Modeling the radiative impact of mineral dust during the Saharan Dust Experiment (SHADE) campaign

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[1] The Oslo chemical transport model (Oslo CTM2) is driven by meteorological data to model mineral dust during the Saharan Dust Experiment (SHADE) campaign in September 2000. Model calculations of the optical properties and radiative transfer codes are used to assess the direct radiative impact in the solar and terrestrial regions of the spectrum. The model calculations are compared to a wide range of measurements (satellite, ground-based, and aircraft) during the campaign. The model reproduces the main features during the SHADE campaign, including a large mineral dust storm. The optical properties and the vertical profiles are in reasonable agreement with the measurements. There is a very good agreement between the modeled radiative impact and observations. The strongest local solar radiative impact we model is around $-115 \text{ W m}^{-2}$. On a global scale the radiative effect of mineral dust from Sahara exerts a significant negative net radiative effect.

INDEX TERMS: Radiative processes; KEYWORDS: aerosols, single scattering albedo, transport model, aircraft measurements


1. Introduction

[2] Among the major components contributing to the direct aerosol effect mineral dust is the most uncertain. Even the sign of the radiative forcing due to mineral dust is unresolved [Intergovernmental Panel on Climate Change (IPCC), 2001]. The main uncertain factors are the degree of absorption of solar radiation (an important part is the refractive index), the influence of mineral dust on longwave radiation, the abundance of mineral dust in the atmosphere, and in particular the human influence on this abundance.

[3] Mineral dust has a cooling effect over ocean in the absence of clouds. Haywood et al. [2001] measured a strong radiative impact (or radiative effect) during a dust event. However, over brighter surfaces (as over desert and above clouds) the radiative impact will be strongly altered. Haywood and Shine [1995] and Hansen et al. [1997] showed that single scattering albedo is a critical factor in determining the sign of the solar radiative impact for various surface albedos. The longwave radiative impact of mineral dust will always be positive, but the magnitude depends strongly on the size of the particles, the refractive indices and the altitude. Therefore the net radiative impact (sum of solar and longwave) exhibits large regional variations, and this explains why the global mean is difficult to estimate. Several studies have investigated the radiative impact of mineral dust [Tegen et al., 1996; Sokolik and Toon, 1996; Myhre and Stordal, 2001; Woodward, 2001; Weaver et al., 2002] with a wide range in the results.

[4] Tegen et al. [1996] and Sokolik and Toon [1996] estimated that the human contribution to the mineral dust abundance can be as large as 30–50%. These numbers are highly uncertain and it is an unresolved question whether this is a direct climate forcing mechanism or a climate feedback (see discussion by Myhre and Stordal [2001]).

[5] The Saharan Dust Experiment (SHADE) took place in September 2000 on the west coast of Africa to improve knowledge about the radiative impact of mineral dust using a wide range of measurements [Tanré et al., 2003]. This study uses meteorological data to model the production and transport of mineral dust during the campaign and compares the results against various in-situ and remotely sensed airborne and surface-based measurements and calculates the direct radiative impact for a larger region surrounding the Sahara.

2. Models

2.1. Transport Model

[6] Oslo CTM2 is an off-line chemical transport/tracer model (CTM) that uses precalculated transport and physical...
fields to simulate tracer distributions in the atmosphere. The model represents the global troposphere and is three-dimensional with the model domain reaching from the ground up to 10 hPa for the current data sets. In the horizontal and vertical the model has a variable resolution that is determined by the input data provided. In this study we use forecast data from the European Center for Medium-Range Weather forecast (ECMWF) with $1.875^\circ \times 1.875^\circ$ degrees horizontal resolution and with 40 levels in the vertical with a resolution of about 600 meters in most of the troposphere.

The second-order moment method is used for advection [Prather, 1986]. Convection is based on the Tiedtke mass flux scheme [Tiedtke, 1989; Sundet, 1997], where vertical transport of species is determined by the surplus/deficit of mass flux in a column. Dry and wet deposition are treated as in the work of Grini et al. [2002a].

