The Effect of Measured Surface Albedo on Modeled Saharan Dust Solar Radiative Forcing

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Abstract

The clear-sky solar radiative forcing of Saharan dust is computed for a case study during the SAharan Mineral DUst ExperiMent (SAMUM) in May 2006. Size-resolved dust concentrations simulated with a regional model and spectrally resolved surface albedo measurements were used as input for a 1D radiative transfer model to study the dependence of the dust radiative forcing at solar wavelengths on surface albedo and particle optical properties. Within the considered parameter range the surface albedo can have a larger impact on the solar radiative forcing of dust at the top of atmosphere (TOA) than the variations of optical properties. At the location of Ouarzazate in Morocco different measured surface albedo values lead to differences in instantaneous solar TOA solar forcing of up to 15 W m\textsuperscript{-2} for identical dust properties. This highlights the importance of using an accurate characterization of surface albedo values for estimating solar dust forcing over land. In the regional average over the Sahara using either the standard model values or satellite-based surface albedos leads to differences in the order of 9 W m\textsuperscript{-2} in the instantaneous solar forcing at TOA, and 5 W m\textsuperscript{-2} for the diurnal mean TOA forcing.
Atmospheric aerosols significantly impact the climate system by influencing the energy balance of solar and thermal infrared radiation (Solomon et al., 2007). Mineral dust aerosol particles emitted from desert surfaces that are transported over large distances are a significant contributor to the global atmospheric aerosol load. Mineral dust impacts the Earth’s radiation balance by scattering and absorbing incoming solar radiation, and by absorbing and emitting radiation in the terrestrial wavelength range. The net solar radiative forcing by dust is defined as the difference in net irradiance (downward minus upward irradiance) of the realistic atmosphere containing dust particles and a hypothetic clear atmosphere without dust. It depends on the optical properties of the particles, which in turn depend on the size distribution and wavelength-dependent complex refractive index of the particles. The magnitudes and even the sign of the net solar radiative forcing at the top of atmosphere (TOA) by dust are highly uncertain. In addition to its optical properties, the magnitude of dust net solar radiative forcing depends on, e.g., the presence of clouds, the vertical distribution of the dust, and on the albedo of the underlying surface (Claquin et al., 1998; Liao and Seinfeld 1998; Tegen and Lacis, 1996). In addition, the irregular shapes of dust particles lead to misrepresentations of their optical properties if their radiative effects are computed by Mie-calculations assuming spherical shapes (e.g. Kalashnikova and Sokolik, 2004). In the global annual average dust net solar radiative forcing at TOA is most likely negative (Solomon et al., 2007), but regionally positive values may occur over bright surfaces like snow-covered or desert areas and regions where dust reaches cold, high altitudes.

Numerous studies show the changes in top-of-atmosphere radiative flux densities as a consequence of the presence of dust using results of global simulations. There still is a considerable uncertainty in these estimates, values for global average TOA forcing by dust range from -0.6 W m$^{-2}$ (Yoshioka et al. 2007) to +0.05 W m$^{-2}$ (solar forcing only) for dust based on optical properties derived from Saharan dust measurements by Patterson and Gilette (1977) (see Balkanski et al., 2007). Case studies of instantaneous dust net radiative forcing during individual dust events find negative TOA values higher than -6 W m$^{-2}$ (Christopher and Jones, 2007, Zhu et al., 2007) or negative values larger than -400 W m$^{-2}$ at the surface (Costa et al., 2006).

Retrievals of dust optical properties from remote sensing (Kaufman et al., 2001; Dubovik et al., 2002; Sinyuk et al., 2003) lead to the conclusion that the net solar radiative forcing due to dust is negative in the global mean, suggesting that enhanced dust aerosol loads would lead to
a global cooling. One cause of uncertainties in computations of the direct radiative forcing of
dust aerosol is the lack of information on the wavelength-dependent complex refractive index
of dust particles from different source regions of the world. Results from laboratory
measurements of dust refractive indices for different samples of airborne dust have been
reported in different publications (e.g., Volz, 1973; Patterson et al., 1977; Sokolik and Toon,
1999, Müller et al., 2009, Kandler et al., 2009). Aerosol optical properties are also obtained
by inverting measured satellite radiances, which requires the assumption of a microphysical
particle model, and from multiple wavelength almucantar measurements with the
sunphotometer instruments used in AERONET. This technique was used to retrieve single-
scattering albedos for Saharan dust particles that were transported to the Atlantic Ocean (e.g.,
Colarco et al., 2002; Dubovik et al., 2002; Sinyuk et al., 2003). Optical properties, in
particular single scattering albedo values that are derived with inversion techniques and those
that are inferred from laboratory measurements of dust samples or specific minerals are often
different, mainly due to different assumptions applied for the respective approaches, in
particular about the shape of dust particles.

Mineral dust originating from the Saharan desert is of particular interest, since it contributes at
least 50% to the global dust load (Goudie and Middleton, 2001; Mahowald et al., 2005). The
first SAharan Mineral dUst experiMent (SAMUM1) aimed at constraining Saharan dust
optical properties in order to clarify the role of Saharan dust in the direct radiative forcing of
the climate system (http://samum.tropos.de/). During SAMUM1, airborne and ground-based
in-situ and remote sensing (active and passive) measurements were carried out in Morocco in
May and June 2006, complemented with satellite remote sensing and regional modeling of
atmospheric dynamics (Heintzenberg, 2009). As part of SAMUM1, dust optical properties
were derived from either in-situ measurements of absorption and scattering of solar radiation
or from measurements of particle size distribution and mineral composition of collected dust
samples (Kandler et al., 2009, Müller et al., 2009, Schladitz et al., 2009, Weinzierl et al.,
2009). Heinold et al. (2009) simulated dust distributions during SAMUM1 using a regional
atmospheric dynamics model system, which takes into account the interactions of dust
radiative forcing and atmospheric dynamics. Radiative transfer in the atmospheric column
was computed by Otto et al. (2009) to investigate the influence of particle nonsphericity
utilizing information on mineral composition and particle sizes of SAMUM1 dust samples.
They found that the radiative forcing at top of atmosphere may be enhanced by up to 30% if
particle nonsphericity is considered for the conditions of the field study at Ouarzazate in
Morocco. Bierwirth et al. (2009) studied the influence of spectral surface albedo on dust solar
and total (solar plus terrestrial) radiative forcing. The authors made use of measurements of spectral upwelling and downwelling irradiance in the solar wavelength range and of surface temperature to determine the dust solar and total radiative forcing under realistic conditions, also implementing measurements of dust and meteorological parameters as input for radiative-transfer simulations along the tracks of SAMUM1 flights.

