Quantifying dust plume formation and aerosol size distribution during the Saharan Mineral Dust Experiment in North Africa

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Abstract

Dust particles mixed in the free troposphere have longer lifetimes than airborne particles near the surface. Their cumulative radiative impact on earth’s meteorological processes and climate might be significant despite their relatively small contribution to total dust abundance. One example is the elevated dust–laden Saharan Air Layer (SAL) over the equatorial North Atlantic, which cools the sea surface and likely suppresses hurricane activity. To understand the formation mechanisms of SAL, we combine model simulations and dust observations collected during the first stage of the Saharan Mineral Dust Experiment (SAMUM–I), which sampled dust events that extended from Morocco to Portugal, and investigated the spatial distribution and the microphysical, optical, chemical, and radiative properties of Saharan mineral dust. We employed the Weather Research Forecast model coupled with the Chemistry/Aerosol module (WRF--Chem) to reproduce the meteorological environment and spatial and size distributions of dust. The experimental domain covers northwest Africa including the southern Sahara, Morocco and part of the Atlantic Ocean with 5 km horizontal grid spacing and 51 vertical layers. The experiments were run from 20 May to 9 June 2006, covering the period of most intensive dust outbreaks. Comparisons of model results with available airborne and ground--based observations show that WRF--Chem reproduces observed meteorological fields as well as aerosol distribution across the entire region and along the airplane’s tracks. We evaluated several aerosol uplift processes and found that orographic lifting, aerosol transport through the land/sea interface with steep gradients of meteorological characteristics, and interaction of sea breezes with the continental outflow are key mechanisms that form a surface–detached aerosol plume over the ocean. Comparisons of simulated dust size distributions with airplane and ground--based observations are generally good, but suggest that more detailed treatment of microphysics in the model is required to capture the full--scale effect of large aerosol particles.

Key words: Regional modeling, fine resolution, WRF--Chem, SAL, boundary layer, dust load
1. Introduction

Mineral dust particles from the Sahara often travel thousands of kilometers across the Atlantic Ocean and into Northern Europe (Dinter et al., 2009). The vertical transport of dust has been the subject of multiple studies (Donnell et al., 2001; Takemi and Seino, 2005; Yasui et al., 2005). However, the quantity and size distributions of dust, as well as the relative importance of the meteorological processes triggering vertical transport over the Sahara, are not completely understood (Marsham et al., 2008; Laurent et al., 2010). Recent studies (Donnell et al., 2001; Cuesta et al., 2009; Solomos et al., 2012) have described dust--uplifting mechanisms associated with moist and dry convection, diurnal variability, and dynamic lifting events. Penetrating convection is an important mechanism that is capable of vertical mixing and transport of chemicals and fine dust particles into the upper layers of the troposphere (Pickering et al., 1992; Stenchikov et al., 1996). Mineral dust mobilization and transport occur at different spatial and temporal scales (Knippertz et al., 2008; Schepanski et al., 2009). On the synoptic scale, extra--tropical depressions and desert heat low are the most important phenomena related to southward penetrating upper--level troughs favorable to Harmattan wind development in winter and spring under meteorological conditions associated with intense anticyclone--genesis (Kalu, 1979; Knippertz and Fink, 2006; Knippertz et al., 2008). On the meso--scale, density currents are important for dust mobilization and transport (Schepanski et al., 2009; Emmel et al., 2010). Density currents can be triggered by various meteorological phenomena such as downdrafts from convective storms, evaporative cooling, and orographic storm activity over mountains (Sutton, 1925; Knippertz et al., 2007). These systems can produce fast--moving “dust walls” generally called “Haboobs” (Sutton, 1925; Membery, 1985; Knippertz et al., 2007). On the micro scale, dust mobilization is associated with dry convective mixing in the atmospheric boundary layer and wind--driven erosion from arid and semi--arid topographical depressions (Gamo, 1996; Flamant et al., 2007). In addition, fast rotating dust devils and non--rotating convective plumes also contribute to dust emissions in this region (Koch and Renno, 2005).

The Sahara desert is the largest dust source worldwide. It contributes more than half of the mineral dust to the earth’s total global atmospheric dust budget (Washington et al., 2005). Given the limited data from monitoring, the small number of field campaigns, and the complexity of the microphysical and radiative properties of atmospheric aerosols, regional dust modeling has been increasingly used to understand and investigate
atmospheric aerosols and their associated meteorological effects (e.g. Cuesta et al., 2009; Solomos et al., 2012; Kalenderski et al., 2013; Prakash et al., 2014).

In this study, we utilize an advanced regional model, WRF--Chem (Grell et al., 2005a), and data from the SAMUM--I field campaign that was conducted from 10 May to 7 June 2006 in Northwest Africa. The main aim of the campaign was to understand the microphysical and optical properties and the radiative effect of dust. Details of the SAMUM--I field experiment, instrumentation and data can be found in Heintzenberg (2008) and Weinzierl et al. (2009) as well as on the SAMUM website (http://samum.tropos.de/). Our research goal was to characterize the size distribution and optical properties of the airborne mineral dust and to improve our understanding of the dynamical mechanisms and meteorological conditions that trigger dust emissions, uplifting and long--range transport in North Africa and over the Atlantic Ocean.

The remainder of the article is organized as follows. Section 2 describes methodology, available observations, and model setup. Section 3 presents results. Section 4 provides a summary and conclusions.

2. Methodology

SAMUM--I was conducted from 10 May to 7 June 2006. During the campaign, horizontal and vertical dust plume structures were sampled (Weinzierl et al., 2009). Based on the dust loading and synoptic conditions over Morocco and north--west Africa, Knippertz (2008) divided the entire campaign period into three phases:

a) Dust phase 1 (DP--1) from 12 -- 15 May 2006.

b) Dust phase 2 (DP--2) from 22 -- 27 May 2006.

c) Dust phase 3 (DP--3) from 31 May -- 7 June 2006.

The DP--2 and DP--3 phases are characterized by aerosols with high atmospheric optical depth, as dust was frequently mobilized by strong winds, dry convection, and density currents during these phases. We therefore focus on the DP--2 and DP--3 periods in this study.

2.1 Observations

Upper air and near--surface station observation data were obtained for the study period from the US National Climatic Data Center (NCDC) and University of Wyoming (http://weather.uwyo.edu/upperair/sounding.html). A list of stations is presented in Table 1. The SAMUM--I airplane meteorological and dust observations were provided by the
Deutsches Zentrum für Luft und Raumfahrt (DLR) team, which conducted the field experiment.

We also utilized Era--Interim (Era--I) reanalysis data produced by the European Center for Medium--Range Weather Forcasts (ECMWF) to test the WRF--Chem model results. Era--I provides a comprehensive set of 3--hourly surface parameters that capture the weather as well as land--surface conditions and 6--hourly upper--air parameters at 37 pressure levels up to 1 hPa on a $0.7^\circ \times 0.7^\circ$ latitude--longitude grid. Details of the dataset can be found in Dee et al. (2011) and on the ECMWF web page (http://apps.ecmwf.int/datasets/data/interim_full_daily/). The spatial distribution of aerosol optical depth (AOD) was obtained from the Moderate Resolution Imaging Spectroradiometer (MODIS) instruments aboard the Terra and Aqua satellites that cover the entire earth surface every 1 to 2 days.

