High-resolution regional modeling of summertime transport and impact of African dust over the Red Sea and Arabian Peninsula

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Key Points

Fine-scale modeling of impact of dust from East Africa to the Red Sea and Arabian Peninsula

Evaluating dust generation mechanisms

Assessing dust radiative forcing
Abstract

Severe dust outbreaks and high dust loading over Eastern Africa and the Red Sea are frequently detected in the summer season. Observations suggest that small-scale dynamic and orographic effects, from both the Arabian and African sides, strongly contribute to dust plume formation. To better understand these processes, we present here the first high resolution modeling study of a dust outbreak in June 2012 developed over East Africa, the Red Sea, and the Arabian Peninsula. Using the Weather Research and Forecasting model coupled with Chemistry component (WRF-Chem), we identified several dust generating dynamical processes that range from convective to synoptic scales, including synoptic cyclones, nocturnal low-level jets, and cold pools of mesoscale convective systems. The simulations reveal an eastward transport of African dust across the Red Sea. Over the northern part of the Red Sea, most of the dust transport occurs above 2 km height, whereas across the central and southern parts of the sea, dust is mostly transported below 2 km height. Dust is the dominant contributor (87%) to the aerosol optical depth, producing a domain average cooling effect of -12.1 W m\(^{-2}\) at the surface, a warming of 7.1 W m\(^{-2}\) in the atmosphere, and a residual cooling of -4.9 W m\(^{-2}\) at the top of the atmosphere. Both dry and wet deposition processes contribute significantly to dust removal from the atmosphere. Model results compare well with available ground-based and satellite observations, but generally underestimate the observed maximum values of aerosol optical depth. The satellite-retrieved mean optical depth at some locations are underestimated by a factor of two. A sensitive experiment suggests that these large local differences may result from poor characterization of dust emissions in some areas of the modeled domain. In this case study we successfully simulate the major fine-scale dust generating dynamical processes, explicitly resolving convection and haboob formation. The future development of this novel approach will be beneficial for dust research, assuming steady growth of available computational power.
Keywords: Red Sea, dust, generation, transport, deposition, haboob, low level jet

1 Introduction

Satellite and ground-based observations show very high dust loading over East Africa and the Red Sea region in summer [Jiang et al., 2009]. Dust is further transported from this region to the Western Arabian Peninsula. Observations suggest that small-scale local dynamics and orography, which are not well resolved by conventional models, strongly contribute to the dust plume formation [Jiang et al., 2009]. A high resolution modeling of dust phenomena is computationally demanding and requires high-resolution input fields. However it is able to better quantify dust source regions [Ginoux et al., 2012; Prospero et al., 2002], meteorological mechanisms that control dust emission fluxes [Cuesta et al., 2009; Knippertz and Todd, 2012], transport pathways [McKendry et al., 2007], dust radiative direct [Perez et al., 2006; Haywood and Boucher, 2000; Liao and Seinfeld, 1998] and indirect [Levin et al., 1996; Satheesh and Krishna Moorthy, 2005] effects, complex atmospheric chemistry [Dentener et al., 1996; Stockwell et al., 1990], and deposition processes [Duce et al., 1991; Pryor et al., 2008].

The Middle East and North Africa (MENA) region (at 0-40° N and 15° W-60° E) is the largest and most persistent source of mineral dust [Prospero et al., 2002; Washington et al., 2003]. It accounts for more than half of the globally emitted dust mass [Goudie and Middleton, 2001]. Dust generation is estimated to be in the range of 1000-6000 Tg a\(^{-1}\) [Evan et al., 2014]. This large uncertainty is due in part to the lack of detailed information on dust sources or/and not accounting for small-scale features that would potentially be responsible for a large fraction of global dust emission [Ginoux et al., 2012]. Under favorable atmospheric conditions, the dust from the MENA region can be transported over thousands of kilometers further towards the North Atlantic Ocean and the United States, the Caribbean and South America.
[McKendry et al., 2007; Prospero and Lamb, 2003], the Mediterranean Sea and Europe [Santese et al., 2007], and towards the Indian Ocean and East Asia [Alizadeh-Choobari et al., 2014; Tanaka et al., 2005], affecting almost half the globe. The MENA dust long-range transport is characterized by strong seasonal variability and is closely related to the dust source areas over the region. For example, Gläser et al. [2015] analyzed transatlantic dust transport, using data from a 5 year time slice simulation with a global chemistry climate model. They found that the transport of dust in boreal winter and spring mainly occurs towards South America, whereas in summer and autumn the preferred pathway is towards the Caribbean. They also identified that the main source regions of transatlantic dust transport are located in northwestern Africa (Algeria, Mali, and Mauritania) for all seasons but not farther east, e.g. the Bodélé Depression and the Arabian Peninsula. Additionally, Tsamalis et al. [2013] analyzed the seasonal vertical distribution of the transported African dust over the tropical Atlantic, based on 5 year Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) observations at 1° horizontal resolution. The observations indicate that the Saharan Air Layer (SAL) presents a clear seasonal cycle and appears more northern in latitude and higher in altitude during summer than during winter.

Dust is generated by surface winds in areas with loose fine-grain soils. There are several important meteorological mechanisms that control near-surface winds. Dry convection in the boundary layer is an effective mechanism for dust uplift through rotating dust devils and nonrotating dust plumes. While dust devils range in width from about 3 meters to 30 meters, the dust plumes can extend up to a few hundred meters. They are of a spatial scale that is not resolved in conventional models, however their global estimated contribution exceeds 3.4% [Jemmett-Smith et al., 2015], and their local effect might be much greater for specific conditions. Moist convection and associated cold-pool outflows (haboobs) can produce severe dust events that typically develop in the late afternoon and span hundreds of
kilometers over desert areas [Roberts and Knippertz, 2014; Solomos et al., 2012]. Another important mechanism for dust emission is the downward mixing of momentum from a low-level jet (LLJ) in the mid-morning when the convective boundary layer grows [Washington and Todd, 2005; Knippertz, 2008; Schepanski et al., 2009]. An LLJ is defined as wind speed maximum in the lower part of the troposphere [Fiedler et al., 2013]. It can form over a flat terrain after sunset, when the daytime mixing stops and near surface sub-geostrophic flow becomes frictionally decoupled from the large-scale flow in the free troposphere. Usually the core (wind maximum) of the jet coincides with the top of the nocturnal inversion [Blackadar, 1957]. Finally, large-scale circulations, such as cyclones, anticyclones, and thermal lows over continents, are sources of strong surface winds that are able to produce dust emissions [Fiedler et al., 2014; Marsham et al., 2008]. Bou Karam et al. [2008] also provide observational evidence of strong dust emissions along the leading edge of a large scale West African monsoon flow in the intertropical discontinuity (ITD) region that is defined as the near-surface convergence zone between the dry harmattan and the moist monsoon flow. A comprehensive discussion of various meteorological processes that lead to dust generation is discussed in Knippertz and Stuut [2014].