The forecast data are constructed from merged forecasts where each forecast is run for 36 hours with the first 12 hours discarded as “spin-up.” The data are stored at three-hour intervals for August and September 2000. Most of the data that are diagnosed from the ECMWF model are standard model output (cloud cover, surface properties etc.) and is thus the state of the art parameterizations of the relevant processes. Alas, since the CTM is off-line there will be no feedback from the CTM to the data that are diagnosed in the ECMWF model. Data that are not part of the ECMWF diagnostic (convective fluxes, rain fall) have been diagnosed explicitly with the built-in diagnostic system in the ECMWF model, and the physical properties of the fields are described by Tiedtke [1987].

2.2. Dust Production

Production of dust is parameterized using the formula

$$\text{production} = \text{const} \cdot u^*_e \left( u_e - u_{\ast, p} \right),$$

where production is in kg m$^{-2}$s$^{-1}$ and $u_e$ is friction velocity (see definition in the work of Stull [1988]) in m/s. The constant used is $1.7 \times 10^{-5}$ kg m$^{-2}$s$^{-5}$. The threshold friction wind speed ($u_{\ast, p}$) is the minimum friction wind assumed to give dust emission.

Different landscapes have different threshold winds for dust emissions, normally between 6 m/s and 20 m/s [see, e.g., Chomette et al., 1999]. We have used global data sets to describe the vegetation. An average procedure was developed using the data set of Olson et al. [1983], Matthews [1983], Wilson and Henderson-Sellers [1985], Ramankutty and Foley [1999], and SARB http://www-surf.larc.nasa.gov/surf/). The lowest threshold wind of 6 m/s was used where all data sets agreed on pure sand deserts. The highest threshold wind of 20 m/s was used in areas where all data sets agreed on shrub. To convert threshold wind ($U_{\ast, p}$) speeds to threshold friction wind speeds, we used the formula $u_{\ast, p} = 0.035 \cdot U_{\ast, p}$ (based on the work of Nickling and Gillies [1987, Table 3]). The constant used in equation (1) gives the same production as in the work of Tegen and Fung [1994] for a wind speed of 10 m/s and a threshold wind speed of 6.5 m/s.

For areas with large roughness elements, the friction wind velocity is very high as the friction velocity increases with roughness length. In arid areas with large roughness, for instance over the Moroccan Atlas mountains, the ECMWF data have a very large friction wind speed. In such areas, we use the formula production $= C u^2 \left( U - U_{\ast, p} \right)$ with C equal to $0.7 \mu g m^{-2} s^{-5}$ as proposed by Tegen and Fung [1994]. For areas with large roughness elements, large wind friction speeds reflect large momentum flux to the surface because of these elements. It will not reflect the momentum appropriate for dust production (we assume that the dust production takes place at smooth surfaces, and it should thus not be influenced by large scale roughness elements). In our model, the threshold wind speeds do not change with surface roughness, but only with vegetation cover. For a further discussion on use of wind or friction wind in dust modeling, see Liu and Westphal [2001].

The soil moisture makes the soil more difficult to erode as the water increases the interparticle capillary forces in the soil. Fecan et al. [1999] proposed relationships to deal with the increase in threshold wind speed with increasing soil water content. The soil moisture from the ECMWF model is not believed to be realistic enough to be used to calculate the increase in threshold wind speed because in the ECMWF model there is no evaporation from soils with humidity lower than a prescribed, globally uniform permanent wilting point of 0.171 m$^3$/m$^3$. This value is much larger than the values discussed by Fecan et al. [1999] (P. Viterbo, personal communication, 2001). To deal with the problem, we chose to set dust emission to zero for the next two days whenever rainfall is higher than 2 mm in 24 hours in a grid square, similar to the approach of Claquin [1999]. Our simple dust production formulation is a source for uncertainty in the model, but this formulation will capture the largest dust outbreaks.