The MODIS instrument provides high radiometric sensitivity (12 bit) in 36 spectral bands ranging in wavelength from 0.4 μm to 14.5 μm. The spatial resolution varies between bands, with two bands at 0.25 km (at nadir views), five bands at 0.50 km, and the remaining 29 bands at 1.0 km. In this study, we utilized MODIS Level 2, Collection 5.1 granule data from the Terra and Aqua platforms. MODIS Level--2 data are derived from Level 1B and produced at a nominal spatial resolution of 10 x 10 km. We use AOD retrievals over land and sea derived from the dark target product (Remer, 2005; Remer, 2008) and the “deep blue” product over bright land surfaces (Hsu et al., 2004) for comparison with the simulated optical properties of aerosols from WRF--Chem.

The directly observed AODs and retrieved aerosol size distributions were obtained from the ground--based remote sensing aerosol robotic network (AERONET) to validate the WRF--Chem aerosol output. AERONET sun photometers provide observations of spectral AOD on up to 8 wavelength channels between 0.340 and 1.640 μm (Holben et al., 1998) and an angular distribution of sky radiance at four wavelengths (0.440, 0.675, 0.870, and 1.020 μm). The maximum AERONET uncertainty in AOD measurements is estimated to be 0.02 with the highest error in the ultraviolet wavelength (Holben et al., 1998; Eck et al., 1999). The calibrated sky radiance measurements typically have an uncertainty of less than 5% (Holben et al., 1998). The AOD data from AERONET are computed at three
quality levels: level--1.0 (unscreened), level 1.5 (cloud screened) and level 2.0 (cloud screened and quality assured). We utilize level 1.5 AOD and level 2.0 data in our analysis. Details of the AERONET data and instrumentation can be found in Holben et al. (1998), Holben et al. (2001), and on the AERONET website (http://aeronet.gsfc.nasa.gov/).

We used various statistical measures such as mean, standard deviation, correlation coefficient, index of agreement (IOA) and root mean squared error (RMSE) to compare model results with observed data. The Pearson’s correlation coefficient describes the co-linearity between observed and predicted values and is defined as:

\[ r = \frac{\sum_{i=1}^{n}(X_i - \bar{X})(Y_i - \bar{Y})}{\sqrt{\sum_{i=1}^{n}(X_i - \bar{X})^2} \sqrt{\sum_{i=1}^{n}(Y_i - \bar{Y})^2}} \]  

where \( X \) and \( Y \) are model prediction and observed values, respectively and \( \bar{X}, \bar{Y} \) are corresponding ensemble averages. The IOA is a dimensionless quantity that determines the degree to which observed values were accurately predicted by the modeled values (Willmott et al., 1985). The value of IOA lies between 0 and 1, where 0 indicates no agreement and 1 corresponds to perfect agreement. The IOA is specifically designed for model performance evaluation and is defined as:

\[ IOA = 1 - \frac{\sum_{i=1}^{n}(X_i - Y_i)^2}{\sum_{i=1}^{n}(|X_i - \bar{Y}| + |Y_i - \bar{Y}|)^2} \]  

The RMSE represents a domain--average difference between observed and model values:

\[ RMSE = [n^{-1} \sum_{i=1}^{n}(X_i - Y_i)^2]^{0.5} \]  

2.2 Model setup

The regional meteorological Weather Research Forecast (WRF) model, developed by the National Center for Atmospheric Research (NCAR) USA, solves fully compressible, non--hydrostatic equations of atmospheric dynamics. The prognostic variables include three--dimensional wind vector \( (u,v,w) \), potential temperature, geopotential, surface pressure, turbulent kinetic energy, water vapor mixing ratio, and cloud water. The model uses a terrain--following, hydrostatic--pressure vertical coordinate
system and horizontal numerical approximation is performed on the Arakawa C--grid (Grell et al., 2005a; Grell et al., 2005b; Skamarock et al., 2005). For advancing in time, the model uses second-- and third--order Runge--Kutta time integration. WRF--Chem extends WRF to include a chemistry module fully coupled with meteorological processes. The module simulates dust emissions, transport, deposition, turbulent and convective mixing as well as atmospheric chemistry and aerosol microphysics (Grell et al., 2005a). In this study, we use WRF--Chem version 3.3.5. It is configured with the Regional Acid Deposition Model 2 (RADM2) photochemical mechanism (Stockwell et al., 1990), the Jerome Fast photolysis scheme (Wild et al., 2000), and the Modal Aerosol Dynamics Model for Europe (MADE) and Secondary Organic Aerosol Model (SORGAM) (Ackermann et al., 1998; Schell et al., 2001). MADE/SORGAM uses the modal approach with three log--normally distributed modes (Aitken, accumulation, and coarse modes) to represent the aerosol size distribution. The aerosol species in MADE/SORGAM are mainly composed of sulfate, nitrate, ammonium, organic matter (OM), black carbon (BC), water, sea salt and mineral dust. We also employed the Goddard Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) dust emission scheme (Ginoux et al., 2001) to calculate dust influx into the atmosphere. GOCART simulates dust emission as a function of the surface wind speed at 10 m altitude above the ground, $U_{10m}$ (Chin et al., 2002). Emission flux, $F_p$, for a specific aerosol size group $p$ is defined as:

$$F_p = \begin{cases} 
CS_p U_{10m}^2 (U_{10m} - U_t) & \text{if } U_{10m} > U_t, \\
0 & \text{otherwise} 
\end{cases},$$  \hspace{1cm} (4)

where $C$ is a dimensional constant coefficient that directly controls the magnitude of the dust emission flux; $S$ is the dimensionless erodibility field taken from Ginoux et al. (2001) with a spatial latitude--longitude resolution of $0.25^\circ \times 0.25^\circ$; $U_t$ is the threshold velocity of wind erosion, which depends on particle size and surface wetness; $s_p$ is a particular particle size mass fraction injected in the atmosphere. Dust is a primary aerosol and does not have a very fine Aitken mode, but comprises accumulation (or fine) and coarse modes.

The dust generation process is very complex. Equation (4) is a simplified empirical approximation of this process and has to be adjusted using available observations to account for different model resolutions and meteorological conditions. For example, coefficient $C$ in Eq. (1) was originally estimated to be equal to 1 mg s$^{-2}$ m$^{-5}$ based
on regional datasets (Ginoux et al., 2001). Zhao et al. (2010) proposed tuning this parameter and, in their calculations, used $C = 0.65$ mg s$^2$ m$^{-5}$ to replicate the consistency of the calculated optical depth with AERONET observations. In our previous simulations (Kalenderski et al., 2013), we used $C = 0.75$ mg s$^2$ m$^{-5}$ to achieve consistency between the simulated and observed AERONET aerosol optical depth at the sites within the study domain. In this study, we simultaneously tuned $C$ and $s_p$ to correctly represent contributions of coarse and accumulation (or fine) modes, respectively. We discuss this adjustment in detail in Section 3.6.

The following physical parameterizations were used to configure the model simulations: the Lin microphysics scheme; the Rapid Radiative Transfer Model (RRTM) both for longwave (LW) and shortwave (SW) radiation; the Mellor--Yamada--Janjic (MYJ) boundary layer scheme; the Noah land surface model; and Grell cumulus parameterization. We used the USGS land use and vegetation data (24 categories) to define the lower boundary conditions (Wang et al., 2014).

Lateral boundary and initial conditions for meteorological fields were provided by the National Centers for Environmental Prediction's (NCEP) global analysis (FNL). We used NCEP's daily global sea surface temperature (SST) analysis (RTG_SST_HR) to update SST every six hours.