The dust radiative effect (DRE) is a measure of the dust impact on the Earth’s energy budget. The DRE is sensitive to aerosol particle size distribution, shape, mixing state, mineralogical composition, the altitude of the dust layer, and underlying surface properties [Otto et al., 2007]. The choice of mixing rule affects both aerosol absorption and scattering [Bond et al., 2006]. Light-absorbing aerosols could absorb more light when they are mixed with non-absorbing materials. Aerosol composition and size distribution are both strongly modified by water uptake [Ghan and Zaveri, 2007]. Balkanski et al. [2007] showed that dust mineralogy could change the magnitude and even the sign of the DRE at the top of atmosphere (TOA). In their study, the radiative effect of dust is linked to the refractive indices of major minerals
[Sokolik and Toon, 1999] whose relative abundance in dust varies with the geographical region. Dust optical properties also evolve during atmospheric transport due to a variety of heterogeneous chemical reactions affecting dust composition [Wu et al., 2011; Zhang et al., 2012] and microphysical processes affecting dust size distribution [Schulz et al., 2012].

Therefore, to realistically simulate dust phenomena, we need to employ models that resolve a large range of scales, from several meters to thousands of kilometers, and include a radiatively interactive air-quality module that is fully coupled with meteorological processes. Currently, global models with their coarse resolution are not able to resolve small-scale features that are responsible for a large part of dust emissions [Roberts and Knippertz, 2014], and they are mainly used to simulate time-average dust features [Shao et al., 2011]. Misrepresentation of moist convective processes in global weather and climate models can lead to large errors in dust emissions [Marsham et al., 2011; Garcia-Carreras et al., 2013]. Furthermore, the convective organization is in fact absent from global models [Moncrieff, 2010]. The parameterized convection up- and down-drafts are assumed to occur within a single model grid-cell, however, the mesoscale convective systems (MCSs) and their associated cold pools, can frequently cover many grid-cells [Pantillon et al., 2015]. Chemical transport models, which calculate air quality offline, are also deficient. Grell [2004] found that, on cloud-resolving scales, wind velocity power spectrum and chemical profiles in offline simulations are susceptible to large errors in the vertical distribution of aerosols and chemical species.
In this study, we use the high-resolution Weather Research Forecast (WRF) model [Skamarock and Klemp, 2008], interactively coupled with an air-quality module (WRF-Chem [Grell et al., 2005]), to investigate intensive dust generation in subtropical East Africa that contributes significantly to dust optical depth over the Red Sea region, Eastern Africa, and the Western Arabian Peninsula. To identify different dust generation mechanisms and to quantify dust emissions, transport, deposition, and radiative impact, we consider the period of June 2012 when a well-developed dust plume was observed from satellites and by the ground-based Aerosol Robotic Network (AERONET) station at King Abdullah University of Science and Technology (KAUST) campus at the eastern coast of the Red Sea. During that event, dust was transported from East Africa mostly through the Tokar Mountain Gap [Davis et al., 2015; Jiang et al., 2009] across the Red Sea towards Western Arabia. Most previous studies of dust phenomena over North Africa on cloud-resolving scales have been focused primarily on western and central regions [Roberts and Knippertz, 2014; Solomos et al., 2012], thus the present study seeks to quantify various effects of dust from East Africa, using a fully coupled aerosol-chemistry-meteorology model explicitly resolving deep convection.

The model results were compared with available ground-based and satellite observations.

The remainder of this paper is organized as follows. In section 2, we present the model description and describe the experimental domain. Section 3 describes the ground- and satellite-based observations. Section 4 quantifies dust emissions, transport, deposition, and radiative effects over the region of East Africa, the Red Sea, and the Arabian Peninsula. Finally, we summarize our work and draw conclusions in Section 5.
2 Model

2.1 Model description

The Weather Research and Forecasting (WRF) model is a numerical weather prediction (NWP) system that is designed to simulate atmospheric processes over a wide range of spatial and temporal scales for both operational forecasting and atmospheric research [Skamarock and Klemp, 2008]. WRF is based on fully compressible nonhydrostatic Navier-Stokes equations that describe atmospheric flow. The model uses a generalized vertical, terrain-following coordinate system and takes into account a variety of unresolved subgrid-scale physical processes, such as boundary layer turbulence, deep and shallow convection, cloud microphysics, and land surface processes. WRF-Chem is an extended version of the WRF model, which includes a chemistry component [Grell et al., 2005] and has been widely used to study dust impacts on local to regional scales [Fast et al., 2006; Kalenderski et al., 2013; Kumar et al., 2014; Prakash et al., 2015; Zhao et al., 2010; Zhao et al., 2011; Zhao et al., 2013]. The chemistry component is fully coupled with the meteorological model by using the same domain grid, advection scheme, physics parameterizations, and time steps. Additionally, the chemistry component takes into account a variety of coupled physical and chemical processes such as deposition, chemical transformations, aerosol interactions, photolysis, and emissions.

2.2 Model configuration and experimental domain

In this study we used the WRF model version 3.5.1, configured with the Mellor-Yamada-Janjic (MYJ) [Janjic, 2002] planetary boundary layer scheme to parameterize the boundary layer processes. We selected the Rapid Radiative Transfer Model (RRTMG) [Mlawer et al., 1997] to represent longwave and shortwave radiation transfer within the atmosphere and the surface. The Lin microphysics scheme [Lin et al., 1983] accounted for non-convective precipitation processes, and the Grell convective scheme [Grell and Dévényi, 2002]
accounted for cumulous cloud parameterizations. The NOAH Land Surface model [Chen and Dudhia, 2001] and the Janjic Eta surface layer scheme [Janjic, 1996] provided surface process calculations.

The air quality component of WRF-Chem was configured with the Regional Acid Deposition Model 2 (RADM2) photochemical mechanism [Stockwell et al., 1990], Fast-J photolysis scheme, and the Modal Aerosol Dynamics Model for Europe/Secondary Organic Aerosol Model (MADE/SORGAM) with included aqueous phase chemistry [Ackermann et al., 1998; Schell et al., 2001].

The MADE/SORGAM aerosol model was coupled with the Goddard Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) dust emission scheme [Ginoux et al., 2001] to account for dust emission processes. The GOCART scheme calculates the dust emission flux \( F_p \left( \mu g \cdot m^{-1} \cdot s^{-1} \right) \) for particle size class \( p \) by the expression:

\[
F_p = \begin{cases} 
C \times S \times s_p \times u_{10m}^2 \times (u_{10m} - u_t) & \text{if } u_{10m} > u_t \\
0 & \text{otherwise}
\end{cases}
\]

(1)

where \( C \) is an empirical proportionality constant \( \left( 1 \times 10^{-9} \text{ kg s}^{-2} \text{ m}^{-3} \right) \); \( S \) is the source function with a value between 0 and 1; \( u_{10m} \) is the horizontal wind speed at the altitude of 10 m; \( u_t \) is the threshold velocity calculated as a function of air and particle density, particle size, and surface moisture [Bagnold, 1941]; \( s_p \) is the fraction of size class \( p \) within the soil. However, the dust emission scheme was used only to calculate the total mass dust emission flux which was then distributed into the two lognormal modes, accumulation and coarse, of the aerosol model with mass fraction coefficients equal to 7% and 93%, volume median diameters of 0.6 \( \mu m \) and 6 \( \mu m \), and geometric standard deviations of 2 and 2.2, respectively [Kalenderski et al., 2013]. For this simulation the value of the proportionality constant \( C \) was tuned to
$0.4 \times 10^{-9}$ kg s$^2$ m$^{-5}$, minimizing the difference between measured and simulated mean optical depth, and to be consistent with the optical depth measurements at the KAUST Campus AERONET station located on the west coast of Saudi Arabia. In addition, a new dust source function, $S$, at higher resolution $0.1^\circ \times 0.1^\circ$ derived from MODIS Deep Blue estimates of dust optical depth in conjunction with other datasets [Ginoux et al., 2012] was used. This new map provides a more detailed representation of the small scale dust sources.