2.3. Dust Optical Properties

The specific extinction coefficient, the single scattering albedo, and the asymmetry factor are modeled using Mie scattering theory. In Mie scattering theory spherical particles are assumed and particle size and refractive index need to be specified. Haywood et al. [2003] show that the mineral dust aerosols are not spherical, but Mishchenko et al.
show that this simplification introduces only a modest uncertainty in calculation of radiative fluxes. The size distribution is explicitly modeled in the transport model and adopted in the optical and radiative calculations. The refractive indices for mineral dust are highlighted as the single most important factor for the large uncertainty in the radiative impact of mineral dust [Myhre and Stordal, 2001]. Kaufman et al. [2001] show that the single scattering albedo at solar wavelengths is much higher and that the imaginary part of the refractive index is substantially lower than in previous studies. We have chosen to use refractive indices in the solar spectrum from the Dubovik retrieval from the Sun photometer at Cape Verde. Real refractive index of 1.48, 1.47, 1.43, at 1.42 at 440, 670, 870, and 1020 nm, respectively and imaginary refractive index of 0.0014, 0.0011, 0.0011, and 0.0011 at 440, 670, 870, and 1020 nm, respectively are adopted. In the thermal infrared spectrum we use refractive indexes from Fougquist et al. [1987] as these compare most favorably with the detailed measurements and modeling presented by Highwood et al. [2003]. The density of the mineral aerosols, used in the calculations of the specific extinction coefficient, are taken to be 2600 kg m\(^{-3}\) [Hess et al., 1998].

2.4. Radiative Transfer Model

We apply separate longwave and shortwave multi-stream models in the radiative transfer calculations using the discrete-ordinate method [Stamnes et al., 1988]. In the calculations in this study eight streams are used. The solar radiative transfer model includes radiative effects of aerosols, clouds, Rayleigh scattering, and the exponential sum fitting method [Wiscombe and Evans, 1977] is used to account for absorption by gases. There are 4 bands in the solar spectrum (see Myhre et al. [2003] for further details). The longwave scheme takes into account absorption by gases and clouds in addition to scattering and absorption by aerosols. The exponential sum fitting method [Wiscombe and Evans, 1977] is used to account for absorption by water vapor, carbon dioxide, and ozone. Five spectral regions are used in the longwave spectrum (see Myhre and Stordal [2001] for further details).

Our calculations are performed in T63 spatial resolution and 40 vertical layers. The meteorological data of temperature, water vapor, clouds used in the radiative transfer calculations during the campaign period are described in section 2.1. Within each model grid column we perform separate radiative transfer calculations in clear sky and cloudy regions to establish an overall radiative flux for the column. Hogan and Illingworth [2000] use radar observations to show that the random cloud overlap is a satisfactory assumption to estimate the total cloud cover. In this model version with 40 vertical layers we combine the random cloud overlap (to describe effect of clouds in different height regions) and the maximum cloud overlap assumption (for nearby levels) to estimate a realistic total cloud cover. The optical properties of clouds in the solar region are calculated using the procedure described in the work of Slingo [1989] with effective radius of 10 \(\mu\)m for low clouds and 18 \(\mu\)m for high clouds [Stephens, 1978; Stephens and Platt, 1987]. In the thermal infrared spectrum non-scattering clouds are assumed with absorption coefficients dependent on the altitude of the clouds [see Myhre and Stordal, 2001]. There is no interaction between the mineral dust aerosols and the cloud microphysics in the model.

The surface albedo is modeled as a function of solar zenith angle and solar spectral region. Over ocean the surface albedo dependence on solar zenith angle is modeled according to Glew et al. [2003], whereas over land as in the work of Briegleb et al. [1986]. Surface albedo over land is based on vegetation data from Ramankutty and Foley [1999] with spectral albedo values from Briegleb et al. [1986].