Initial and boundary conditions for aerosol-- and gas--phase species were from the default WRF--Chem profiles, which were obtained from various field studies to capture clean atmosphere, maritime, and mid--latitude conditions (McKeen et al., 2002). Anthropogenic emissions were provided by the REanalysis of the TROpospheric (RETRO) chemical composition inventories (http://retro-archive.iek.fz-juelich.de/) and by the Emission Database for Global Atmospheric Research (EDGAR) (http://edgar.jrc.ec.europa.eu). Biomass burning emissions were obtained from the Global Fire Emissions Database, Version 2 (GFEDv2.1) (Randerson et al., 2007) with a 1° x 1° latitude--longitude spatial resolution and 8--day temporal resolution. All emission inventories were preprocessed by the PREP--CHEM--SRC v1 emissions preprocessor (Freitas et al., 2011). Biogenic emissions were not included in this study. We assumed zero initial dust concentrations and no dust inflow through the lateral boundaries.

2.3 Choice of a domain
To cover the entire observation area and account for processes that are responsible for the development of dust plumes during the entire experiment, the WRF--Chem model was run in the spatial domain from 14.62°N to 36.10°N in latitude and 17.56°W to 12.56°E in longitude with 5 km horizontal grid spacing. In the calculations, we used 550 grid points in the west--east direction and 484 grid points in the south--north direction, as well as 51 vertical levels with the top of the model domain at the 10 hPa level. The domain covers Northwestern Africa including the southern Sahara, Morocco and parts of the Atlantic Ocean and Mediterranean Sea (Figure 1a). As a sensitivity study, we also ran WRF--Chem with the identical settings but in a smaller domain from 10.5°W to 1°W in longitude and from 26°N to 36°N in latitude with 181x181 grid points (Figure 1b). Below, we refer to the simulations in the large domain (LD) as WRF--LD, and in the small domain (SD) as WRF-SD.

The choice of the domain appears to be a very important issue for correctly simulating dust plumes during the SAMUM--I campaign. The following requirements must be considered:

a) The domain should be as small as possible to make calculation more feasible.
b) The domain has to be large enough to allow small--scale features to develop inside the domain despite coarse boundary conditions, insuring correct simulation of meteorological fields with a fine spatial resolution.
c) The domain should include all important dust sources to make the emission--tuning process (i.e., adjusting $C$ in equation (4)) physically meaningful (Kalenderski et al., 2013).

The third requirement appears to be crucially important to our case. First, we ran simulations in SD and were able to reproduce meteorological processes in the SAMUM--I area relatively well. However, the AOD in comparison with AERONET observations was unrealistically small. When we ran the simulation in LD with the same default settings, the AOD increased by about three times. Further increase of the domain did not affect the AOD significantly. Figure 2 compares AOD from two simulations, WRF--LD (blue solid line) and WRF--SD (green solid line), with that from four Aeronet sites (red dots) within the SAMUM--I region: Ouarzazate, Ras EL Ain, Sada, and Tamanrasset (see Figure 1). The simulations were conducted with the default $s_p$ values and empirical constant $C = 0.65$ mg s$^2$ m$^{-5}$, which is consistent with the value suggested by Zhao et al. (2010) for the Sahara region. The simulated AOD from the WRF--LD run was able to reproduce
the magnitude and temporal evolution of observed AOD at all four Aeronet sites, while WRF--SD simulations severely underestimated AOD throughout the simulation period. Figure 3 shows the dust emission areas (A, B, C, D, E, F) in LD that were activated during the WRF--LD simulations. These areas are discussed in detail below. The color shading indicates the average rate of dust emission. The SD, bordered by a black dashed line, includes only source region D and part of source region B but is missing all major emission areas. The WRF--SD simulation therefore severely underestimated AOD. Thus, we further base our analysis on the WRF--LD simulations.

3. Results

The WRF--Chem model with the above settings was integrated for 21 days from 20 May to 9 June 2006 through most of the duration of the SAMUM--I experiment. The analysis of the simulations is discussed below.

3.1 Meteorology

The SAMUM--I study area was located between 29.5\(^\circ\)N to 31.5\(^\circ\)N and 9.0\(^\circ\)W to 4.0\(^\circ\)W (Figure 1). Deep convection frequently develops over the high terrain during the afternoon hours, leading to precipitation and thunderstorms in the evening. During the study period, northwestern Africa was affected by weather systems from the north. The meteorological analysis of this period was presented by Knippertz et al. (2008). To evaluate the performance of WRF--Chem during the entire simulation period, we first compared the mean Era--I (http://apps.ecmwf.int/datasets/) fields for the period from 22 May to 7 June 2006 with the WRF--Chem output for the same period. Both the correlation coefficient and index of agreement between 2 m temperatures (T\(_{2m}\)) in WRF--Chem and Era--I were greater than 0.97 with an RMSE of 1.57 K. The domain average time means (T\(_{2m}\)) were 28.0 °C and 29.0 °C from WRF--Chem and Era--I, respectively, because of higher terrain and associated orographic cooling captured in WRF--Chem in comparison with the lower resolution Era--I data. Over most of the Sahara, WRF--Chem is also slightly colder than in Era--I, however the model over--estimates T\(_{2m}\) in the northern foothills of the Atlas mountain range and further north over the Moroccan plains. The distribution of wind at 10 m (U\(_{10m}\)) is consistent with the topographical features of the model domain. The correlation coefficient between WRF--Chem and Era--I winds is greater than 0.50 with an Index of Agreement (IOA) of 0.74 and an RMSE of ~ 1.39 m s\(^{-1}\). The domain average time mean, WRF--Chem U\(_{10m}\), is 1 m s\(^{-1}\) stronger (5.5 m s\(^{-1}\)) than that from Era--I. The WRF--
Chem model simulates higher wind speeds in the vicinity of elevated terrain, whereas over the flat desert, the mean difference between the model and observations is small.

To evaluate how WRF--Chem performs on fine temporal and spatial scales, the meteorological characteristics from the WRF--Chem simulations were compared with those measured along the flight track of the research aircraft. During the SAMUM--I campaign, a DLR Falcon research aircraft was used for 17 research flights from 18 May to 7 June 2006. The durations of these flights varied from 37 minutes to more than 3 hours. Presented here are the plots for the longest flight that took off on 4 June 2006 (Figure 4). Figure 4 shows observed (red) and simulated (blue) wind speed, wind direction, air temperature and specific humidity along the flight track shown in Figure 1b. The correlation coefficient between simulated and observed air temperature is 0.99. The basic meteorological characteristics are reproduced more accurately at higher elevations where there is less meteorological noise from high--frequency boundary layer processes. The agreement for winds is not as good as that for temperature, especially in the lower troposphere where penetrating convection causes variable winds. However, the overall IOA exceeds 0.90. The RMSE for wind speed is 2.89 m s\(^{-1}\), which is two times smaller than the wind standard deviation (4.67 m s\(^{-1}\)). At altitudes higher than 6 km, the predicted wind speed reaches 17.0 m s\(^{-1}\) and is higher than the observed wind speed that reaches 13.0 m s\(^{-1}\). The IOA for both U and V components is around 0.88, and the correlation coefficient and RMSE are 0.89 and 3.77 m s\(^{-1}\), respectively. The observed standard deviations for both components of horizontal wind speed are twice of RMSE, which indicates the ability of WRF--Chem to reliably reproduce atmospheric flow at high spatial resolution. WRF--Chem generally over--estimates water vapor at high elevations above 6 km (see Figure 4d); at lower elevations, however, WRF--Chem shows better capability in predicting water vapor.