The parent (lat/lon) and nested (lat/lon) domains are shown in Figure 1. The parent domain has a grid spacing of 10 km and uses 440x297 grid points to cover East Africa, the Arabian Peninsula, and the Red Sea, while the nested domain uses 756x656 grid points and 2 km grid spacing. The nested domain is centered over the dust source regions of North Sudan. It is assumed that, in the high-resolution child domain, convective processes are explicitly resolved and convective parameterization is turned off [Costantino and Heinrich, 2014]. The model has 35 vertical layers designed to provide increased resolution in the boundary layer, with the second model level placed at ~20 m above the ground and the top of the model atmosphere located in the stratosphere at 50 hPa.

The European Centre for Medium-Range Weather Forecasts (ECMWF) operational analysis with a horizontal resolution of about 16 km was used to prepare the lateral boundary and initial conditions for the meteorological fields for the parent domain. The child domain was driven by two-way boundary conditions imposed using fields from the parent domain. Initial and boundary conditions for aerosol and gas-phase species were assembled using the default WRF-Chem profiles. These profiles were obtained from various field studies to represent clean atmosphere, maritime, mid-latitude conditions [McKeen et al., 2002]. We used anthropogenic emissions from the REanalyzer of the TROpospheric (RETRO) chemical composition inventories (http://retro.ens.es/index.shtml) and the Emission Database for
Global Atmospheric Research (EDGAR) datasets (http://edgar.jrc.ec.europa.eu). The biomass burning emissions were provided from the Global Fire Emissions Database, Version 2 (GFEDv2.1) [Randerson et al., 2005]. All emission inventories were preprocessed using the PREP-CHEM-SRC v1 emissions preprocessor [Freitas et al., 2011].

The simulations were performed for a period extending from 15 to 26 June 2012, which includes a major dust outbreak from East Africa across the Red Sea and the Arabian Peninsula. We applied a five-day spin-up period and analyzed results only from 21 to 26 June.

3 Observations

Satellite retrievals and ground-based observations were used to test and tune the model.

3.1 AERONET data

AERONET is a ground-based remote sensing aerosol network established by NASA to measure the optical properties of aerosols and validate satellite retrievals [Holben et al., 1998]. It uses CIMEL sun and sky radiometers, which are automatic robotically-operated instruments located across the world. Aerosol products are routinely retrieved from AERONET raw data following the approach described by Dubovik and King [2000] and Dubovik et al. [2000]. The uncertainty of AOD at 600 nm under cloud-free conditions is less than ±0.01 [Holben et al., 1998]. In this study, AERONET version 2, level 1.5 cloud-screened data were employed throughout the analysis of the optical properties of aerosols. For comparison of the optical properties from the model output at 600 nm wavelength and the AERONET measurements, the Angstrom power law was used:

\[
X(600\text{nm}) = X(675\text{nm}) \times \left(\frac{600\text{nm}}{675\text{nm}}\right)^{-a}
\]
where $\alpha$ is the Angstrom exponent provided by AERONET measurements and given by

$$\alpha = \ln \left( \frac{X(440\text{nm})}{X(675\text{nm})} \right) / \ln \left( \frac{675\text{nm}}{440\text{nm}} \right),$$

(3)

$X$ is aerosol optical depth. For comparisons, the simulated data were sampled at the same times and locations as were measurements.

### 3.2 Satellite data

In this study we used aerosol optical depth (AOD) fields from Moderate Resolution Imaging Spectroradiometer (MODIS) and Spinning Enhanced Visual and Infrared Imager (SEVIRI), Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) backscatter profiles and rainfall retrievals from the Tropical Rainfall Measuring Mission (TRMM) instrument.

MODIS Collection 06 “Deep Blue” 550 nm AOD [Hsu et al., 2004] Level 2 data were used in combination with the standard ocean algorithm [Tanré et al., 1997] for comparison with simulated aerosol optical properties. Additionally, true color imagery Level 1B data were used to identify cloud and dust transport across the Red Sea. MODIS Level 2 data were provided at a spatial resolution of 10 km x 10 km, and Level 1B data were provided at a spatial resolution of 1 km x 1 km. Both datasets have poor temporal resolution of two to four times per day for a given location.

SEVIRI false color dust imagery [Brindley and Ignatov, 2006; Lensky and Rosenfeld, 2008] was used for tracking the development of dust storm systems and the SEVIRI 600 nm AOD product [Brindley et al., 2015] was used to evaluate simulated time averaged AOD spatial distribution. SEVIRI products cover the entire experimental domain and were provided at a spatial resolution of 5 km x 5 km and at a temporal resolution of 15 minutes.
CALIOP Version 3.02 Level 1 attenuated backscatter profiles at 532 nm with 30 m vertical resolution and a profile spacing of 333 m [Winker et al., 2007] were used to evaluate dust vertical structure.

TRMM 3B42 V7 [Huffman et al., 2007] rainfall retrievals were used to analyze synoptic condition and to verify the simulated precipitation. These gridded estimates were provided at a spatial resolution of 0.25×0.25 and at 3-hour temporal resolution in a global belt extending from 50°S to 50°N latitude.

3.3 ECMWF ERA-Interim

The European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim reanalysis [Dee et al., 2011] provides meteorological fields. In this study we used dew point surface temperature, geopotential-height, and wind fields to analyze synoptic-scale features. ERA Interim output is available at a 6-hour temporal resolution on a reduced Gaussian grid with a horizontal resolution of 0.7° (approximately 78-km grid spacing), and 60 terrain following vertical levels. ERA-Interim reanalysis assimilates a large number of ground-based and satellite observations using a 4D-Var technique and is often viewed as best estimate of the past state of the atmosphere.