Here we compute the solar part of the Saharan dust radiative forcing for May 19, 2006, which is one of the ‘golden days’ for which numerous different observations from the SAMUM1 field campaign are available. We use results of the regional dust model by Heinold et al. (2009) as input to the radiative transfer model to study the dependence of dust solar radiative forcing on the optical properties and surface albedo, and to provide spatial context to the results by Bierwirth et al. (2009).

2. The Dust Event on May 19, 2006

The solar radiative forcing of Saharan dust was computed for the size-resolved spatial dust distribution on May 19, 2006. A dust optical thickness of 0.4 was determined by sunphotometer measurements at 12:00 UTC at the location Ouarzazate (30°56’ N, 6°54’ W) in Morocco (Müller et al., 2010). The meteorological conditions that controlled dust emission and transport during this period are described by Knippertz et al. (2009) and Heinold et al. (2009), and are only briefly outlined here. During May 16-22, 2009 dust was transported into the region of Ouarzazate from a south-easterly direction. Further dust transport from central Algeria to southern Morocco until May 15 was observed after the formation of a lee cyclone. During the following days an upper-level air pressure ridge developed over northwestern Africa, accompanied by a surface high-pressure system centered over the eastern Atlas, causing air mass transport from eastern and central Algeria towards southern Morocco (Knippertz et al., 2009). High surface winds caused dust emissions in north-eastern Algeria, Libya, and on the border of Mali and Niger. This dust was advected over Morocco and subsequently towards the Iberian Peninsula. Dust clouds formed in Mali and Niger following convective activities during this period. The regional dust model that was used to simulate the size-resolved regional dust distributions during the SAMUM1 field experiment reproduced
the actual dust conditions in terms of dust optical thickness, particle size distributions, and extinction profiles that were measured in Ouarzazate (Heinold et al., 2009). Some inaccuracies in the model results were attributable to imprecision in the modeled location of dust emission in the presence of wet convective activity.

3. Model Description and Methods

3.1 Regional Dust Model COSMO-MUSCAT

Spatial and temporal dust distributions during the SAMUM1 period were computed with the mesoscale atmospheric dynamics model system COSMO-MUSCAT (Heinold et al., 2007, Heinold et al., 2009, Tegen et al., 2006). It is a parallelized model system consisting of two major components: the non-hydrostatic COSMO model (Steppeler et al., 2003) as meteorological driver, and the online-coupled 3D chemistry tracer transport model MUltiScale Chemistry Aerosol Transport Model (MUSCAT) (Wolke et al., 2004). In MUSCAT, advection of chemical species and aerosol particles is computed by a third-order upstream scheme. Temporal integration is carried out by an implicit-explicit method. For the SAMUM1 dust case, a horizontal grid resolution of 28 km was used. The model has 40 vertical layers, the lowest layer extends to 68 m altitude, the layer thickness increases with height. The model domain for the simulations of SAMUM1 covers the area between 13.86°N; 25.35°W (lower left corner) and 47.78°N; 38.16°E (upper right corner). The model computes dust emissions in non-vegetated areas depending on surface wind friction velocities, surface roughness, soil particle size distribution, and soil moisture (Heinold et al., 2007). Surface wind and soil moisture fields are assimilated from the meteorological model COSMO. The threshold friction velocities for initiation of dust emission are computed depending on soil particle size distribution following Marticorena and Bergametti (1995), assuming constant surface roughness within each model grid cell. Soil particle size distributions were derived from soil texture data (Zobler, 1986), assuming each texture class to be a composite of four particle size modes (clay, silt, fine sand, coarse sand). These size classes are assumed to be lognormally distributed with mode diameters at 2 μm, 15 μm, 158 μm and 720 μm (Tegen et al., 2002). The MUSCAT tracer scheme transports dust as a passive tracer in five independent size bins, covering the size range from 0.1−24 μm. The size bin limits are at 0.1 μm, 0.3 μm,
0.9 \mu m, 2.6 \mu m, 8 \mu m, and 24 \mu m. Dust is removed from the atmosphere by dry and wet
deposition processes. The computation of dry deposition follows Seinfeld and Pandis (1998);
wet deposition (both rain-out and wash-out) is parameterized following Berge (1997).

From these computations of size-resolved atmospheric dust distributions, wavelength-
dependent dust aerosol optical thicknesses \( \tau(\lambda) \) are computed as:

\[
\tau = \sum_i \frac{3Q_{ext}(\lambda, r_i)M_i}{4r_i \rho}
\]

where \( Q_{ext} \) is the dimensionless specific particle extinction efficiency depending on
wavelength \( \lambda \), \( \rho \) is the particle density, and \( M_i \) is the aerosol mass load of size fraction \( i \)
characterized by the effective radius \( r_i \) (Lacis and Mishchenko, 1995).