To further evaluate the simulated fields, we made comparisons with the available upper atmosphere soundings. The radiosonde data (potential temperature, specific humidity, wind speed, and wind direction) at 1200 UTC on 4 June 2006 from three stations in the Sahara Atlas region (In--Salah, Bechar and Tamanrasset, see Table 1 and Figure 1 for station locations) were obtained from University of Wyoming (Figure 5) ([http://weather.uwyo.edu/upperair/sounding.html](http://weather.uwyo.edu/upperair/sounding.html)). The stations are characterized by quite different meteorological conditions. The temperature at Tamanrasset is up to 12 degrees higher than at In--Salah and Bechar. This difference is also reflected in the structure of the
planetary boundary layer (PBL). The PBL height at In--Salah and Bechar is lower (< 4000 m) than in Tamanrasset (~5000), because Tamanrasset is further south in the desert, where higher vertical instability leads to stronger turbulence in the boundary layer. The observations at In--Salah show a well--mixed layer in the first 2 km and a stable layer above with a capping inversion at 4 km. The simulated specific humidity is less consistent with observations, as could be expected, but is still quite reasonable. The reduction in moisture above the PBL in the simulations indicates cloud formation at this level (Figure 5b).

The strongest simulated and observed winds in the PBL are recorded at In--Salah with the wind speed exceeding 12 m s$^{-1}$ from the NE direction (Figure 5g). The weakest PBL winds are recorded at Bechar (<5 m s$^{-1}$) while at Tamanrasset, we see a moderate westerly flow of ~7 m s$^{-1}$. During daytime, the north--easterly flow prevails. Tamanrasset is located south of Ahaggar Mountains, i.e., on the lee--side of the north--easterly flow. The existence of a lee--side cyclone to the south of Ahaggar results in westerly flow at Tamanrasset. Wind speed at the top of the boundary layer reaches 15 m s$^{-1}$ at Tamanrasset and Bechar, while a relatively weaker northerly wind of 12 m s$^{-1}$ is recorded at In--Salah. The strong wind shear at 800 m above sea level (asl), observed at In--Salah, is indicative of the existence of a low level jet in this region.

3.2 Dust sources in the SAMUM--I region

North Africa is considered to be the most active dust source in the world (Knippertz et al., 2008). A variety of meteorological processes cause dust emissions over this region. They range in scale from synoptic to local and comprise disturbances from subtropical troughs, the Saharan Heat Low, Haboobs, and dust devils. Interaction of multi-scale subtropical disturbances with continental--to--local scale Saharan meteorological processes further complicates the picture. The SAMUM--I period is generally characterized by interactions between the West African Heat Low (HL), upper level circulation in the subtropics, and the Intertropical Convergence Zone (ITCZ) (Knippertz, 2008). Convective cold pools and the breakdown of nocturnal low--level jets (NLLJs) are key regional--scale meteorological mechanisms responsible for dust emissions over Northwestern Africa during summer (Heinold et al., 2013).

From 22 May to 7 June 2006, six emission areas were activated. They are shown in Figure 3 (areas $A$, $B$, $C$, $D$, $E$, $F$), which depicts average dust emissions for the entire
period from WRF--Chem simulations.

Area A is the region of Chotts. It comprises a large area of ephemeral lakes that stretches from the low lands of Tunisia to the Atlas Mountains of Northern Algeria and is a well--known dust source region (Prospero et al., 2002; Mahowald et al., 2003; Ginoux et al., 2012). The Chotts area is in the rain shadow of the Algerian Atlas mountains. Precipitation in the Chotts area is less than 100 mm per year. According to Prospero et al. (2002), the most intensive dust emission occurs in an area centered at (7.5° E, 33.5° N) during April – May and it could extend to August – September. The model captures significant dust emissions from this area on 24, 25, 26, 30 May, and 1 June. The strong wind gusts that lead to emissions in the Chotts area could be attributed to lee cyclones that develop to the south of the Atlas Mountain range (Knippertz et al., 2008).

Area B is the topographical low region to the west of the Ahaggar Mountains comprising hydro (ephemeral lakes) and non--hydro natural dust sources (Ginoux et al., 2012). These lakes formed because of weathering and runoff from the mountains (Prospero et al., 2002). The West African Heat Low controls the circulation in this area. During the simulation period, this emission area was active every day except for 27 and 29 May. Most of the high emission events occurred between 0900 and 1200 UTC, when strong solar heating of the ground leads to convection and rapid growth of the Convective Boundary Layer (CBL) that allows downward momentum transport from the Low Level Jets (LLJs) resulting in strong and gusty winds of 10--15 m s$^{-1}$ (Knippertz et al., 2008).

Area C is located in the Tiris Zemmour region in northern Mauretania and the coastal plains of southern Morocco and comprises natural and anthropogenic dust sources and ephemeral lakes (Ginoux et al., 2012). Dust emissions in Tiris Zemmour were simulated for 22--25 May. Most of the dust events occurred between 1200 and 1800 UTC. Tiris Zemmour has a high frequency of dust emission episodes and acts as the main source of fine--grained material in this region (Oberle, 2001; Varga, 2012).

Area D is a region of density currents (DC) (Knippertz et al., 2007; Knippertz et al., 2008). In the Sahara--Atlas region, evaporative cooling associated with deep moist convection is an important meteorological process responsible for the formation of density currents while the high Atlas mountain range serves as a trigger to the moist convection in this region. These systems frequently produce haboobs (Sutton, 1925; Membery, 1985; Knippertz et al., 2007). Knippertz et al. (2007) reported seven major DCs in the study region during the SAMUM--I campaign. The WRF--Chem model successfully captures the
DC activity and the modeled meteorology is in agreement with the available observations.

Area E is in the Senegal River basin where dust generates from overgrazing reactivated playas and ephemeral lakes (Gill, 1996; Ginoux et al., 2012).

Area F is in the Upper Niger River basin where dust generated from alluvial sediments mostly results from overgrazing and sediments from seasonal flooding of the Niger river (Mulitza et al., 2010).

The model simulations show that most of the dust was emitted from areas A, B, and C during the simulation period. Areas E and F are remote with respect to the flight tracks and relatively unimportant for our analysis of the flight observations.

3.3 Dust generation and transport

During the SAMUM--I period, dust was generated and transported within the domain. To demonstrate the model’s capability to realistically simulate the dust plume, we compare in Figure 6 the simulated and observed MODIS aerosol optical depth. The spatial-temporal structure of the simulated and observed dust plumes compare well despite that vast areas, shown in gray, in the MODIS observations are hidden by clouds. The aerosol plumes are quite extensive with optical depths exceeding 1. The high AOD on 25 and 26 May is attributed to emissions from areas A, B and F (Figure 6 b and d). The model simulates strong dust emissions from area A from early morning to late evening. Dust in area A and B is mobilized due to the presence of a sub-tropical cut-off low centered over northwest Africa accompanied with lee cyclone to the south of the Atlas mountain range (not shown here) that causes strong northeasterly winds near the surface in the Chotts region. In area F, dust emissions occurred during the mornings when the stable nocturnal boundary layer broke down and the LLJ momentum mixed downwards, resulting in strong surface winds and higher dust emissions in the dust source regions. The MODIS AOD (Figures 6 a and c) is consistent with the model results. On 27 May, there are no significant emissions from any of the dust source regions and most of the simulated AODs indicate remnants of the previous two days in the elevated dust layer (Figure 6f).