4 Results and Discussion

In this section we validate the model meteorology and AOD output using reanalysis and observation data. The considered case study allows us to identify several important mechanisms of dust generation over the model domain and to estimate the total dust emissions from these mechanisms. In our simulations, we explicitly resolve deep convection and therefore are able to describe haboob genesis in more detail than in previous studies. Further, in this section, dust transport, deposition, and radiative impact are discussed.
4.1 Weather synoptic conditions

We used ERA-Interim data to analyze the evolution of large-scale weather conditions during the simulation period. Figure 2 shows the geopotential height, at the 925 and 500 hPa isobaric surfaces, the estimated 24 hour accumulated rainfall from TRMM, and 14°C 2 m dew point temperature locations (for estimating the ITD position) at 00 UTC for 20-25 June. On 20 June (Figure 2a), about three days before the beginning of the dust outbreak, the circulation at 500 mb was characterized by a marked upper level trough centered over the eastern Mediterranean Sea and a large-scale ridge covering both North Africa and the Arabian Peninsula. The main features of the low level (925 mb) circulations were the Sahara heat low over Algeria, northerly Harmattan flow over East Africa, and south-west monsoon current and associated rainfall over the southern part of the domain (see Figure S1 and S2 in the supporting information). The ITD was mostly zonal around 15°N. The high pressure system over the Ethiopian highlands was present during the entire period of interest.

On the next day (Figure 2b), the upper-level trough amplified and moved further southeastward into East Africa and the northern Arabian Peninsula, splitting the large-scale ridge into two distinct regions with one ridge over north Algeria and the other over the central part of the Arabian Peninsula. In the lower levels, the Sahara and Arabian heat lows intensified and their locations remained approximately unchanged as did the position of the ITD.

Figures 2c and 2d characterize the synoptic situation during the most intensive days of dust generation (Figure 3c). At the time the upper-level trough moved east-northward, the ridge over the Arabian Peninsula moved slightly west and the ridge over north Algeria was nearly stationary. The preexisting surface large horizontal temperature gradient between the Nubia desert and the Ethiopian highlands, combined with the north-west located upper level trough,
caused a significant pressure decrease in the lower troposphere and formation of a low-pressure system in north Sudan (Sudan heat low), seen as a 7032 m$^2$ s$^{-2}$ minimum on 925 mb surface. The low-pressure system enhanced the south-west monsoon flow, thus favoring a northward advance of the ITD in north-east Sudan up to ~20°N. Additionally, a strong surface wind of 15 m s$^{-1}$ developed to the east of the low, which activated several dust source regions bounded between ~15° - 21°N, 28° - 34°E [Prospero et al., 2002].

Favorable synoptic conditions for dust transport over the Red Sea on 24 and 25 June are shown in Figures 2e and 2f. A high-low dipole pattern was formed by the upper-level high over the Arabian Peninsula and the upper-level low over Syria, which favors eastward transport of dust in the mid troposphere over the northern Red Sea. Hypothetically, the reverse dipole (i.e. high in the north and low in the south) would favor westward transport of dust over the Red Sea. Additionally, while the ridge intensified and continued to move westward crossing the Red Sea on 24 June, the trough was nearly stationary. At that time, the surface low pressure system was no longer present over north Sudan and the ITD retreated southward becoming mostly zonal again, ~15°N on 25 June. The precipitation bands propagating westward in the 5 - 15°N latitude zone could indicate the presence of African Easterly Waves [Burpee, 1974] with wavelength of ~2000 km, corresponding to a period of about 3 days.

The model-simulated large-scale circulation (not shown) was consistent with the ERA-Interim reanalysis. The spatial correlation coefficient between simulated and ERA-Interim time averaged surface winds was greater than 0.81 for the parent domain and was 0.36 for the nested domain. The latter value is low, as reanalysis fields retain processes at 75 km scale and lose their skill at 2 km scales in the nested model domain. The model simulated higher wind speeds in the vicinity of elevated terrain and much stronger gap winds along the coast of Red
Sea, up to 6 m s\(^{-1}\) for the nested domain (see Figure S3 in the supporting information). The mean difference of simulated and observed surface winds was small over flat areas. As expected, the agreement for the surface temperature was better than that for the surface winds (see Figure S4 in the supporting information), with correlation coefficients of 0.94 for the parent domain and 0.85 for the nested domain. Similar results were obtained for other meteorological parameters.

Dust emissions

The emission fluxes of mineral dust were calculated interactively by Eq. (1) within the model. The spatial distribution of time averaged dust emissions (Figure 3a) was largely controlled by southwest and northeast winds and quite different from the source function (Figure 3b) indicating that not all source regions were active throughout the study period. The most intense dust emissions were located in the Tokar delta region, where a large alluvial deposit was produced near the Baraka River at the Red Sea coast. A high level of dust activity was also observed in the large area bounded on the east by the western flanks of the Ethiopian highlands, and the Ennedi Plateau on the west, which includes several dust sources like ephemeral channels that extend from the highlands towards the Nile River [Prospero et al., 2002].

The maximum value of the dust emission flux over the nested domain during the post-spin-up period of the simulation reached 391 µg m\(^{-2}\) s\(^{-1}\). For comparison, the dust emissions simulated here were well within the range reported by Kang et al. [2011]. The time series of daily total emissions in the nested domain (Figure 3c) shows that dust generation was most intense on 22 and 23 June with the peak value of ~2.8 Tg day\(^{-1}\). The emitted total dust amount was 54.5 Tg for the entire domain and simulation period.
Several dynamical processes spanning a large range of scales were identified by the model as potential sources of dust emissions, including synoptic cyclones, nocturnal low-level jets, and cold pools of mesoscale convective systems. In the following subsections, examples to illustrate the occurrences of the highlighted processes in different dynamical regions are provided. The analyses were conducted over the high-resolution nested domain.

4.1.1 Sudan heat low

The Sudan heat low was a key factor of atmospheric circulation over the region and a major mechanism of strong surface wind generation during the simulation period, which at that time was centered near to 19°N, 30°E. The simulated heat low over north Sudan was strongest during the daytime and weakened or even disappeared during the night. Figure 4 shows the spatial distribution of near-surface wind field and dust emissions at 08:00 UTC on 22 June 2012. It can be seen that the heat-low vortex enhanced the south-west monsoon flow and, as a consequence, favored northward advance of the ITD (Figure 2c). The strong winds related to the heat-low vortex over an area comprised between 15–22°N and 28–34°E reached up to 20 m s⁻¹ and generated 0.19 Tg of dust between 07:00-08:00 UTC.