The spectral optical properties (extinction efficiency \( Q_{ext} \), single scattering albedo \( \omega_0 \) and
asymmetry parameter \( g \)) used to compute the changes in irradiances in the model are
calculated for each size bin from Mie theory (Mishchenko et al., 2002). The radiative
properties of the Saharan dust are prescribed in the model in different ways (Helmert et al.,
2007). An internal mixing of the major components kaolinite (98\%) and hematite (2\%) is
assumed. To compute the size-dependent optical parameters (called ‘more absorbing’ dust)
the wavelength-dependent complex refractive indices from Egan and Hilgeman (1979),
Querry (1978) as reported by Sokolik and Toon (1999) were applied, where the optical
properties of the particle mixture were estimated using the Bruggeman approximation.
Assuming internal mixtures of the minerals is justified since Kandler et al. (2009) found only
a minor amount of single iron-rich particles in samples taken during the SAMUM1 field study
(from single particle analysis of dust particles). Instead iron oxides are mostly found in
coatings or embedded in the mineral grains. Alternatively, a ‘less absorbing’ dust with optical
properties of Saharan dust derived from inverted almucantar sunphotometer measurements
(Dubovik et al., 2002; Sinyuk et al., 2003) can be used at visible wavelengths. The resulting
values for single scattering albedo for the example of a particle radius of 1.5 \mu m are 0.79 for
the ‘more absorbing’ and 0.91 for the ‘less absorbing’ dust in the COSMO wavelength band
of 250-700 nm, and 0.98 and 0.99 in the wavelength band of 700 nm to 1530 nm, respectively
(Helmert et al., 2007). To facilitate comparisons with other studies, the single scattering
albedo values for particles with 1.5 \mu m effective radius are also provided for the wavelengths
of AERONET sunphotometers (e.g. Dubovik et al., 2002) at 440 nm, 670 nm, 870 nm,
1020 nm as well as at the reference wavelength of 550 nm (Table 1).
Dust optical properties are implemented in the COSMO radiation scheme by spectral integration for each of the three solar and five terrestrial (thermal infrared, IR) bands. In COSMO-MUSCAT, the change in irradiances due to the presence of dust impacts the atmospheric dynamics, and feeds back on dust emissions and atmospheric distributions of the dust (Heinold et al., 2008). Here, 3D fields of modeled size-resolved dust distribution for May 19, 2006 12:00 UTC were used as input for the 1D radiative transfer model libRadtran (Mayer and Kylling, 2005), which was used to compute dust radiative forcing within the solar wavelength range.

### 3.2 Radiative Transfer Model libRadtran

Similar to Bierwirth et al. (2009), the radiative transfer for the selected Saharan dust case was computed with the software package libRadtran, described by Mayer and Kylling (2005). The input fields of libRadtran include profiles of thermodynamic variables such as air temperature, density, water vapor and trace gas concentrations, wavelength-dependent surface albedo and extraterrestrial radiation to calculate radiation quantities within the atmosphere. The radiative transfer equation is solved in the 6-stream plane-parallel DISORT2 algorithm by the Discrete-Ordinate Method (Stamnes et al. 1988). Gas absorption is computed using the parameterization SBDART/LOWTRAN (Ricchiazzi et al. 1998). Bierwirth et al. (2009) use observations of thermodynamic and dust parameters measured during SAMUM1 as input for libRadtran. In contrast, in this work the clear-sky solar radiative forcing of the size-resolved dust distribution resulting from the regional model simulations was computed using libRadtran (version 1.4). For each size class the computations of the optical properties were performed by Mie calculations according to Mishchenko et al. (2002). Dust extinction, single scattering albedo, and asymmetry parameter were computed as external mixtures of the different simulated size classes as in Mallet et al (2003).

Below 12 km height the layer heights of the radiative transfer model were modified to fit the layers of the COSMO model. Further input parameters for the libRadtran model that were taken from COSMO output were air temperature, pressure, and humidity. The solar zenith angle was calculated using the geographic coordinates, date, and local time. The ozone column concentrations were set to 290 Dobson Units, according to retrievals with the Ozone Monitoring Instrument (OMI, Levelt et al., 2006) on May 19, 2006. Standard values (US standard atmosphere) were assumed for concentrations of other atmospheric trace constituents
While in Bierwirth et al. (2009) the dust single-scattering albedo and the asymmetry parameter were derived from ground-based measurements at Tinfou (Schladitz et al., 2009), here we have computed the optical properties according to the ‘more absorbing’ and ‘less absorbing’ dust parameterizations in COSMO-MUSCAT.

Two sets of experiments were carried out with the libRadtran model. In one set the conditions at Ouarzazate on May 19, 2006 were considered, and atmospheric radiative transfer was computed at the solar wavelength range (320 to 2100 nm) with and without the presence of the modeled dust for this location. In another set of experiments a 1°x1° degree map of solar dust radiative forcing was computed for the Sahara for the area between 14.5°W and 19.5°E, 15.5°N and 34.5°N. The computations were performed using three datasets of surface albedo: airborne measurements with the SMART Albedometer within SAMUM1 (Bierwirth et al., 2009), retrievals from the Moderate Resolution Imaging Spectroradiometer (MODIS) satellite instrument, and surface albedo values from the COSMO model.

### 3.3 Surface Albedo Measurements During SAMUM1

Surface albedo, the ratio of upwelling and downwelling irradiances at the surface, strongly influences the TOA radiative forcing of clouds and aerosols. Surface albedo measurements were performed on 13 measurement flights starting from the airport of Ouarzazate (Bierwirth et al., 2009) as part of the SAMUM1 field experiments. The flights covered the region between 9°W and 4.5°W and 30°N and 31.5°N. Solar spectral irradiances were measured in the spectral range from 290 to 2200 nm by the Spectral Modular Airborne Radiation Measurement System (SMART)-Albedometer (Wendisch et al., 2001, 2004). The SMART Albedometer was operated with optical inlets for the detection of upwelling and downwelling spectral irradiances and actinic flux densities. The surface albedo was retrieved applying an iterative algorithm by Wendisch et al. (2004) that eventually removes the impact of aerosol from the albedo measurements at flight level. Measurements were taken for different surface types. An important surface type in the vicinity of Ouarzazate and in the Zagora basin is Hamada, a brown-red stone desert. Other typical surface albedo spectra were measured, e.g., for oasis surfaces and for the Iriki dry salt-lake near the border to Algeria. The salt-lake surface is characterized by high albedo values (0.18 at 500 nm, increasing to 0.5 at
1500 nm wavelength). The surface albedo values of the darker Hamada surface are at 0.1 at 500 nm, increasing to approximately 0.28 at 1500 nm wavelength. Typical spectral surface albedo data measured for different surfaces during SAMUM1 are presented in Bierwirth et al. (2009).