We have performed radiative transfer calculations including and excluding mineral dust. The difference between these two simulations at the top of the atmosphere is taken as the radiative impact of the aerosols. This is similar to the radiative forcing concept, except that we include both anthropogenic and natural abundance of the aerosols.

3. Results

The purpose of our calculations is to compare our results with intensive observations during the SHADE period 19 to 28 September 2000. We first calculate the dust distribution using the CTM to simulate production of dust, the transport, and deposit of the dust particles. These calculations are started 1 August, in order to spin up the model. Thereafter we calculate optical properties and radiative impact of the dust during the campaign period.

3.1. Production

Figure 1 shows the modeled production of mineral dust during the SHADE campaign. A substantial diurnal variation can be seen, with much larger production during the day. This indicates that solar heating is an important factor in the generation of the surface winds. Further, the model indicates an important shift in the source regions during the campaign. During the first part of the campaign the source regions seem to be closer to the coast compared to the main dust storm during the period 22–24 September 2000. This is due to a shift in the region with friction velocity above the threshold value. The threshold friction velocity is somewhat higher at the coast than in the main source region during the campaign, affecting the magnitude of the production. Based on the model, the mineral dust transported over the campaign area mainly has its origin in the western part of Sahara.

3.2. Modeled Aerosol Optical Depth

The modeled aerosol optical depth at 550 nm during the campaign is shown in Figure 2a. It clearly reveals the large dust storm starting in western Sahara the 22nd and developing over the next days and with a westward movement of the mineral dust plume. Over the main campaign area near Sal Island and Dakar it shows that AOD is moderate during the 2 or 3 first days. Thereafter there are some days with low AOD before the dust starts to move over ocean the 24th. A maximum AOD over the campaign area can be seen the 26th. The AOD weakened as the mineral dust move westward, but is still not insignificant as the dust plumes cross the Atlantic ocean in accordance with earlier measurements.

3.3. Comparison With MODIS Aerosol Optical Depth

In Figure 2b the AOD from the Moderate Resolution Imaging Spectroradiometer (MODIS) [Kaufman et al.,
aboard the Terra satellite is shown for 25–26 September 2000. The pattern with high AOD along the coast 25 September and the mineral dust transported further out over the ocean is clearly evident and rather similar to the model. Note in particular the similar shape of the dust plume for 26 September in the satellite retrieval and the model results. However, it seems that the plume is slightly displaced eastward in the model compared to the satellite retrievals. Further the maximum AOD is somewhat higher in the satellite retrievals than in the model, which can partly be explained by the much higher horizontal resolution in the satellite data. The AOD values from MODIS compare well with AERONET observations over ocean [Remer et al., 2002]

3.4. Comparison With AERONET Aerosol Optical Depth

Aerosol optical depths from the model are compared to surface Sun photometers at Cape Verde and Dakar during the campaign period and shown in Figure 3. These instruments are part of the AERONET which is a global ground-based network of Sun photometers (see Holben et al. [1998] for a description). The general pattern which can be divided in four periods is in reasonable agreement given the large
Figure 2. AOD at 550 nm. (a) Modeled during the period 19–28 September 2000 at 0 and 12 UTC. (b) MODIS data for 25–26 September between 9 and 14 UTC. Note that both panels have the same color scale. Cape Verde is located at 16.73N and 22.94W, and Dakar is at 14.39N and 16.96W.
uncertainties in the source function of mineral dust. In the beginning of the campaign there was moderate AODs, thereafter much lower values, followed by large AOD with maximum values above 1, and finally more moderate values. The modeled AOD is lower than the maximum observed values, in particular at Cape Verde. Further higher modeled AOD is indicated during the period from 21st to 23st with very low observed values.

Another feature is the stronger wavelength dependence of AOD in the model compared to the observations. This is most pronounced during high AOD conditions. The model indicates that in the beginning of the campaign period there was a different main source region compared to in the period with heaviest dust loading, which may be the reason why the observed size distribution changes during the campaign.