4.1.2 Nocturnal low level jet

The simulated intense dust emissions over the Tokar delta region were associated with the breakdown of a strong and persistent nocturnal LLJ. The nocturnal LLJ embedded in the monsoon flow was present on 5 of 6 days of the study period and strong enough to induce dust emissions. Figure 5a shows the simulated dust emission fluxes at the time (08:00 UTC) of the strongest surface winds on 23 June 2012 and vertical cross-section (Figure 5b) of LLJ at the time (03:00 UTC) of the maximum wind speed found in the core on the same day. The vertical cross-section, denoted with a straight white line in Figure 5a, is located at the exit region of the Tokar Gap, which is approximately 50 km inland from the Tokar delta on the
Red Sea coast. The Red Sea Hills not only channeled the southwestern winds through the
Tokar Gap but also increased their speed due to the orographic effect. The core of the jet in
this region was located near 300 m above ground level (agl), where the wind speed reached
its highest value. Figure 5c shows the average wind speed profiles over the study period at
18.61°N, 37.82°E (see Figure 5a, the white dot). The LLJ wind ranged from 22 m s\(^{-1}\) to 28 m
s\(^{-1}\) during the simulation period, with the mean value of 23 m s\(^{-1}\) and mean height of 350 m
agl at 02:00 UTC. The maximum near-surface winds ranged from 13 m s\(^{-1}\) to 19 m s\(^{-1}\), with
the mean value of 16 m s\(^{-1}\) at 06:00 UTC, which is about 3 hours after sunrise (05:53 LT),
and thus lagged the mean maximum wind speed in the core by 4 hours. The estimated mean
vertical wind difference between the LLJ core and the level 500 m above was approximately
-5 m s\(^{-1}\) and between the LLJ core and the level 100 m below was 1.1 m s\(^{-1}\). The potential
temperature vertical gradient exceeded 0.009 K m\(^{-1}\) in the layer below the LLJ core,
indicating a stably stratified surface layer. This is consistent with the Allen and Washington
[2014] LLJ detection criteria developed, based on the high-quality Fennec observational
dataset for Bordj-Badji Mokhtar, Algeria, during June 2011.

It should also be noted that the area of intensive dust uplifting (up to 100 µg m\(^{-2}\) s\(^{-1}\)) was
found north of the Tokar delta along the coast of the Red Sea. It was supposedly activated by
strong land/sea breezes in the morning hours.

**4.1.3 Haboob**

During the study period several haboobs were simulated over the domain of interest,
however, the presence of deep convective clouds prevented many of these events from being
verified by satellite observations. Figure 6 presents a haboob event that occurred on 21 June
2012. The development of the dust storm started with several small convective events around
14°N, 30.5°E at about 12:00 UTC. By 14:00 UTC the cold pool can be clearly identified in
the vertical cross sections (Figure 6c) of virtual potential temperature, turbulent kinetic energy (TKE), and vertical wind speed along the direction of travel of the dust storm (Figure 6b). The contours of the TKE indicate the change of the boundary layer structure, caused by advancing density currents during the daytime. The cold pool cooled and stabilized the lower layer near the ground behind the leading edge and created strong vertical mixing up to 4.2 km height along the interface between the gust front and the undisturbed well-mixed layer in front of the traveling system. At 16:00 UTC the haboob system was fully developed (Figure 6, middle row). The simulated density currents (Figure 6f) propagated eastward and westward from the buoyancy source, forming two distinct lobes at (15°N, 30°E) in Figure 6e, which is in good agreement with the SEVIRI imagery at that time. Strong surface winds of 25 m s\(^{-1}\) caused an extreme dust emission flux of 1670 \(\mu g m^{-2} s^{-1}\) (Figure 6c), and the updrafts at the leading edge of the cold pool reached a maximum of 5.1 m s\(^{-1}\) (Figure 6f). The haboob propagated with an average speed of 14 m s\(^{-1}\). Figure 6 (bottom row) displays the dissipation stage of the dust storm system at 18:00 UTC. The emission rates significantly decreased to less than 400 \(\mu g m^{-2} s^{-1}\) and no precipitation was simulated (Figure 6h) at that time. SEVIRI RGB dust imagery (Figure 6g) captures the dust plume elevated at 2-3 km that developed from the aged cold pool as well as the formation of a new low-level (0-1 km) dust plume near 18°N.

The considered dust storm case lasted for more than 7 hours, spreading over a distance of 500 km in the northeast direction and developing the 400 km front line. The estimated emitted dust amount was 0.31 Tg over the area covered by the dust storm. A similar haboob case was recently simulated by Heinold et al. [2013] over Mauritania, West Africa. They calculated offline dust emission fluxes using convection-permitting simulations provided by the UK Met Office Unified Model to estimate the contribution of different dust generation processes. In their study the simulated dust storm lasted over 6 hours spreading at a speed of about 10 m s\(^{-1}\).
over a distance of 600 km and forming a dust plume of over 400 km in length. However, our haboob case is quite different from that reported in Roberts and Knippertz [2014]. They simulated an unusually large haboob that was supported by the African easterly jet and lasted over two days, producing a dust plume over 1000 km long over the central Sahara.

4.2 Dust transport

A clear Aqua MODIS true color image (Figure 7a) over the Red Sea on 24 June captured a massive transport of mineral dust from the African continent to the east. The large scale circulation simulated by WRF-Chem (Figure 7c) on 24 June shows positive sea level pressure anomalies (SLPAs, color filled-contours) over north-east Africa, the north-central part of the Arabian Peninsula, and the western flanks of the Ethiopian highlands. At the same time a weak isolated low-pressure system can be seen over the east coast of the Red Sea between 17°N and 23°N. Under such conditions, the model simulation suggested a synoptic flow that produced large-scale eastward advection of the dust plume at low levels in the central and southern parts of the Red Sea. Moreover, our analysis of the synoptic conditions in Sec. 4.1 shows that an upper-level geopotential height dipole favored eastward transport in the mid-troposphere over the northern part of the sea at that time. To further evaluate dust transport routes over the Red Sea, we used forward trajectories [Lee and Chen, 2014] initiated at different locations, altitude, and time. The 72h trajectories (Figure 7c) were released on 23 June 00:00 UTC at a starting altitude of 4.5 km (blue line) and 40 m (green line) based on the observed vertical profiles of dust layers. The forward-trajectory and dust load contour lines (Figure 7c) show that eastward transport of dust at low and high altitudes affects the western and north-central parts of the Arabian Peninsula. The low level eastward transport due to mostly dust sources close to the Red Sea coast line is impeded by surface winds that have a very strong tendency to blow parallel to the axis of the sea and eventually get trapped into the Red Sea convergence zone [Pedgley, 1966]. These processes are additionally enhanced by the
steep orography bounding the southern part of the Red Sea, with elevations between 2 and 3 km above sea level, continuously for about 1000 km.

CALIOP’s 532 nm total attenuated backscatter profiles (Figure 7b) on 24 June 10:34-48 UTC is qualitatively compared with a model-predicted vertical cross-section of dust concentration (Figure 7d) on 24 June 11:00 UTC to evaluate model performance and vertical structure of dust transport along the CALIOP’s orbit track (Figure 7a, blue line). CALIOP’s vertical feature mask data (not shown) identify the presence of dust aerosols over the entire domain at that time. In general, the model-simulated vertical structure of the dust plume compares relatively well with CALIOP’s data. In both profiles the dust plume extends from the surface upwards up to height of about 6 km and along the entire CALIOP track. Additionally, low level and high level transport of dust over the southern part of the Red Sea is presented in both profiles (see Figure 7b and 7d). The simulated vertical distribution of dust in the cross-section (parent domain) shows that the model captured dust vertical mixing up to a 6 km height caused by the steep orography along the south-east coast of the Red Sea. The vertical dust mixing in Figure 7c is not well seen in the lower part of the boundary layer, as CALIOP observations near the surface are a subject of large uncertainty [Misra et al., 2012].