The SMART-Albedometer measurements provide high-quality spectral surface albedo information for individual locations. To utilize this information in large-scale estimates of aerosol radiative forcing it must be put into a regional context. Large-scale surface albedo information is available from satellite remote sensing. Here, we use retrievals from the MODIS satellite instrument.

The 16-day surface albedo product MOD43B3 is available at 1 km resolution. Provided are albedo data for MODIS bands 1-7 (620-670 nm, 841-876 nm, 459-479 nm, 545-565 nm, 1230-1250 nm, 1628-1652 nm, 2105-2155 nm) and for three broad bands (300-700 nm, 700-5000 nm, and 300-5000 m). The operational albedo algorithm makes use of a semi-empirical model to determine a global set of parameters that describe the bidirectional reflectance of the land surface. These parameters are then used to determine the diffuse albedo (white-sky) and direct-beam albedo in the absence of a diffuse component (black-sky surface albedo) (Moody et al, 2008). The actual surface albedo is interpolated between these two values as a function of the fraction of diffuse light. In addition to the tiled 1-km resolved product, the International Satellite Land Surface Climatology (ISLSCP) Initiative (Hall et al 2006) provides climatologic land-surface information at regular grids of 0.25, 0.1 and 1 degree resolution useful for climate modeling. Both the 1-km product and the 1-degree product of MODIS surface albedo data are compared with the local spectral albedo results from the SMART Albedometer measurements and used as input for the radiative transfer model libRadtran.

4. Results

4.1 Controls of External and Dust Optical Properties on Dust Solar Radiative Forcing

(a) Solar zenith angle (SZA)
The model results for the SAMUM1 period are described in detail in Heinold et al. (2009), including the comparison of the simulated dust distributions with results from different measurements during the field experiments in terms of dust extinction profile, optical thickness, and particle size distribution at ground and elevated levels (Esselborn et al., 2009, Petzold et al., 2009, Schladitz et al., 2009, Tesche et al., 2009). In addition, the modeled horizontal dust distributions were compared to satellite remote sensing retrievals of dust properties. The model reproduced the observed distributions of optical thickness and dust particle size distributions. The spatio-temporal evolution of the dust plumes in the model generally matched the observations during the SAMUM1 period (Heinold et al., 2009).

Deficiencies were related to inaccuracies in the placement of dust emission events by the model, in particular for dust events related to wet convective activities (Heinold et al., 2009, Reinfried et al., 2009). Figure 1 shows the modeled dust optical thickness distribution at 550 nm on May 19, 2006, 12:00 UTC. An arc-shaped plume of dust transported toward the Moroccan coast can be identified in the northwestern part of the Sahara. The dust optical thickness reached values up to 1 in northern Mauritania, as well as in parts of Niger and Chad. In the northeastern part of the Sahara the model simulations result in relatively dust-free conditions. At 12:00 UTC the model computed a dust optical thickness of 0.38 at the location Ouarzazate in Morocco, which is in good agreement with the measured value of 0.4 (Müller et al., 2009).

At this location, the modeled dust extinction profiles and meteorological fields were used as input for libRadtran to compute the clear-sky solar radiative forcing of the modeled dust. Optical properties of the dust aerosol particles were calculated from the mixtures of the individual size distributions in each layer. The modeled size distribution shows a decrease in the contribution of supermicron-sized particles at higher altitudes as larger particles are removed from the atmosphere more quickly due to gravitational settling than smaller particles (Figure 2). The dependence of the computed dust solar radiative forcing at TOA in the solar wavelength range of 320-2100 nm on different solar zenith angles (SZA) is shown in Figure 3a for both the ‘more’ and ‘less’ absorbing dust. Results are shown for two observed spectral surface albedo types, the dark Hamada and the bright Iriki salt-lake surfaces (Bierwirth et al. 2009), and compared to the results obtained for a spectrally constant surface albedo of 0.2 (black line in Figure 3a). In this case study, the TOA dust solar radiative forcing depends more strongly on variations of the surface albedo than on the optical properties for the chosen parameter range. At low solar zenith angles (values less than 10°) the difference in TOA solar radiative forcing assuming ‘more’ or ‘less’ absorbing dust is between 5 and 7 W m⁻², while
assuming either the measured surface albedo values for Hamada or Iriki salt lake conditions results in differences in TOA solar forcing between 12 and 17 W m$^{-2}$ for dust aerosol at Ouarzazate with the same optical properties. Positive solar forcing at the low SZA near noontime is evident for all cases except for the ‘less absorbing’ aerosol over the darker Hamada surface, while all cases result in negative dust solar forcing at TOA at high solar zenith angles. The result from Bierwirth et al. (2009) is marked in Figure 3a; it agrees well with the case of more absorbing aerosol over Hamada surface.

The radiative transfer solver DISORT used in the libRadtran computations is based on a plane-parallel approximation to compute the radiative transfer. Using a pseudo-spherical solver (SDISORT) within the libRadtran model (Mayer and Kylling, 2005) leads to small differences of less than 2 W m$^{-2}$ at high SZAs above 75 degrees (not shown) compared to the standard computations using DISORT. Given the uncertainties in input parameters in the model it is acceptable to use the plane-parallel approximation for the radiative transfer calculations.