The modeled size distributions compare relatively well with the observed distributions for particles below 1.0 \( \mu m \), but the model clearly underestimates the fraction of larger aerosols. We have tried replacing the size distribution we have adopted in the source region with an alternative distribution \( [\text{Schultz et al., 1998}] \) (a three mode lognormal distribution \( D_{\text{mode}} = 0.011, 2.52, \) and 42.3 \( \mu m \), respectively and \( \sigma = 2.13, 2.00, \) and 1.89, respectively) without significant change in the number of large particles.

### 3.5.2. Vertical Profiles

Figure 5 shows the vertical profile of scattering, single scattering albedo, and asymmetry factor. A comparison is made between the C-130 aircraft observations in the main dust layer and the model for two locations during the a797 flight (see Haywood et al. [2003] for a description of the flight). Figures 5a and 5d show the vertical profiles of aerosol scattering at 550 nm from the model compared...
against those from the nephelometer on board the C-130 aircraft. The comparisons are made near Cape Verde and near Dakar during the a797 flight on the 25 September 2000 (see Haywood et al. [2003] for a full description of this flight). The C-130 derived scattering coefficient includes corrections for variations from STP, truncation of the scattered radiation, and for deficiencies in the illumination source [Anderson and Ogren, 1998]. An additional correction factor of 1.5 is applied to account for the significant contribution to the aerosol scattering from supermicron particles that are not sampled efficiently by the inlet system [Haywood et al., 2003]. Figures 5b and 5e show comparisons of the single scattering albedo at 550 nm from the model against those derived from the C-130. In deriving the single scattering albedo from the C-130, vertically resolved 1-min averages of the aerosol size distribution from the PCASP and identical refractive indices to those used in the model are assumed in Mie scattering calculations. This method is preferred to determining the single scattering albedo from measurements of particle scattering by the nephelometer and particle scattering by the Particle Absorption Soot Photometer (PSAP), because it is difficult to obtain representative values from two vertical profiles owing to variability. The refractive index used in this study is derived from campaign averages of scattering and absorption. An additional benefit from using the PCASP size distribution to determine the optical parameters is that it provides the asymmetry factor which is compared in Figures 5c and 5f.

The profiles of the scattering show that the mineral dust is between 1 and 5 km, but with somewhat different distribution in the model and the observations. The observations show more distinct peaks, which the model with much coarser spatial resolution does not reproduce. The modeled single scattering albedo and asymmetry factor both show substantial vertical variations. The modeled single scattering albedo increases with altitude and the asymmetry factor decreases with altitude, which both indicate smaller particles in the upper part of the dust plume. For the observed single scattering albedo and asymmetry factor no such systematic changes with altitude can be found. In comparison with the observations the modeled single scattering albedo and asymmetry factor are higher and lower, respectively, in both cases mainly because the modeled particles are slightly smaller than the observed ones. The differences between the modeled and measured values are relatively small, and the modeled single scattering albedo shown in this figure is within the range of averaged measured single scattering albedo for the C-130 flights during SHADE (range from 0.95 to 0.98 and in the transit from Ascension Island it was 0.99). Actually, the measured single scattering albedo during this particular flight (a797) was the lowest of the 6 flights [Haywood et al., 2003].

In Figure 6 vertical profiles of scattering for the entire flight a797 are shown. The observed averaged profile shows a rather homogeneous scattering from 1 to 4–5 km, whereas in the model there is a more clear maximum around 3 km. The observed standard deviations are largest in the lower part of the dust plume, in particular around 1 km. However, in the model the largest standard deviations are in the middle and upper part of the dust plume. Overall the
agreement in the vertical profile is good, with the vertical position of the top and the bottom of the dust plume being well represented as well as the absolute magnitude of the scattering when an entire flight of scattering data is included.

3.5.3. Radiative Measurements

[31] To allow for an easier comparison of the radiative impact of mineral dust between the measurements and the model we introduce a normalized radiative impact (radiative impact divided by AOD). Clouds are excluded in the radiative transfer calculations also for a better comparison with the measurements which also are performed for clear sky. The model results are shown for the region 16.9–22.5°W and 13.1–16.9°N.