4.3 Dust radiative impact

We obtain the simulated dust impact on aerosol optical depth, $AOD_{(dust)}$, and the instantaneous DRE as a difference between the total aerosol optical depth, $AOD$, and radiative fluxes, $F_{(all)}$, calculated for all aerosol species, and a second time calculated within the same model run on the same meteorological fields, but excluding the effect of mineral dust ($AOD_{(all except dust)}$, $F_{(all except dust)}$) [Zhao et al., 2013]:

$$AOD_{(dust)} = AOD_{(all)} - AOD_{(all except dust)}$$

$$DRE = F_{(all)} - F_{(all except dust)}$$

(4)
Here $F$ is the net radiative flux (SW, LW, or total) calculated at the surface (SFC) or at the top of the atmosphere (TOA) and defined as a difference between the downward $F_\downarrow$ and upward $F_\uparrow$ fluxes, i.e. $F = F_\downarrow - F_\uparrow$. A negative sign of $F$ and DRE corresponds to a cooling effect, and a positive sign implies a heating effect.

### 4.3.1 AOD

First, we use SEVIRI and AERONET observations to evaluate the spatial and temporal variability of the AOD at 600 nm, simulated by WRF-Chem during the study period. Figure 8a-b compares the daytime (from 06:00 to 16:00 UTC) average spatial distribution of the AOD from SEVIRI retrievals and WRF-Chem model results. The comparison shows that the model results are generally consistent with the observations with a spatial correlation coefficient of 0.58. The high optical depth values in northeastern Sudan and the southern part of the Red Sea are consistent with dust transport pathways (Sec. 4.3) and are well represented in both spatial patterns. However, the model significantly underestimated the AOD over northeastern Sudan. Several additional simulations were conducted to study the sensitivity of the averaged AOD field to the model configuration. Using different aerosol models coupled with different dust emissions schemes as well as different erodibility and soil datasets revealed that the dust load can vary over some areas of the domain up to 120%. For example, the MOSAIC aerosol model coupled with the DUSTRAN emissions scheme [Shaw et al. 2008] produced a dust horizontal distribution more consistent with the satellite observations over northeastern Sudan, but overestimated the average AOD, up to 90%, in the central part of the Arabian Peninsula (see Figure S5 in the supporting information). A sensitivity test with the MOSAIC aerosol model and the GOCART emission scheme showed bad consistency and overestimation of the hourly average AOD, up to 120%, at most AERONET locations. Similar results were obtained by using the MADE/SORGAM aerosol model.
coupled with the GOCART emission scheme, using a lower resolution erodibility dataset [Ginoux et al. 2001]. These discrepancies are more likely due, in part, to poor characterization of dust emissions in some areas of the modeled domain.

Figure 9 shows time series of hourly average AOD, derived from WRF-Chem simulation at six sites in comparison with the co-located AERONET and SEVIRI observations during the simulation period. The model and AERONET time series correlate well, with the correlation coefficient ranging from 0.33 to 0.87 at all the sites, although the simulated AOD does not reach the magnitude of the observed peaks most of the time. Both observational and model data at the KAUST AERONET site, which is affected the most by the dust storm, show a significant increase of the AOD and a decrease of the Angstrom exponent, \( \alpha \) (not shown), during 24–25 June. The observed AERONET AOD reached as high as 2.4, when the Angstrom coefficient \( \alpha \) is equal to 0.05, indicating the presence of large dust particles [Toledano et al., 2007]. The observed single scattering albedo, SSA, of 0.95 corresponds to typical dust absorption.

In contrast to the KAUST site, the model and observed AOD at Sede Boker station (Israel) show much lower AOD, indicating that the location was not affected significantly by the dust storm. The observed mean \( \alpha \) of 0.43 and 1.29 respectively for the two stations also indicates the presence of different types of aerosols. The observed average values for AOD and \( \alpha \) of 0.14 and 1.29 respectively, are well consistent with the seasonal mean values of 0.1 and 1.1 reported by Cuevas et al. [2015]. The large value of \( \alpha \) for this location could be explained by the coexistence of mineral dust and fine pollution aerosols in summer, when wet removal is practically absent and the accumulation of pollution is favored [Basart et al., 2009]. It should also be noted that a significant increase of observed and simulated AOD is present at the
Mezaira and Masdar Institute sites, and the model considerably underestimated observed AOD peaks at the Solar Village and Kuwait University sites during the dust storm. The observed mean $\alpha$ at all of these stations ranged from 0.21 to 0.34, indicating that the dust was dominant in aerosols during the simulation period.

Figure 10 shows the combined instantaneous spatial distributions of AOD at 550 nm from Level 2 MODIS satellite retrievals at 10:40 and 10:45 UTC on 25 June, as well as the total AOD and AOD[all except dust] from the model simulation at 11:00 UTC. The spatial patterns of the AOD from satellite retrievals (Figure 10a) and model simulation with dust (Figure 10b) are similar, both showing intense dust transport over East Africa. However, the simulated AOD[all except dust] exceeds 0.3 only over the southern Red Sea and the Arabian Sea (Figure 10c), indicating the dominant role of dust in the entire region. The simulated domain average contribution of mineral dust in the total AOD was 87% at that time. The model captures well the aerosol SW absorption. The domain average SSA was 0.975 and 0.970 from MODIS (550 nm) and the model (600 nm), respectively.

4.3.2 DRE

The DRE calculated by Eq. 4 characterizes the instantaneous dust impact on the radiation energy budget. However, the magnitude and even the sign of the model-simulated effect at the TOA are sensitive to the solar absorptive properties of the dust particles and albedo of the bright arid and semiarid areas [Balkanski et al., 2007; Kalenderski et al., 2013]. In our simulations, the imaginary part of the shortwave refractive index of dust particles was set to 0.003, corresponding to the moderate absorbing dust of Balkanski et al. [2007], and was not wavelength dependent. The spatial distribution of daily averaged clear-sky direct DRE for SW, LW, and Net (SW+LW) radiation fluxes at the TOA, in the atmospheric column (ATM), and at the SFC are shown in Figure 11. At the surface (bottom row), the SW DRE is negative.
everywhere ranging from \(-81.7\) to \(-0.01\) W m\(^{-2}\), with the domain average value of \(-21.1\) W m\(^{-2}\). However, the instantaneous SW DRE at the surface is considerably larger, reaching up to \(-301.4\) W m\(^{-2}\). These results are comparable with results obtained in previous studies: \textit{Kumar et al.} [2014] simulated a domain average SW radiative perturbation at the surface of \(-10.1\) W m\(^{-2}\) and instantaneous cooling of up to \(-227\) W m\(^{-2}\) over northern India during a typical pre-monsoon dust storm; \textit{Slingo et al.} [2006] reported that the incoming solar fluxes at the surface dropped by 250 W m\(^{-2}\) during a major Saharan dust storm. The LW DRE warmed the surface, with a domain average value of \(9.1\) W m\(^{-2}\) and maximum instantaneous warming of up to \(28.5\) W m\(^{-2}\) over the central plateau of the Arabian Peninsula. The Net DRE ranged from \(-58.0\) to \(2.1\) W m\(^{-2}\), with a domain average value of \(-12.1\) W m\(^{-2}\), indicating that the SW flux reduction by dust was dominant in the net flux changes at the surface.