On 1 June the model simulates strong dust emission flux in the Chotts region (Area A) under northeasterly winds from morning to evening. The northeasterly flow results from interaction of upper level circulation with lee cyclone to the south of the Atlas mountain range. These emissions are reflected in the AOD for 1 June (Figure 6h). The main dust source regions on 3 June remain in areas A and F. On this day, emissions mostly
occur between 0800 and 1100 UTC. Analysis of the wind field (not shown here) indicates that these emissions are caused by low level jet (LLJ). The dust–laden air rapidly moves northward. This is reflected in the AOD for 3 June in both MODIS and WRF–Chem (Figures 6i and j). In the Sahelian region (bottom of the domain at 15°N), the dust AOD is enhanced by biomass burning aerosols transported from the Niger River and the Senegal River basins. Emission from biomass burning in the dry season is an important (but not dominant) source of airborne aerosols in this region (Bond et al., 2004; Haywood et al., 2008).

3.4 Saharan Air Layer formation

Long–range cross–Atlantic transport of Saharan dust in the Saharan Air Layer (SAL), located above the marine boundary layer, is well documented (Prospero et al., 2002). Here, we take advantage of SAMUM–I observations to understand the formation of SAL in the high Atlas mountain region. A 5–km deep dust layer is found on 3–4 June both to the south and north of the Atlas Mountains. However, northward dust transport occurs only on 6–7 June. The period from 5–7 June is characterized by the upper level cut–off low over the Atlantic Ocean and a weak high over most of northwest Africa. Interaction of this upper level circulation with lee cyclone south of the Atlas mountain range results in strong near surface southerly flow. During the mid–morning on 5 and 6 June, the model simulates two medium level dust outbreaks. The first dust outbreak on 5 June is initiated by emissions from the hydro–dust source area A (Figure 3), under strong southeasterly/southerly winds. The second dust outbreak on 6 June is driven by emissions from source area B (Figure 3) to the west of the Ahaggar Mountains (25.5°N, 3.0°E) under strong northeasterly winds. The uplifted dust is transported to the north of the high Atlas mountains by southerly winds on 6–7 June.

To better demonstrate the formation of SAL, Figure 7 shows meteorological characteristics and the dust concentration in two latitude–height cross sections at longitudes of 7.79° W (AA’ cross section, Figures 7 a,c,e) and 6.3° W (BB’ cross section, Figures 7 b,d,f) on 7 June at 4:00 UTC, when the land breeze is still strong, and at 15:00 UTC, when sea breeze develops. The AA’ cross section contains a mountainous area with elevation above 3 km. The BB’ cross section covers the area closer to the ocean behind the mountains where the elevation does not exceed 1 km.

The solid lines in Figures 7a and 7b depict the height of the Residual Layer (RL)
calculated as the maximum height of the Planetary Boundary Layer (PBL) in the previous diurnal cycle. The dashed lines show the depth of the shallow nocturnal boundary layer (NBL). The vertical velocity contours in Figures 7a and c show the regions of vertical convection that could mix dust above PBL. The wind barbs in Figure 7b show the predominant southerly flow in the lower troposphere and the land breeze near the surface. The CBL, which by 15:00 UTC has already consumed the RL, reaches the height of PBL.

In the morning (Figure 7 a,b) before crossing the High Atlas Mountains, most dust is confined to the 4--km high RL over land. The orographic lifting on the southern side of the Atlas Mountains and the resulting hydraulic jump, as well as the convection developed on the slopes (vertical velocity is shown by black [upward] and gray [downward] contours in Figure 7b) cause some of the dust--laden air to escape the RL (Figure 7 a,b).

At night or in early morning, most dust is transported northward in the RL. Strong southerly winds rapidly advect the dust--laden air mass towards the ocean where, in the region of the strong gradient of the RL height and the weak night inversion, it relatively easily becomes part of the free troposphere.

The growth of the CBL in the morning entrains aerosols in the RL over the land but does not affect the aerosol plume over the ocean, as the marine PBL is very thin. In the afternoon, the fully developed PBL at 15:00 UTC is much higher over land (about 5 km) than over sea (200--300 m) with a steep gradient of PBL height and all meteorological characteristics in a transition zone between land and sea (Figure 7 d--f). A strong convection develops on the slopes of the Atlas Mountains (Figure 7c). This convection mixes the dust--laden layer from the PBL into the free troposphere, eroding the capping inversion.

The sea breeze strengthens by 15:00 UTC (Figure 7 e,f). Its propagation is well marked by transport of moisture and potential temperature. The breeze front reaches 31°N in Figure 7e and 33.3°N in Figure 7f. The convection in the breeze front could also be a powerful mechanism for mixing dust in the free troposphere within the land/sea transition zone.

3.4 Interaction of Sea Breezes with Continental Dust Outflow

The SAMUM--I observations show the formation of the elevated aerosol layer over the sea that is clearly detached from the surface. We were able to reproduce this effect in
our simulations and found that the interaction of the coastal outflow with the breeze plays an important role. Sea breezes are known for trapping pollutants emitted from the surface layer of the coastal regions (e.g. McKendry, 1989; Miller et al., 2003). WRF--Chem simulates sea breezes from 10:00 to 19:00 UTC at the Moroccan coast of the Atlantic Ocean. Figures 7e and 7f show the vertical cross--sections of land and sea at 15:00 UTC. The potential temperature indicates the sea--breeze inflow layer depth and its horizontal extent while the water vapor mixing ratio shows the difference in moisture content between the cold dense marine air and the warm dry land air. The average height of the inland branch of the simulated sea breeze ranges from 0.7 km to 1 km above ground level (agl), while the wind speed within the sea breeze flow varies from 8 to 13 m s\(^{-1}\). The inland extent of the sea breeze reaches 100--150 km, while the seaward extent of the sea breeze reaches about 200 km (Figures 7 e,f). The water vapor mixing ratio in the sea--breeze flow ranges from 15 g kg\(^{-1}\) in the first 150 m over the sea to 7.5 g kg\(^{-1}\) over land. Strong updrafts forms due to convergence between the sea--breeze air and land air along the frontal zone. The collision of the two air masses (moving from sea and from land) results in the formation of the sea breeze head that grew from 2.0 to 3.5 km high, nearing the height of the PBL in the transition zone and generating strong updrafts reaching 1.5 m s\(^{-1}\). The collision of cold and warm air enhances turbulence and mixing of dust in a vertical column (Figures 7 c,d,e,f).

The northward moving dust--laden land air mass is warmer than the landward moving sea breeze. The dust is well mixed during afternoon hours within the warmer PBL over land. The cold and moist marine air rapidly propagates inland. Behind the sea--breeze head, the warm air over--rides the marine air and forms a strong inversion above the cold and moist landward flow. Due to this inversion, the relatively warm dusty air is unable to mix downward. The air that is uplifted in the sea breeze convergence zone rapidly moves seawards with strong northward prevailing winds. The scattered downdrafts could be seen in few tens of kilometers from the shore, but due to the inversion over the sea--breeze flow, they were unlikely to reach the surface layer. Therefore, most of the dust--laden air moves out of the domain without being involved in the landward branch of the sea--breeze flow. The breeze effectively cleans the lower part of the dust plume over the sea detaching the aerosol layer from the sea surface as observed (see Figure 7d).