[32] Figure 7 shows that the modeled normalized radiative impact is between around 80 to 120 Wm\(^{-2}\) during the day. The normalized radiative impact is stronger at 9 and 18 UTC than closer to noon (local noon deviates from UTC noon by 1–2 hours in the campaign region) as the magnitude of the radiative impact of non-absorbing particles is highest for large solar zenith angles ([see Haywood and Shine, 1997]. This effect is almost compensated by surface albedo of the ocean being higher for large solar zenith angles. The day-to-day variation in the normalized radiative impact is small at 12 and 15 UTC; however, at 9 and 18 UTC the day-to-day variation is significant. This day-to-day variation is mostly caused by changes in AOD with weakest normalized radiative impact for high AOD. Furthermore, the variation is also slightly related to the size distribution.

[33] The modeled and measured normalized radiative impact are relatively similar during the middle of the day and this is an important result. In the model the values are around 80 Wm\(^{-2}\), whereas around 90 Wm\(^{-2}\) in the measurements. The normalized radiative impact depends slightly on the AOD, with decreasing normalized radiative impact with increasing AOD. Results from the measurements are only shown for small solar zenith angles as the uncertainties in the measurements increase significantly for larger solar zenith angles owing to uncertainties in the surface albedo of ocean for large solar zenith angles. The aircraft measurements of the radiative impact compare well with satellite derived radiative impact from the CERES instrument [Haywood et al., 2003].

3.6. Radiative Impact

[34] Above normalized radiative impact is discussed, below solar and longwave radiative impact without normalization is presented.

3.6.1. Solar

[35] The solar radiative impact of mineral dust in the period 24–26 September is shown for clear sky and when clouds are included in Figure 8. The radiative impact follows mainly the pattern of AOD. Clouds reduce the radiative impact of mineral aerosols. This is most pronounced west of 30°W. The radiative impact shows strong negative values, with some small positive values associated with very limited cloudy regions. The maximum radiative impact is 6 Wm\(^{-2}\). Mineral dust influences the solar radiation strongly not only near the source regions and its associate regions, but we can see a significant influence even in the western part of the Atlantic ocean. The strongest values reach almost 115 Wm\(^{-2}\). This is in accordance with measurements in the work of Haywood et al. [2003], who reported values ranging down to nearly 130 Wm\(^{-2}\). The strongest radiative impact is rather similar over land (24 September) and over ocean (26 September), despite the AOD is stronger 24 September. This is due to the fact that the normalized radiative impact is at least a factor two stronger over ocean than over land owing to the difference in surface albedo.

[36] In the work of Myhre et al. [2003] it was found that neglect of spectral and solar zenith angle variations in the surface albedo weakened the radiative impact of aerosols from biomass burning by 25–30% over the region which is

Figure 6. Vertical profiles of scattering for the C-130 observation and the model for flight a797. Standard deviations are shown as horizontal lines.

Figure 7. Modeled and observed normalized clear sky radiative impact (given as Wm\(^{-2}\) per unit AOD) for various times during the day. Results from the model are shown in the period 19–28 September 2000. The measurements are shown from a797 (R6; see Haywood et al. [2003]) on 25 September with about 5 min averages. Measurements were made at the top of the dust layer (about 5 km) and the model results at the top of the atmosphere (consistent with the rest of the model results). In the model the difference in radiative impact of mineral dust is negligible between 5 km and the top of the atmosphere.
Influenced by aerosols from biomass burning in southern Africa. For mineral dust, the influence of spectral and solar zenith angle variation is much smaller (5–10%). The reduction due to neglect of such variations is slightly larger over land than over ocean. The weaker response to surface albedo parameterization for mineral dust compared to biomass burning aerosols is due to the much stronger spectral dependence of specific extinction coefficient and single scattering albedo than for biomass aerosols.