Figure 1 (upper row) displays the spatial distribution of daily averaged DRE at the TOA. The dust SW radiative effect (Figure 11a) is generally negative with values comprising between \(-38.2\) and \(4.4\) W m\(^{-2}\), with a domain average value of \(-7.3\) W m\(^{-2}\). The effect is negative over dark surfaces, like oceans and forests, with low surface albedo less than 1.5, and changes to positive over bright surfaces, like deserts and snow covered areas with high surface albedo larger than 0.3. The dust LW radiative effect (Figure 11b) is small and positive almost everywhere, with values comprising between \(-0.02\) and \(8.9\) W m\(^{-2}\) and a domain averaged value of \(2.4\) W m\(^{-2}\). However, the overall Net DRE (Figure 11c) is negative with a domain average value of \(-4.9\) W m\(^{-2}\). For comparison, \textit{Satheesh et al.} [2006] calculated daily averaged SW radiative effect ranging from \(-16.5\) to \(-6.0\) W m\(^{-2}\), LW effect ranging from \(0.8\) to \(2.2\) W m\(^{-2}\), and Net effect ranging from \(-14.3\) to \(-5.2\) W m\(^{-2}\) at the TOA, using multiyear island based and ship-borne measurement over the Arabian Sea. \textit{Zhang and Christopher} [2003] reported a dust LW warming effect of \(7\) W m\(^{-2}\) over the cloud free Sahara desert regions. Finally, \textit{Osborne et al.} [2011] modeled a daily averaged SW DRE at TOA
between 0 and -23 W m$^{-2}$, depending on particle shape, and a LW effect of 14 W m$^{-2}$ over the West Sahara region during Geostationary Earth Radiation Budget Intercomparison of Long-wave and Short-wave (GERBILS).

In the atmosphere (Figure 11, middle row), DRE is calculated as the difference between the radiative effect at the TOA and at the SFC. The SW radiative effect (Figure 11d) warms the atmosphere column by absorbing solar radiation with a domain average value of 13.8 W m$^{-2}$; the LW radiative effect (Figure 11e) cools the atmosphere column by increasing the emitted terrestrial radiation with domain average value of -6.7 W m$^{-2}$. The overall effect (SW+LW) in the atmosphere is positive with a domain average of 7.1 W m$^{-2}$, which compares well with the 6.94 W m$^{-2}$ simulated by Zhao et al. [2011] over West Africa.

### 4.4 Dust deposition

Both dry and wet deposition processes contributed significantly to dust removal from the atmosphere during the simulation period. However, their contributions show significant spatial variation. Figure 12 shows mean dry (Figure 12a) and wet (Figure 12b) deposition fluxes over the model domain and the time series of daily deposition rates integrated over the entire domain (Figure 12c). Dry deposition (Figure 12a) is the dominant removal mechanism of dust particles in the vicinity of source regions and remote desert areas, since the lack of precipitation there and the tendency of large particles, present near the source regions, to settle down quickly after they are emitted. The highest dry deposition fluxes are found in north Sudan and the east coast of the Arabian Peninsula, with the maximum greater than 2500 g m$^{-2}$ a$^{-1}$ at a few isolated locations with a domain averaged value of 139.9 g m$^{-2}$ a$^{-1}$. Dry deposition contributes nearly homogeneously (about 20 g m$^{-2}$ a$^{-1}$) over the entire Red Sea basin. The simulated wet deposition (Figure 12b) is the dominant removal mechanism for dust particles relatively far from the source regions, where precipitation is present, such as the
Sahel region and oceans. The highest wet deposition fluxes are found in the Sahel and Gulf of Aden, with the maximum greater than 1200 g m$^{-2}$ a$^{-1}$ and a domain averaged value of 70 g m$^{-2}$ a$^{-1}$. Contrary to the dry deposition, wet deposition distributes heterogeneously over the Red Sea. While there is practically no contribution from wet deposition in the northern part of the basin (north of 21°N), this process is dominant in the southern part. The large amount of deposited dust in the southern part of the Red Sea is contributed from the Tokar delta source region, which is very active in summer [Jiang et al., 2009]. The time series of daily deposition rates (Figure 12c) reveals a single peak on 24 June, which is well correlated with the most intensive days of dust generation (Figure 3c). While dry deposition rates ranged from 3.9 to 6.0 Tg per day in the entire domain, wet and total depositions rates ranged from 1.9 to 3.1 and from 5.9 to 8.2 Tg per day, respectively. The overall total deposition for the entire domain during the simulation period is 49.2 Tg, which is about 90% of the emitted total dust amount reported earlier in section 4.2. The total dust deposition to the Red Sea reaches 0.2 Tg.

For comparison, the observed dust deposition rates from O’Hara et al. [2006] at 15 sites, located across northern Libya during a year-long field-based dust monitoring study, ranged from 38.6 to 311.0 g m$^{-2}$ a$^{-1}$, with an average of 129.1 g m$^{-2}$ a$^{-1}$, which is in good agreement with the simulated averaged value of 139.9 g m$^{-2}$ a$^{-1}$ presented earlier. Zhang et al. [1997] estimated dust deposition rates over Chinese deserts ranging from 7.7 to 2300 g m$^{-2}$ a$^{-1}$, well consistent with the simulated range from 10 to 2500 g m$^{-2}$ a$^{-1}$. 
5 Summary and Conclusion

In this study, we employed the WRF-Chem model to quantify emissions, transport, deposition, and the radiative effect of dust aerosols over the region of East Africa, the Red Sea, and the Arabian Peninsula during June 2012. The model simulations were conducted in two nested domains. The child domain was focused on the haboob area and used 2 km grid spacing to resolve deep convection and density currents. We identified a few meteorological mechanisms that cause intense dust emissions. The simulated heat low over north Sudan produced strong winds of up to 20 m s\(^{-1}\) that cause dust emission of 0.19 Tg between 07:00-08:00 UTC on 22 June 2012. The simulated intense dust emission flux over the Tokar delta region was associated with the strong near surface winds during the breakdown period of a nocturnal low-level jet. The cold pools of mesoscale convective systems were identified as another important mechanism causing dust emission. The considered haboob case lasted for more than 7 hours and lifted 0.31 Tg of dust. The emitted total dust amount over the entire domain and simulation period reached 54.5 Tg.