To more clearly demonstrate this point we present in Figure 8 the vertical profiles of dust concentration and meteorological parameters at 6.3°W, 33.5°N (this position is
indicated in Figure 7f). Figure 8 shows formation of the vertical structure of the sea breeze
head and its interaction with the continental outflow. The location is approximately 150
km inland from the coastline with a land elevation of 500 m above sea level (asl). Figure
8a shows how a CBL grows and erodes the residual layer above it. By 1300 UTC, CBL
reaches its maximum at 5.0 km asl (4.5 km agl). At 1500 UTC a 0.7 km deep surface layer
forms. The air temperature in the surface layer drops by 3°K, the wind direction is onshore
(Figure 8b), and the moisture content reaches its maximum value of 10 g kg\(^{-1}\) (Figure 8d).
Abrupt changes in the meteorological characteristics indicate the arrival of a cold, moist
sea breeze front. Figure 8b shows the Ekman clockwise rotation of the surface winds
across the diurnal cycle due to the Coriolis force that induces an oscillating alongshore
wind component that lags the onshore--offshore component by 6 hours or 1/4 cycle
(Adams, 1997; Miller et al., 2003). The 11:00 UTC profile shows the sea--breeze flow
from the southwest direction. At 13:00 UTC, the wind rotates to its right and the westerly
sea breeze blows towards the land. At the sea breeze peak time (15:00 UTC), the wind
further rotates and comes at a right angle (northwest) to the coastline.

A distinct feature of the dust profiles in Figures 8 e,f (from 11:00 UTC to 15:00
UTC) is the gradual decrease of the dust mixing ratio in the surface layer. Before the
arrival of the sea breeze, the dust concentration reaches 400 µg m\(^{-3}\) at 11:00 UTC in the
surface layer. As the clean cold and dense marine air propagates inland, a warm and dusty
land air glides over the sea--breeze flow. The inversion over the sea breeze flow inhibits
free downward mixing between the two air masses. This is evident from the vertical
profiles of total and fine dust concentrations at 15:00 UTC that show a lower dust
concentration within the surface layer, whereas above the inversion, the dust
concentrations at 15:00 UTC are practically equal to those at 13:00 UTC.

3.5 Dust partitioning between the PBL and the Free Troposphere

Over land, dust particles are first entrained into the surface layer and then during
the day they mix in the CBL that, at about 15:00--16:00 UTC, reaches its maximum height
and consumes the RL. At night, the PBL collapses and most of airborne dust remains in the
RL. However, in the morning hours, the growing CBL rapidly erodes the RL and entrains
the previous day’s uplifted dust. Therefore, at any given time most of dust mass stays in
the PBL, where PBL= NBL/CBL+RL. The height of the PBL over land is 4--5 km in our
simulations. Because of turbulent mixing, the lifetime of dust particles in the PBL is
relatively short, while particles above the top of the PBL are disconnected from the surface and could stay in the free troposphere for extended times, which would be sufficient for long--range transport. To better understand partitioning of the dust mass between the PBL and the free troposphere, we calculated the average dust loadings above and below the top of the PBL separately over land and over sea areas in the LD. The top of the RL is defined as a maximum height of the PBL in the preceding diurnal cycle. Figure 9 shows these loadings as function of time in g m\(^{-2}\). Red and blue lines with filled circles correspond to dust loadings above the PBL over land and sea respectively, whereas thin red and blue solid lines correspond to dust loading within the PBL over land and sea, respectively. The maximum loading over land is about 1.2 g m\(^{-2}\), and over the ocean it is 0.5 g m\(^{-2}\). Over land, about 10--15% of dust is in the free troposphere. Over the ocean, the dust loading in the free troposphere (i.e. in SAL) often exceeds that in the shallow marine boundary layer. This is in good agreement with the analysis conducted in the previous section.

3.6 Evaluation of dust size distribution

Optical properties, life time, and radiative effects of dust particles are critically dependent on particle size and composition (Sokolik and Toon, 1999; Dubovik et al., 2006). The size range of particle radii spreads across two orders of magnitude, roughly from 0.1 to 10 microns and the Mie solution predicts that the extinction efficiencies of individual coarse--to--fine particles range by four orders of magnitude, because efficiencies are proportional to squared particle radii. Nevertheless, small particles are optically more efficient per unit mass because the number of particles is inversely proportional to the cubed particle radius. Large particles have shorter lifetimes in the atmosphere because of their higher deposition velocities. However, based on the measurements conducted during the ACE--2 campaign of 1997, Otto et al. (2007b) found that the contribution of coarse particles in atmospheric heating rates and their effect on the top of the atmosphere and surface radiation balances prevails over those of fine particles. Weinzierl et al. (2009) confirmed this observation under SAMUM--I.

We evaluate aerosol size distributions in the framework of the SAMUM--I case study, using model output and available airplane and ground--based observations. In our simulations, we use the MADE/SORGAM modal aerosol model (Ackermann et al., 1998; Schell et al., 2001) that approximates the dust size distribution using the two--mode lognormal size distributions to represent coarse and fine (accumulation) modes:
\[
\frac{dN}{dD} = \frac{N_{\text{fine}}}{D \ln \sigma_{\text{fine}} \sqrt{2\pi}} e^{\frac{-(\ln(D) - \ln(D_{\text{fine}}))^2}{2 \ln^2 \sigma_{\text{fine}}}} + \frac{N_{\text{coarse}}}{D \ln \sigma_{\text{coarse}} \sqrt{2\pi}} e^{\frac{-(\ln(D) - \ln(D_{\text{coarse}}))^2}{2 \ln^2 \sigma_{\text{coarse}}}}
\]  

(5)

where \(N_{\text{fine}}\) and \(N_{\text{coarse}}\) are total number concentrations, \(D_{\text{fine}}\) and \(D_{\text{coarse}}\) are modal diameters, and \(\sigma_{\text{fine}}\) and \(\sigma_{\text{coarse}}\) are the geometric standard deviations of fine and coarse modes, respectively. The MADE/SORGAM model predicts the spatial--temporal evolution of the modal diameter and number concentration for each mode, however the geometric standard deviations for both modes are kept fixed:

\[
\sigma_{\text{fine}} = 2 \quad \text{and} \quad \sigma_{\text{coarse}} = 2.2
\]  

(6)

The number and size distribution of dust particles entering the atmosphere from surface in equation (4) depend on surface wind, as well as soil type and its physical conditions, e.g., moisture content and surface roughness (Kok, 2011a; Kok, 2011b; Kok et al., 2012). The size of particles is especially important because it defines particle lifetime and affects aerosol optical properties. In the standard WRF--CHEM setup, the emitted dust mass flux in equation (4) is distributed in two modes (coarse and fine), with modal diameters of \(D_{\text{fine}} = 0.6 \, \mu m\) and \(D_{\text{coarse}} = 6 \, \mu m\), mode mass fractions \(s_{\text{fine}} = 0.07\) and \(s_{\text{coarse}} = 1 - s_{\text{fine}} = 0.93\) (\(s_p = 0\) for nucleation or Aitken mode as dust is a primary aerosol), and geometric standard deviations \(\sigma_{\text{fine}}\) and \(\sigma_{\text{coarse}}\) as given by equation (6). In the atmosphere dust particles are advected by winds, mixed by turbulence, microphysically interact with aerosols and water, experience chemical transformations, and deposit forming an airborne aerosol particle distribution that is different from that coming from surface (Zender et al., 2003; Shao, 2008). The microphysical processes have small sub--grid scales and at present cannot be simulated from the first principles. To estimate the associated “microphysical” uncertainties Zhao et al. (2013) conducted WRF--Chem quasi--global simulations with tuned meteorology using bin MOSAIC and modal MADE/SORGAM microphysical dust size parameterizations. They found that the different microphysical models could result in significant differences in aerosol size distribution. Thus, model predictions have to be reconciled with available observations.