In the work of Myhre and Stordal [2001] a large number of sensitivity calculations of the radiative forcing due to mineral dust were performed, pointing to refractive indices as a major source to the large uncertainty. In addition to the calculations performed using the refractive indices derived from Sun photometers, we also performed calculations using the refractive indices of d’Almeida et al. [1991] which is the base refractive index used in the calculations of Myhre and Stordal [2001]. Due to the much higher imaginary part of these data for the refractive index compared to the one based on the Sun photometers this yields a reduction in the single scattering albedo at solar wavelengths; see Haywood et al. [2003] for discussion. The radiative impact over the region of interest decreases by about 25% for clear sky and as much as 50% when clouds are included. The strongest radiative impact is weakened from around $-115 \text{ Wm}^{-2}$ to around $-80 \text{ Wm}^{-2}$. Positive radiative impact of up to $50 \text{ Wm}^{-2}$ was then estimated, as opposed to $6 \text{ Wm}^{-2}$ in the original calculation. The increase in the radiative impact was much larger over land than over ocean owing to the brighter surface.

Figure 8 shows modeled diurnal mean solar radiative impact when clouds are excluded and included in the radiative transfer calculations over the region shown in Figure 9. The radiative impact is somewhat stronger during the end of the campaign period than in the beginning. The figure shows that the radiative impact for clear sky is 40–50% stronger than when clouds are taken into account. When clouds are included in the simulations the solar radiative impact is around $-6 \text{ Wm}^{-2}$ over the region, corresponding to a global mean radiative impact of $-0.4 \text{ Wm}^{-2}$ (scaled according to the fraction of the area of this region relative to Earth’s surface). Note that these numbers take into account almost the entire radiative effect of mineral dust particles originating from Sahara, but not from other sources.

### 3.6.2. Longwave

[39] The diurnal mean longwave radiative impact is shown in Figure 10a. The longwave radiative impact is much smaller in magnitude than the solar part in our simulations. The radiative impact is strongest over land near the source regions where the AOD is highest, but in addition the surface temperature is higher and the troposphere is less humid compared to over ocean contributing to an even larger land/ocean contrast in the results. The maximum diurnal mean longwave radiative impact is close to 8 Wm$^{-2}$. Figure 4 indicates that the model underestimates the number of larger mineral particles. This causes that the longwave radiative impact is underestimated in our calculations. In a calculation using the lognormal size distribution based on the observations shown in Figure 4 resulted in 10–20% stronger longwave radiative impact.

A sensitivity calculation with the refractive index from d’Almeida et al. [1991] instead of from Fouquart et al. [1987] is performed. This resulted in a weakening of the longwave radiative impact by a factor more than 2. The optical properties changes substantially with the change in refractive indices [see Highwood et al., 2003].

[40] Figure 11 shows the regional mean longwave radiative impact during the campaign (region as shown in Figures 9 and 10. As in Figure 8 we see that the radiative impact is stronger during the end of the campaign. Clouds weakened the radiative impact by about 20%. The regional mean longwave radiative impact varies from around 0.8 to slightly more than 1.0 Wm$^{-2}$ and is in magnitude about a factor 6–7 weaker than the solar radiative impact.

### 3.6.3. Net Radiative Impact

[41] Figure 10b shows the diurnal mean net radiative impact of mineral aerosols during the SHADE campaign period. The radiative impact is strongest on 26 September with diurnal mean values close to $-50 \text{ Wm}^{-2}$ off the coast of western Africa. The small positive values of the net radiative impact is associated with clouds. The ratio of the magnitude of the longwave and solar radiative impact is generally a factor of 2 higher over land than ocean. The diurnal and regional mean net radiative impact of mineral aerosols is slightly stronger than $-5 \text{ Wm}^{-2}$ when clouds are included in the simulations.

