured optical depth with a correlation coefficient higher than 0.70. The other erodibility give lower correlation coefficients, and the optical depths are too high in many of the dust episodes.

The good correlations obtained by the MDSLNR and MDSSQR are encouraging given the simplicity of these datasets. We believe there is connection between soil erodibility and soil reflectivity. Both MDSSQR and MDSLNR are only simple concepts which can probably be refined to give better datasets for soil erodibility.

Acknowledgments. AG, GM, JKS and ISAI acknowledge support from the Norwegian research council grant 139810/720 (CHEMCLIM). CSZ acknowledges support from NASA grants NAG5-10147 (IDS) and NAG5-10546 (NIP)
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This preprint was prepared with AGU’s LATEX macros v4, with the extension package ‘AGU++’ by P. W. Daly, version 1.6b from 1999/08/19.