(b) Albedo and optical properties

The results for instantaneous 12:00 UTC and diurnal mean solar radiative forcing for different surface albedos and dust optical properties are shown in Tables 2 and 3 for TOA and surface, respectively, for the location of Ouarzazate on May 19, 2006. In addition Figures 3b and 3c show the change in TOA solar radiative forcing at 12:00 UTC and the diurnal average for different dust optical properties in dependence on surface albedo, which is here assumed to be constant in the solar wavelength range. For the diurnal mean, the solar radiative forcing values were computed as 24hr averages, using hourly values of dust extinction profiles and meteorological parameters from the model simulations as input parameters for libRadtran. In Tables 2 and 3 the results are shown for both TOA and surface forcing, for the spectral surface albedo data of dark Hamada, bright salt-lake surface, and several spectrally constant values between 0 and 0.3. The values are computed for different optical properties, including simulations with the simulated dust size distribution for the ‘more’ and ‘less’ absorbing dust cases, as well as for the assumption that the dust aerosol consists of a single particle radius of 0.5, 1.5 and 4.6 μm, respectively. Those values represent the effective dust particle radii of the smallest size bins in the regional dust model. For the size distribution simulated by the model, instantaneous TOA solar radiative forcing values of the dust event at 12:00UTC range
from -3.7 to +17 W m\(^{-2}\) when using measured values of surface albedo as model input, and from -28 W m\(^{-2}\) for constant surface albedo 0 to +23 W m\(^{-2}\) for constant surface albedo 0.3, considering different dust types. In the diurnal average, the solar radiative forcing at TOA is generally negative with the exception of the case of ‘more absorbing’ aerosol over a very bright surface (spectrally constant albedo of 0.3) (see Table 2, Figure 3c). For realistic surfaces, the values of diurnally average TOA forcing range between -9.8 W m\(^{-2}\) and -0.36 W m\(^{-2}\). The assumption of a fixed particle radius modifies the results of the radiative solar computations considerably. Due to their lower single scattering albedo at solar wavelength, large particles (here represented by particles with an effective radius of 4.6 μm) would lead to positive forcing at solar wavelengths over many land surfaces even in the diurnal mean. While the assumptions of mono-disperse dust particle distributions with effective particle sizes of 0.5 μm or 4.6 μm are clearly unrealistic for the dust distribution at the location of Ouarzazate, the size distribution of particles of the same composition nevertheless plays an important role when determining the dust solar radiative forcing.

At the surface, the dust clear-sky solar radiative forcing is always negative, but the magnitude of the surface forcing by dust can vary by a factor of 2-3 for the different assumptions of dust particle size and surface albedos. The values range between -94 W m\(^{-2}\) and -29 W m\(^{-2}\) for 12:00 UTC and -48 W m\(^{-2}\) and -20 W m\(^{-2}\) in the diurnal average (Table 3). For the particle size distribution computed by the regional model and for a realistic surface albedo, the surface solar forcing ranges between -52 W m\(^{-2}\) and -31 W m\(^{-2}\) for 12:00 UTC and -30 W m\(^{-2}\) and -20 W m\(^{-2}\) for the diurnal mean considering the cases of ‘more’ and ‘less’ absorbing dust. Thus in this case study the surface solar forcing varies only by about 10 W m\(^{-2}\) for different dust optical properties and surface albedos at the given size distribution in the diurnal average. For comparison, Bierwirth et al (2009) report values of instantaneous forcing for the SAMUM1 experiment that range from -15 W m\(^{-2}\) to +10 W m\(^{-2}\) at TOA, and from -65 W m\(^{-2}\) to -45 W m\(^{-2}\) at the surface at individual locations.

### 4.2 Comparison of Surface Albedo Products

Computations of dust solar radiative forcing at TOA require an accurate characterization of the underlying spectral surface albedo. Surface albedo values in the COSMO model are determined by the soil type, soil moisture, vegetation cover and coverage of the surface with snow or ice. For the Sahara desert, we compare the use of surface albedo values from the
COSMO model for solar radiative forcing calculations with the use of surface albedo values retrieved from satellite remote sensing by the MODIS instrument (Moody et al., 2005). The 16-day MODIS MOD43B3 product at 1 km resolution for May 2006 is compared at 7 locations to the SMART Albedo meter measurements from Bierwirth et al., (2009) (Figure 4). The locations include measurements at two sites with Hamada surface, Iriki salt lake, a farmland site, an oasis, and a small reservoir, covering different surface characteristics. In addition, climatologic surface albedo data from the monthly (May) ISLSCP product interpolated to 1°×1° resolution are shown in Figure 4 at the grid cells containing the locations of the SMART Albedo meter measurements. While Liu et al. (2009) estimate the overall uncertainty of the MODIS surface albedo product to be within the range of 0.05, the error bars of the MODIS results in Figure 4 indicate the ranges between the ‘black sky’ and ‘white sky’ surface albedo values. The data show good agreement between the 1-km MODIS product and the SMART Albedo meter measurements in most cases. For the Hamada and Iriki salt-lake locations the agreement between the different surface albedo products is even very good. Differences between the spectral albedo products agree within the range given for the MODIS product for wavelengths under 1000 nm, and remain lower than 0.05 for longer wavelengths. For wavelengths longer than 700 nm the MODIS values clearly exceed the SMART Albedo meter measurements at the reservoir location; for the oasis location this is the case in for wavelength above 1300 nm. Those discrepancies may be due to small-scale surface features that are not resolved at 1-km resolution. Some deviations can be expected for the comparison of the albedo measurements and the MODIS product, since the aircraft and satellite measurements have different fields of view that also do not necessarily overlap completely. In particular at the reservoir location where a water surface with very low albedo is surrounded by bright desert surface small deviations in the field of view can lead to considerable differences in the areal average value of surface albedo. Nevertheless, even for the coarser grid 1°×1° the climatologic MODIS surface albedo product mostly agrees with the aircraft measurements and this motivates the use of MODIS surface albedo values as input for the radiative transfer model to compute Saharan dust solar radiative forcing on a regional scale.