### 4. Summary

[42] In an attempt to reduce the large uncertainties linked to the direct effect of the mineral dust, we model the radiative impact of the mineral dust with a comprehensive comparison with measurements obtained during the SHADE campaign. We use meteorological data for the campaign period, which enables us to reproduce the main pattern of mineral dust during SHADE. The general agreement with observations is very good. Satellite retrievals show that the general pattern of mineral dust during the campaign is reproduced. MODIS data is used during the major dust storm, and also TOMS aerosol index (not shown) reveals a very similar feature to the model during
the entire campaign, even the transport of mineral dust to the north of Sahara. Comparison with ground-based Sun photometers strengthens the agreement of the main features. In addition the possibility to compare the model against in situ aircraft data is very valuable. Most important in this respect is the radiative effect of aerosols, which is of large scientific value. Given the fact that the model is completely independent of the measurements, a radiative effect in the model and observations within about 10% is very encouraging, demonstrating the usefulness of larger aerosol campaigns. Both in the model and observations a weak dependence of the normalized radiative impact on AOD is found close to noon. The modeled size and vertical profiles are compared with the aircraft measurements with reasonable agreement, except for the larger sized mineral particles which are underestimated in the model. This affects most strongly the longwave radiative impact results, and only to a small extent the shortwave results.

Figure 9. Radiative impact of mineral dust in the period 24–26 September for (Figures 9a–9d and 9i–9l) four times each day, and (Figures 9q–9t) clouds are included in the calculations, whereas in Figures 9e–9h, 9m–9p, and 9u–9x, clouds are excluded.
The problem with the size distribution is a result of the currently limited knowledge of the physical processes in the source of mineral dust. Further, few observations in these areas are available. The physical uncertainties of the source are not only linked to dependence on friction velocity and surface composition, but also on data characterizing the surface. A problem for a global model is also to represent physical processes that occur on a small scale. A major result of the SHADE campaign is that it confirms the recent findings in the work of Kaufman et al. [2001], who reported a much higher single scattering albedo than earlier measurements indicate [Haywood et al., 2003]. This can either be due to the fact that in earlier measurements mineral dust has been mixed with aerosols from biomass burning or that the source regions are located in other regions than during SHADE. Clarke et al. [2001] find a similar high single scattering albedo for dust over China, but on the other hand Claquin et al. [1999] show that the mineral composition varies substantially in Sahara and the absorbing hematite has a much higher abundance in the Sahel region than in the rest of Sahara.

We estimate a much stronger solar radiative impact than the longwave radiative impact, and thus a strong net radiative impact. Figure 10 shows the diurnal mean radiative impact of mineral dust during the campaign period 19–28 September. (a) Longwave. (b) Net.
negative radiative impact or cooling effect of the mineral dust from Sahara during the SHADE campaign. The strongest modeled net radiative impact is around $-110 \text{ Wm}^{-2}$, occurring around local noon on 26 September. Over a region (latitude 0°N–30°N, longitude 60°W–40°E) around Sahara the diurnal mean radiative impact during the SHADE campaign is between −5 and −6 Wm$^{-2}$. This translates to a global mean net radiative impact of Saharan dust aerosol of around $-0.4 \text{ Wm}^{-2}$ when averaged over the globe. The solar and longwave radiative impact results are based on refractive indices which are from observations or compared closely to measurements. In accordance with earlier studies we find a strong sensitivity to the refractive indices and the use of refractive indices from commonly used sources resulted in solar and longwave radiative impact a factor of 2 weaker than our best estimates.

The diurnal and regional mean of the solar radiative impact is between $-8$ and $-10 \text{ Wm}^{-2}$ when clouds are excluded. This is somewhat stronger than the global mean due to all kinds of aerosols, as found in several studies based on satellite data [Haywood et al., 1999; Boucher and Tanré, 2000; Loeb and Kato, 2002]. Our results yield stronger radiative impact than the ones derived globally from observations, as we are focusing on a region with large contributions from strongly reflecting dust aerosols. However, also large areas in the regions considered in our study are only weakly influenced by dust aerosols.

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