In our preliminary WRF--LD calculations with the default mode mass fraction setup and \(C = 0.65 \, \text{mg} \, \text{s}^2 \, \text{m}^{-5}\), the simulated visible total aerosol optical depth
compares well with AERONET observations at 0.5 µm (see Figure 2) at all four AERONET sites. However, the model with standard settings incorrectly represents contributions of the fine and coarse modes in the total optical depth. It overestimates the optical depth of the fine mode and underestimates the optical depth of the coarse mode (not shown) in comparison with the AERONET Spectral Deconvolution Algorithm (SDA) Level 2.0 retrieval available at one AERONET station at Ouarzazate. SDA provides column integrated optical depth for the coarse and fine modes separately showing that in observations the optical depth of coarse mode exceeds that of the fine mode (see Figure 10) in agreement with findings in Ryde et al. (2013) and Otto et al. (2007a). To overcome this deficiency and to reproduce the observed contributions of coarse and fine modes in our simulations we decreased in equation (4) the mass fraction for a fine mode by a factor of seven, redirecting the residual mass into a coarse mode, and increased C slightly to keep the total optical depth as in the AERONET observations. Kok (2011a), consistent with our findings, mentioned that the current parameterizations overestimate the fine mode emission flux by an order of magnitude. With the modified model settings:

\[ S_{\text{fine}} = 0.01; \ S_{\text{coarse}} = 0.99; \ C = 0.75 \times 10^{-9} \ \text{kg} \ \text{s}^{-2} \ \text{m}^{-5}, \]  

we were able to obtain a qualitative and quantitative agreement between simulated and observed optical depths for coarse and fine modes (Figure 10). In Figure 10 the coarse mode optical depth (blue) in both simulations and observations is twice as large in comparison with the fine mode optical depth (red). The simulated (solid line) and observed (filled circles) AODs are in good agreement for the entire period of simulations.

Figure 11 shows the AERONET Level 2.0 (red) at 07:07:41 UTC and simulated at 07:00 UTC (fine mode – green, coarse mode – black, and total – blue) column--integrated (for the entire depth of the atmosphere) instantaneous dust size distributions two hours prior to the research flight 060604a on 4 June 2006. The observed and simulated distributions are in good agreement. WRF--CHEM underestimates the numbers of fine (0.1--0.2 µm) and coarse (3--10 µm) particles, and is fairly close to the observations in the range of particle diameters around 0.4--0.6 µm. These particles produce most of the radiative effect in visible range and therefore are well constrained by the 0.5 µm aerosol optical depth that was tuned to be close to that observed by AERONET. The discrepancy for small particle diameters could be explained by the fact that AERONET retrievals tend
to overestimate the number of small particles (McConnell et al., 2008). However, the discrepancy for large particles is probably because of the model’s deficiency.

To further compare aerosol characteristics along the DLR Falcon flight tracks, we analyzed twelve flights during the simulation period. While all of them capture similar general behaviors and compare well with WRF--CHEM simulations, flight 060604a was chosen for demonstration due to its extended duration and the richest distribution of sampled altitudes. The complete description of the flights is given in Weinzierl et al. (2009).

Flight 060604a took off on 4 June at 09:18 UTC from Casablanca airport located near the Atlantic coast, crossed over the Atlas Mountains, continued to the Ouazazate site, then moved to Zagora, and returned back to Casablanca at 12:34 UTC. Figure 12 shows the flight altitudes as a function of time, the land elevation at the location of the airplane, as well as the PBL height as defined above.

Instrumental observations of aerosol size distributions on the airplane during the SAMUM--I campaign were conducted by a set of instruments including a Forward Scattering Spectrometer Probe (FSSP--300, FSSP--100), a Passive Cavity Spectrometer Probe (PCASP), a set of condensation particle counters (CPC/CPSA) operated at different lower cut--off diameters, a Grimm Optical Particle Counter (Grimm OPC), and a Differential Mobility Analyzer (DMA) (Weinzierl et al., 2009). For use in models, Weinzierl et al. (2009) also approximated the observed aerosol size distributions by four lognormal modes and reported their parameters. These distributions are listed in Table 4 of Weinzierl et al. (2009). We have used that information to compare the simulated and observed aerosol size distributions along those time intervals of the flight track in Figure 12, where the height is constant (see Figure 13). All chosen heights are within the PBL. For all six time intervals in Figure 13, the simulated and observed size distributions compare well for most optically active (for visible wave lengths) aerosol diameters, 0.3 -- 1 µm. However, the model overestimates the number of very fine particles (contrary to what was observed in comparison with AERONET) and underestimates the number of large particles. We have to mention that the MADE/SORGAM is a modal aerosol model and, by design, is unable to account for the giant particles of 30--50 µm diameter observed by Weinzierl et al. (2009), as it assumes that particles follow a two--mode log--normal size distribution and fixes the width of the distributions.

4.0 Summary and conclusions
Dust aerosols are the most abundant particulate constituent in the earth’s atmosphere. In addition to its regional/local weather, societal, and health effects, dust impacts climate on a global scale. Dust particles are capable of travelling thousands of kilometers across continents and oceans. Dust emission and transport mechanisms, interaction with clouds and other atmospheric species, and the effect of dust on radiation transport are of continuous interest. In this study, we used the WRF--Chem model along with observational data from SAMUM--I to investigate dust generation and transport from the Northwest Sahara towards the North Atlantic Ocean and Mediterranean Sea. The model simulates the background meteorological fields and meteorological phenomena relatively well. It captures complex topographic and land surface effects. The model simulates major dust outbreaks observed during the simulation period. The spatial distribution and magnitude of AOD are also well captured. In the model simulations, more than 15% of aerosol mass is mixed up above the PBL by dry convection, lee mountain effects, haboobs, and by mixing in the breeze front. The orographic lifting and interaction of the sea breeze with the continental outflow are key processes leading to formation of the dust--laden SAL above the marine boundary layer. We were able to tune the model to produce the observed ratio of AODs generated by fine and coarse dust modes. The model’s predicted aerosol number density compares well with the AERONET retrievals and SAMUM--I airplane observations in the most optically important 0.5 µm diameter range as the model was tuned to AERONET’s visible optical depth. However, the computationally efficient MADE/SORGAM modal microphysical model is unable to correctly predict the numbers of very small and very large particles.

The main results of the study could be formulated as follows:

- The model produces meteorological fields that compare well with observations for the simulation period both across the entire domain, in station locations, and along flight tracks.

- WRF--CHEM captures major dust outbreaks during the SAMUM--I campaign as verified by observation data. The size of the calculation domain has to be large enough to include all major dust source areas activated during the study period. Simulations conducted in a smaller domain that include only haboob dust sources produced less than 30% of observed optical depth, suggesting that haboobs were
responsible for 1/3 of dust generation in the simulation domain during the SAMUM--I period.