4.3 Regional Dust Solar Radiative Forcing and Dependencies
For the area between 14.5°W and 19.5°E, 15.5°N and 34.5°N the modeled profiles of dust extinction, optical properties, and meteorological parameters on May 19, 2006 were re-gridded to 1°×1° horizontal resolution. Each of the resulting 700 dust profiles was then used as input for the libRadtran model to obtain a spatial distribution of dust forcing. In one set of model experiments, the surface albedo value in each of the 1°×1° grid cells was taken from the prescribed surface albedo of the COSMO model, which is assumed to be constant in the solar wavelength range and re-gridded to the coarser resolution. In another set of experiments the MODIS surface albedo product was used, computed as mean value of the ‘black sky’ and ‘white sky’ products, re-gridded to 1°×1° horizontal resolution. Spectral surface albedo values in libRadtran were interpolated from the wavelengths bands 1-7 of the MODIS data product. Saharan surface albedo maps in the solar wavelength range of MODIS (averaged for the wavelength bands 1-7) and the COSMO model differ considerably (Figure 5). The MODIS surface albedo product reaches values up to 0.5 with maximum values in the Bodélé region. Lower surface albedo values are observed in mountain regions like the Atlas or the Ahaggar mountains (Figure 5, left panel). In the COSMO model the surface albedo values are prescribed for each grid cell from soil composition, which is not always well known for the desert region. Saharan surface albedo values that are prescribed in the regional model are generally lower than the satellite-derived values, rarely exceeding a value of 0.25 (Figure 5, right panel).

The sign of the TOA clear-sky solar radiative forcing computed for the SAMUM case of May 19, 2006 is determined by both surface albedo and the optical properties of the dust aerosol (Figure 6). While the instantaneous TOA dust solar forcing at 12:00 UTC is always positive in the regional average over the Sahara when computed with MODIS surface albedo values, the dust forcing results are more often negative when COSMO surface albedo values are used for the desert surface. For the case of the ‘less absorbing’ dust and COSMO surface albedo the TOA solar forcing is negative at almost every location in the Sahara. The regional average instantaneous TOA solar radiative forcing for the configurations shown in Figure 6 range from +11 W m⁻² (for the ‘more absorbing’ dust and MODIS surface albedo) to -7.2 W m⁻² for
the ‘less absorbing’ dust and COSMO surface albedo. Maximum positive instantaneous TOA solar forcing can vary by nearly an order of magnitude depending on the dust optical properties and surface albedo values, in this case maximum results range from +13 W m$^{-2}$ up to +105 W m$^{-2}$ in the Bodélé region in Chad where the dust optical thickness reaches values of 1 and the surface albedo in the MODIS data is as high as 0.5 due to the presence of bright diatomite deposits (Washington and Todd, 2005). Negative instantaneous TOA forcing over the ocean surface ranges from -106 to -144 W m$^{-2}$ for the different experiments. Differences between the TOA solar radiative forcing results for separately considering the different dust optical properties or surface albedos are shown in Figure 7. Using the different surface albedo fields (Figure 7b) results in the same order of magnitude of TOA forcing differences as computing the solar radiative forcing by either assuming ‘less absorbing’ or ‘more absorbing’ dust aerosol (Figure 7a). The differences in instantaneous solar TOA forcing for different dust optical properties are up to -50 W m$^{-2}$ and occur in areas where the dust optical thickness reaches values of 1. The regional average of this difference over the area of the Sahara that is considered in this computation is -9.0 W m$^{-2}$. The difference for different surface albedos (Figure 7b) ranges between -60 and +50 W m$^{-2}$ with a regional average of -9.4 W m$^{-2}$, in this case some cancellation occurs between positive and negative differences. This demonstrates the importance of the complete characterization of both optical properties and surface albedo to characterize the dust solar radiative forcing over a bright surface. It has to be noted that the assumption of particle sphericity for the non-spherical particles may lead to an underestimation of the instantaneous TOA forcing by dust of up to 30% at mid-visible wavelengths (Otto et al., 2009), which adds another uncertainty to the computation of radiative forcing by Saharan dust.

In field studies, the dust size distribution and its optical properties are often determined from dust samples that are taken near the surface (e.g., Schladitz et al., 2009). However, a near-surface dust particle size distribution may not be representative for conditions in the atmospheric column, as large particles may be deposited to the ground quickly and do not reach higher atmospheric layers. To test to which extent such assumptions could influence the computed dust solar radiative forcing, a sensitivity study was performed in which the size-dependent dust optical properties in the entire column were replaced by the optical properties of the lowest model layer. Thereby the column optical thickness remained unchanged compared to the standard case (‘more absorbing’ dust and MODIS surface albedo). The differences in TOA solar forcing from this experiment and the standard case (Figure 8) show that the positive solar radiative forcing using first layer dust optical properties is
overestimated compared to the standard case, with differences of up to 30 W m\(^{-2}\) in regions with high dust optical thickness. This difference is due to the fact that larger dust particles are present at the surface level (compare also Figure 2), which are more absorbing at solar wavelengths compared to the smaller-sized dust particles that dominate the size distribution at higher layers. The increased absorption by larger-sized dust particles (see, e.g., Tegen and Lacis, 1996) compared to the standard case leads to more positive TOA solar forcing, but the differences to the computation with the standard case are smaller than those that are caused by using different surface albedo fields, which may lead to differences of up to 50 W m\(^{-2}\).

In contrast to dust solar radiative forcing at TOA, the choice of surface albedo plays a smaller role for the solar radiative forcing at the surface (Figure 9). In the regional average the instantaneous forcing for the test case is -43 W m\(^{-2}\) and -57 W m\(^{-2}\) prescribing the ‘more absorbing’ dust with MODIS and model surface albedo, respectively; and -29 W m\(^{-2}\) and -38 W m\(^{-2}\) for surface solar forcing of the ‘less absorbing’ dust. Maximum negative solar forcing values range from -390 W m\(^{-2}\) to -256 W m\(^{-2}\). Differences between the computed surface solar radiative forcing for considering the different dust optical properties or surface albedos separately are shown in Figure 10. The surface forcing is more negative for the ‘more’ than for the ‘less’ absorbing dust. Assuming different optical properties leads to differences in instantaneous solar surface forcing of 22 W m\(^{-2}\) in the regional average, with local maxima in the differences of 70 W m\(^{-2}\). In contrast the differences in assuming different surface albedo fields is -8 W m\(^{-2}\) for the difference in regional average surface forcing when using COSMO and MODIS surface albedos and ‘more absorbing’ dust (Figure 10b). The differences in instantaneous TOA and surface forcings result in an atmospheric forcing at solar wavelengths of 53 W m\(^{-2}\) and 30 W m\(^{-2}\) for the regional average for the ‘more’ and ‘less’ absorbing dust, respectively, independent of the surface albedo.

The diurnal averages of the dust forcing for May 19, 2006 were computed in addition to the instantaneous values at 12:00 UTC using hourly values for profiles of dust extinction and meteorological parameters from the COSMO—MUSCAT model as input for the libRadtran computations. For the computation of the diurnal average it was assumed that the surface albedo has no dependence on the solar zenith angle. In the 24-hour average the results for TOA forcing in the wavelength range of 320-2100 nm are generally negative, ranging from -1.7 W m\(^{-2}\) to -9.4 W m\(^{-2}\) for the regional average, as the solar radiative forcing by dust is generally negative for solar zenith angles above 40° (compare Figure 3a). Only for the more realistic surface albedo from the MODIS measurements the solar radiative forcing is positive
over very bright desert regions like the Bodélé depression where surface albedo values of up to 0.5 are indicated by the MODIS dataset (Figure 11). At that location the TOA solar forcing by dust can reach -30 W m\(^{-2}\). The difference in diurnally averaged solar clear-sky forcing reaches a negative maximum of -18 W m\(^{-2}\) and is -3.7 W m\(^{-2}\) in regional average for the difference of ‘less’ and ‘more’ absorbing dust using MODIS surface albedo (Figure 12a). The difference between using different surface albedos and the ‘more absorbing’ dust (Figure 12b) is -5.1 W m\(^{-2}\) in the regional average with local differences ranging between -25 and +20 W m\(^{-2}\).

5. Discussion and Conclusions

The clear-sky solar radiative forcing by Saharan dust was computed with the radiative transfer model libRadtran for a case study during the SAMUM1 experiment in Morocco in May 2006. Size-resolved dust concentration and optical thickness were simulated by the regional transport model COSMO-MUSCAT. Dust optical properties in the atmospheric column were computed from the mixture of different particle sizes in the model. ‘More absorbing’ and ‘less absorbing’ dust was prescribed by either assuming a mineral composition containing 2% hematite, which is highly absorbing in the solar wavelength range, or assuming absorption properties derived from sunphotometer measurements. The results presented here are negative in the diurnal average for the range of single scattering albedos if the surface albedo remains below 0.25. This narrows down earlier estimates by e.g. Miller et al (2004), who investigated the sensitivity of the direct effect of dust in a global model for single scattering albedo values that were increased and decreased by 10% to represent more and less absorbing dust aerosol. Within that range they found possible values for dust radiative TOA forcing at solar wavelengths between -5 and +4 W m\(^{-2}\) for the Sahara for northern hemisphere spring conditions. The difference between the instantaneous and diurnally average solar dust forcing agrees with earlier results by Christopher et al (2003), who point out the importance of considering the diurnal cycle of dust optical thickness for radiative forcing estimates due to the SZA dependence of instantaneous forcing by dust. Their results in solar dust forcing over ocean regions (12 W m\(^{-2}\) at TOA and 18 W m\(^{-2}\) at the surface, ‘less absorbing’ dust) are consistent with the results presented here considering the dust optical thickness in their experiment ranged at about 50% of the values observed at Ouarzazate during SAMUM1. Regional variations in TOA dust solar forcing are strongly dependent on the spectral albedo of the underlying surface. Locally the differences in solar dust forcing due to the prescription
of the albedo of the underlying surface can exceed the uncertainty range that occurs due to differences dust optical properties. Claquin et al (1998) already showed that dust aerosol with a single scattering albedo of 0.89 changes the sign of the forcing from negative to positive when surface albedo values increase above 0.25, which is within the range of the results presented here. Over the Sahara the use of surface albedo values from remote sensing provides a useful alternative to deriving surface albedo values from prescribed soil types, as the information on soil types in the desert areas can be patchy.

The impact of the presence of dust on atmospheric dynamic is due to radiative heating by dust, which depends on its optical properties. Complex refractive indices depend on the mineral composition of the dust as well as on the structure of the individual particles. However, the optical properties also depend on the particle size distribution which can be described sufficiently well by dust transport models.

One of the major goals of SAMUM1 was to reduce uncertainties in dust radiative forcing by appropriate measurements. Absorption by dust particles depends in particular on the imaginary part of the refractive index of the particles. The investigation of dust properties in the field campaign using different methods results in a range from $1.5 \times 10^{-3}$ (Müller et al. 2009) to $6.6 \times 10^{-3}$ (Kandler et al, 2009) for the imaginary part of the dust refractive index at 530 nm wavelength. The prescribed optical properties of the ‘more’ and ‘less’ absorbing dust used in this model study correspond to imaginary parts of the complex refractive indices of $5.8 \times 10^{-3}$ and $2.2 \times 10^{-3}$, respectively. Thus the range of the prescribed values represents the range of the measurements well, reflecting in part the remaining uncertainty in solar radiative forcing by Saharan dust. The assumption of particle sphericity for the non-spherical dust particles adds to the remaining model uncertainty. Otto et al (2009) note that for the conditions of the field study at Ouarzazate in the dust radiative forcing at TOA may be enhanced by up to 30% if particle nonsphericity is considered.
Acknowledgements

This work is a contribution to the project SAMUM (Saharan Mineral Dust Experiment) funded by the German Research Foundation (DFG). We thank the Deutscher Wetterdienst (DWD) for good cooperation and support.
References


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Schladitz, A., T. Müller, A. Massling, N. Kaaden, K. Kandler, and A. Wiedensohler (2009), In situ measurements of optical properties at Tinfou (Morocco) during the Saharan Mineral Dust Experiment SAMUM 2006, *Tellus B, 61*, 64-78.


Tables

Table 1: Single scattering albedo for dust particles with an effective radius of 1.5 \( \mu m \). The values are given for the wavelengths of AERONET sunphotometers and additionally for 550 nm.