- The model produces realistic spatial--temporal distributions of aerosol optical depth compared with satellite observations and the AERONET retrievals, as well as matches individual contributions of the coarse and fine modes into a total visible optical depth.

- It appears that the orographic lifting of dust in the lee shadow in the Atlas Mountains, convection on the mountain slopes and at the breeze front, as well as cross--inversion mixing in the land/sea transition zone are the most effective mechanisms of dust penetration into the free troposphere. Interaction of a continental outflow with a sea breeze is a key process that leads to formation of a surface--detached dust plume over a marine boundary layer.

- WRF--CHEM predicts that the fraction of dust above the top of a residual layer over land is about 15%. This fraction is more than 50% over the Atlantic Ocean.

- WRF--CHEM, both in comparison with AERONET retrievals and SAMUM--I instrumental observations, predicts well the number density of the most optically active particles with diameters of 0.4--0.6 µm. However, its simplified microphysical model is unable to correctly reproduce the tails of the dust size distribution, overestimating the number--density of small particles and underestimating the number--density of large particles. This suggests that more sophisticated multi--mode or bin aerosol microphysical models should be employed in future studies.

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modeling dust mass balance and radiative forcing from size parameterization.
Table 1: Names and the lat/lon locations of stations used in the analysis.

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Figure 1

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(34)
Figure 2

(a) Ouarzazate (OZT)
(b) Ras_El_Ain (RAS)
(c) Sadaa (SAD)
(d) Tamanrasset (TAM)
Figure 3
Figure 4
Figure 5
Figure 5  Cont.
Figure 6

(a) 25 May

(b) 25 May, 13:00 UTC

(c) 26 May

(d) 26 May, 13:00 UTC

(e) 27 May

(f) 27 May, 13:00 UTC

(g) 01 Jun

(h) 01 Jun, 13:00 UTC

(i) 03 Jun

(j) 03 Jun, 13:00 UTC
Figure 7
Figure 7 Cont.
Figure 8
Figure 9
Figure 10
Figure 11
Figure 12
Figure 13
FIGURE CAPTIONS

Figure 1. The maps of simulation domains with terrain height (shaded), SAMUM--I flight track (purple solid line) on 4 June 2006, and locations of observation stations used in the analysis: (a) the large domain with the black dashed box indicating the area of the small domain; (b) small domain, lines AA’ and BB’ show locations of vertical cross sections at 7.79°W and 6.3°W, respectively. The description of the observation stations is provided in Table 1.

Figure 2. The AOD time series for the period of the SAMUM--I experiment of the observed at 0.5 μm (red) from four AERONET stations and the simulated at 0.6 μm in the large (blue) and small (green) domains at (a) Ouarzazate (OZT), (b) Ras--Al--Ain (RAS), (c) Sadaa (SAD), and (d) Tamanrasset (TAM). The geographical coordinates of the AERONET stations are provided in Table 1.

Figure 3. The time average dust emissions (shaded) in μg m⁻² h⁻¹ from 22 May to 7 June 2007. The colored circles marked as A, B, C, D, E and F show emission areas of Chotts Aljard, Grand Erg Occidental, Tiris Zemmour, haboob activity area near Tindouf, the Senegal River basin, and the upper Niger River basin, respectively. The black dashed line box shows the small domain area.

Figure 4. A comparison of the model (blue) wind speed (m s⁻¹), wind direction (degrees), air temperature (C) and specific humidity (g kg⁻¹) with the aircraft observations (red) conducted during the research flight on 4 June 2006 from 9:18 UTC to 12:35 UTC. The solid purple line shows the height of the flight trajectory (km) with respect to the vertical axis on the right.

Figure 5. The observed (red) and simulated (blue) soundings of potential temperature (K), specific humidity (g kg⁻¹), wind speed (m s⁻¹) and wind direction (degrees) at 12:00 UTC on 4 June 2006 for In--Salah (a, b, g and h, 2.50°E, 27.23°N), Bechar (c, d, i, and j, 2.25°W, 31.50°N) and Tamanrisset (e, f, k and l, 5.51°E, 22.78°N).

Figure 6. The comparison of 0.55 μm AOD for the period of SAMUM--I experiment from
Terra--MODIS observations (a, c, e, g, i) and WRF--Chem (b, d, f, h, j) simulations at 0.6 µm. The MODIS plots for each date are generated by overlaying a few images taken during a period from 10:00 to 13:00 UTC. The WRF--Chem AOD is shown at 13:00 UTC.

**Figure 7.** (a) Dust concentration (shaded, µg m⁻³), height of the nocturnal boundary layer (dashed gray line, km), vertical velocity (m s⁻¹, upward -- black solid line, and downward -- grey dashed line) at the AA` meridional cross section on 7 June 2006 at 04:00 UTC; the solid gray line shows the height of the PBL (km) at 15:00 UTC the day before on 6 June that we associate with the RL height in the morning; (b) Same as (a) but for BB` cross section and vertical velocity contours are replaced by (V,W) wind barbs (m s⁻¹); (c) same as (a) but on 7 June 2006 at 15:00 UTC and the PBL height is shown for 7 June 2006 at 15:00 UTC; (d) same as (c) but for the BB` cross section; (e) Specific humidity (g kg⁻¹) and potential temperature (K, contour lines) in the AA` cross section on 7 June 2006; (f) Same as (e) but for the BB` cross section. The red arrow on the x--axis indicates the location of the vertical soundings in Figure 8.

**Figure 8.** Vertical profiles of (a) potential temperature (K), (b) wind direction (degrees), (c) wind speed (m s⁻¹), (d) specific humidity (g kg⁻¹), (e) total aerosol concentration (µg m⁻³) and (f) aerosol fine mode concentration (µg m⁻³) on 7 June 2006 at three different times in the location of 6.3°W, 33.35°N as indicated by the red arrow in Figure 7f.

**Figure 9.** The domain average dust loadings (g m⁻²) above (solid line with filled circles) and below (thin solid line), the PBL height over land (red) and over the ocean (blue) as functions of time.

**Figure 10.** The column AOD at 0.6 µm from the WRF--Chem simulations (solid lines) and AERONET observations (filled circles) for the fine (red) and coarse (blue) modes at the location of the Ouarzazate AERONET station. The AERONET retrievals were calculated using the Spectral Deconvolution Algorithm (SDA).

**Figure 11.** The column integrated aerosol number size distributions [m⁻³] from the WRF--Chem simulations at 7:00 UTC (green--fine mode, black--coarse mode, blue--total) and the AERONET Inversion Level 2.0 (red) at Ouarzazate at 7:07:41 UTC on 4 June 2006.
Figure 12. The height (km) of the flight track as a function of time (solid magenta line) for SAMUM--I flight 060604a on 4 June 2006. The blue line shows the height of the PBL=CBL/NBL+RL and gray shading shows the land elevation at the location of the aircraft.

Figure 13. The simulated (blue) and observed (red) aerosol number--size distribution [$m^{-4}$] averaged in time over the constant height elements of the airplane track during research flight 060604a on 4 June 2006. The observed time intervals and the heights of the elements of the flight track are shown in the titles of each panel. The simulated size distributions are sampled at corresponding locations at 10:00 UTC (a--c) and at 11:00 UTC (d--f) on 4 June.