<table>
<thead>
<tr>
<th>Wavelength (nm)</th>
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<th>‘Less absorbing’ dust</th>
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<tr>
<td>1020</td>
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<td>0.99</td>
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</table>

Table 2: TOA dust clear-sky solar forcing (in Wm\(^{-2}\)) on 19.5.2006, Ouarzazate, Morocco, for different surface albedos (Hamada and Salt Lake from Bierwirth et al, (2009)). Aerosol
optical thickness is 0.41. First value: diurnal average/ second value: instantaneous forcing at 12:00 UTC (SZA: 13°). Wavelength range is 320-2100 nm. Compare also Figure 3b, 3c.

- Type 1 dust: Particle size distribution from regional model, varies with height, 'more absorbing' dust as defined in text.
- Type 2 as type 1, but 'less absorbing' dust.
- Type 3 as type 2, but with $r_{\text{eff}}=0.5 \, \mu m$,
- Type 4 as type 2, but with $r_{\text{eff}}=1.5 \, \mu m$,
- Type 5 as type 1, but with $r_{\text{eff}}=0.5 \, \mu m$,
- Type 6 as type 1, but with $r_{\text{eff}}=1.5 \, \mu m$,
- Type 7 as type 1, but with $r_{\text{eff}}=4.6 \, \mu m$
Table 3: Clear-sky solar surface dust forcing (in Wm$^{-2}$) on 19.5.2006, Ouarzazate for different surface albedos. Description see Table 1.

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<th>Albedo 0.2</th>
<th>Albedo 0.3</th>
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Figure Captions

Figure 1: Dust optical thickness (550 nm) at May 19, 2006 12:00 UTC computed with COSMO-MUSCAT (Heinold et al 2009), regridded to 1°x1° resolution. The location of Ouarzazate (Morocco) is indicated by the star symbol. Maxima of modeled dust optical thickness reach values up to 4.

Figure 2: Normalized dust particle extinction size distribution computed by COSMO MUSCAT for May 19, 2006 12:00 UTC, in the surface layer (solid line) and at 4.3 km height (dashed line).

Figure 3: (a): Solar dust TOA clear-sky solar radiative forcing (wavelength range of 320-2100 nm) in dependence on solar zenith angle (SZA) for different dust optical properties and surface albedo values, at Ouarzazate, May 19, 2006, 12:00UTC. The color of the lines and symbols indicates the surface albedo. Red: Hamada, green: Iriki Salt-lake (both from Bierwirth et al, (2009)), black: spectrally constant surface albedo value of 0.2. Symbols: diamonds: Dust optical properties following Sokolik et al. (1999) assuming a mixture of 2% Hematite, 98% kaolinite; asterisks: dust optical properties according to Dubovik et al. (2002). The square symbol shows the result from Bierwirth et al (2009) computed from observed dust aerosol properties at Ouarzazate (see Fig 8 therein). (b),(c): Solar dust TOA clear-sky solar radiative forcing (wavelength range of 320-2100 nm) in dependence on surface albedo (spectrally constant) for different dust optical properties (compare Table 2) and surface albedo values. Black: dust size distribution from model results, blue: \( r_{\text{eff}} = 0.5 \, \mu m \), red: \( r_{\text{eff}} = 1.5 \, \mu m \), orange: \( r_{\text{eff}} = 4.6 \, \mu m \). Shown are results for instantaneous solar forcing at 12UTC (b) and diurnally average forcing (c); note the different ranges of radiative forcing values.

Figure 4: Comparison of spectral surface albedo measurements by Bierwirth et al. (2009) (black dots) and MODIS retrievals of surface albedo. The error bars indicate the difference between white-sky and black-sky albedos in the MODIS data. Blue: Climatological ISLSCP MODIS surface albedo product for 05/2002, 1 deg. horizontal resolution. Red: 16-day MODIS MOD43B3 product at 1 km resolution for May 2006.

Figure 5: Saharan surface albedo at solar wavelengths, derived from MODIS (left panel) and COSMO (right panel).

Figure 6: Instantaneous TOA solar clear-sky radiative forcing of modeled dust distribution on May 19, 2006, 12:00 UTC, computed by libRadtran. Top panels: MODIS surface albedo,
lower panels: COSMO surface albedo. Left panels: ‘more absorbing’ dust (containing 2% hematite), right panels: ‘less absorbing’ dust from AERONET inversions. Note the use of a logarithmic color scale.

**Figure 7:** (a) Difference between instantaneous TOA dust clear-sky solar forcing (May 19, 2006, 12UTC, MODIS surface albedo) between the ‘more absorbing’ and the ‘less absorbing’ dust case, and (b) between the case computed with COSMO surface albedo and MODIS surface albedo (‘more absorbing’ dust case.)

**Figure 8:** Difference in TOA solar forcing between assuming optical properties of the lowermost model layer for the entire column and the standard case (‘more absorbing’ dust, MODIS surface albedo)

**Figure 9:** As Figure 6, for instantaneous surface clear-sky solar forcing.

**Figure 10:** As Figure 7, for instantaneous solar surface clear-sky forcing.

**Figure 11:** As Figure 6, but for 24-hour average solar TOA clear-sky solar forcing.

**Figure 12:** As Figure 7, but for 24-hour average solar TOA clear-sky forcing.
Messungen Bierwirth et al., 2009

MODIS Terra, 19.5.2006, 1km Produkt
MODIS 5/2002 1deg Produkt
- Messungen Bierwirth et al., 2009
More Absorbing Dust
Less Absorbing Dust

MODIS

COSMO

W/m²

-120.0  -70.0  -20.0  -10.0  -5.0  -2.0  2.0  5.0  10.0  20.0  70.0  120.0
'More Absorbing' Dust

'Less Absorbing' Dust

MODIS

COSMO

-120.0 -70.0 -20.0 -10.0 -5.0 -2.0 2.0 5.0 10.0 20.0 70.0 120.0

W/m²
'More Absorbing' Dust

'More Absorbing' Dust

'More Absorbing' Dust

'More Absorbing' Dust