Model results and satellite data were analyzed to examine the horizontal and vertical transport pathways of East African dust across the Red Sea. The simulations indicate that the dust transport pathways are highly dependent on the synoptic meteorological patterns. When a high pressure system over the Arabian Peninsula is located to the south of a low pressure system over the East Mediterranean Sea and Syria in the mid-troposphere, East African dust plume trajectories go clockwise around the high pressure system and dust is preferentially transported in the eastward direction across the northern part of the Red Sea. A very persistent eastward dust transport at low altitudes was identified over the southern part of the Red Sea.
The AOD\textsubscript{[dust]} and DRE calculated from Eq. (4) show that the contribution of mineral dust in the total AOD is dominant during the simulation period, reaching 87\% of total AOD. The presence of dust particles in the atmosphere causes a significant reduction in solar radiation reaching the surface, with a domain average effect of -21.1 W m\textsuperscript{-2} and a maximum of -301.4 W m\textsuperscript{-2}. The SW radiative effect at the TOA is generally negative, with a domain average value of -7.4 W m\textsuperscript{-2}, however, the effect becomes positive over bright surfaces like desert areas. The atmosphere warms due to absorption of solar radiation with an average value of 13.8 W m\textsuperscript{-2}. The domain averaged LW effect of dust is warming at the surface reaching 9.0 W m\textsuperscript{-2}, cooling of the atmospheric column, -6.7 W m\textsuperscript{-2}, and warming at the TOA, 2.4 W m\textsuperscript{-2}.

Ground-based and satellite observations were exhaustively used to evaluate spatial and temporal variability of AOD at 600 nm simulated by WRF-Chem during the study period. Analysis of the time evolution of the AOD over six AERONET sites indicates close agreement between the temporal behavior of observed and model data with the correlation coefficient ranging from 0.33 to 0.87. However, large local disagreements in the AOD spatial distributions obtained from SEVIRI retrievals and simulated results were found over northeastern Sudan. The model underestimated by a factor of two observed AOD over this region even though the domain averaged values of the two fields are very similar, 0.46 for the model, and 0.47 for the SEVIRI retrievals. To further investigate this discrepancy, several additional simulations were conducted to study the sensitivity of averaged AOD field to model configuration and input datasets. The results from these experiments suggest that the discrepancy in the spatial distribution of AOD between satellite retrievals and simulations is more likely due, in part, to poor characterization of dust emissions in some areas of the modeled domain. It should also be noted that variability in the dust cycle variables from different model configurations and using different input datasets on average is less than 30\%.
which provides confidence to use the model for regional climate application over the modeled domain.

Our analysis shows that both dry and wet deposition processes contribute significantly to the removal of dust from the atmosphere. Dry deposition is the dominant removal mechanism of dust particles in the vicinity of the source regions and remote desert areas, with a domain averaged value of 139.9 g m$^{-2}$ a$^{-1}$, while wet deposition is the dominant removal mechanism of dust particles relatively far from the source regions, where precipitation is present with a domain averaged value of 70.2 g m$^{-2}$ a$^{-1}$. The simulated overall total deposition for the entire domain and simulation period is 49.2 Tg. The total dust deposition to the Red Sea reaches 0.2 Tg.

The spatial and temporal scales of the meteorological processes involved in the dust generation, transport, and deposition spans over few orders of magnitude. Therefore, high resolution analysis is extremely important for evaluating major uncertainties associated with the dust phenomena and better understanding the impact of dust on regional climate and ecological systems.

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British Atmospheric Data Centre (BADC) at

http://catalogue.ceda.ac.uk/uuid/d8a5e58e59eb31620082dc4fd10158e2. The simulation results and figures are available from the authors upon request (stoitchko.kalenderski@kaust.edu.sa)

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Figure 1. The parent (external) and nested (bounded by a blue line) domains with the 10 km and 2 km grid spacing. The orography (m) is shown by shading.
Figure 2. Synoptic weather conditions during the study period. ECMWF ERA-Interim geopotential height (m² s⁻²) of the 925 hPa (blue lines) and 500 hPa (red lines) isobaric surfaces, the estimated 24 hour accumulated rainfall (mm) from TRMM (color shading), 10 m winds, and 14°C 2 m dewpoint temperature (dashed green line) at 00 UTC on (a) 20 June, (b) 21 June, (c) 22 June, (d) 23 June, (e) 24 June, (f) 25 June
Figure 3. (a) Spatial distribution of time averaged dust emission flux [$\mu g m^{-2} s^{-1}$] over the high resolution nested domain and the study period, (b) dust source function, (c) time series of daily dust emission rates integrated over the nested domain (Tg day$^{-1}$).
Figure 4. Spatial distribution of near-surface winds (m s$^{-1}$, vectors) and dust emissions (µg m$^{-2}$ s$^{-1}$, shading) at 08:00 UTC on 22 June 2012.
Figure 5. (a) Spatial distribution of the simulated dust emission fluxes ($\mu$g m$^{-2}$ s$^{-1}$) and surface winds (m/s, vectors) at 08:00 UTC on 23 June 2012, (b) Vertical cross-section of the low-level jet located at the exit region of the Tokar Gap, (c) Average wind speed profiles at (18.61° N, 37.82° E). The white line and the white dot in Figure 5a denote the location of the vertical cross-section and the point of calculated average wind speed profiles.
Figure 6. SEVIRI RGB dust imagery (left column: a, d, g) - dust appears pink and clouds appear red (deep, high), black (shallow, high), orange (midlevel), and green (low). Spatial distribution of dust emission fluxes (grey shading at 200 μg m$^{-2}$ s$^{-1}$ intervals, starting at 100 μg m$^{-2}$ s$^{-1}$), simulated precipitation rates (mm, color shading), and wind vectors (m s$^{-1}$) near surface (center column: b, e, h). Vertical cross-sections of virtual potential temperature (K, shading), vertical wind speed (m s$^{-1}$, black contours), and TKE (m$^2$ s$^{-1}$, white contours), shown by blue lines on the maps in the central column (right column: c, f, i). Plots are shown for 14:00 UTC (top row: a, b, c), 16:00 UTC (middle row: d, e, f), and 18:00 UTC (bottom row: g, h, i) on 21 June 2012.
Figure 7. (a) True color image combined from two (10:40 and 10:45 UTC) Aqua MODIS retrievals on 24 June, (b) CALIOP total attenuated backscatter profiles (km$^{-1}$ sr$^{-1}$) at 532 nm, taken at about the same time as the MODIS observations, (c) Spatial distribution of sea level pressure anomalies (hPa, shading), dust load (g m$^{-2}$, black contours), and 72 hour trajectories (blue and green curves) released on 23 June 00:00 UTC, (d) Model-predicted vertical cross-section of dust concentration (µg m$^{-3}$) on 24 June 11:00 UTC along the CALIOP orbit track (Figure 7a, blue line).
Figure 8. AOD averaged for daytime from 6 UTC to 16 UTC and for the period of 20-26 June. (a) Spatial distribution of 600 nm AOD from WRF-Chem model, (b) Spatial distribution of 630 nm AOD from SEVIRI retrievals. The locations of the six AERONET sites are shown on the maps by black dots.
Figure 9. Time series of hourly average 600 nm AOD derived from WRF-Chem, AERONET, and SEVIRI datasets at six AERONET sites in the model domain. The correlation coefficients (r) and mean observed $\alpha$ are also shown at all six sites.
Figure 10. (a) MODIS 550 nm AOD (combined standard ocean and Deep Blue products) at 10:40-10:45 UTC on 25 June, (b) WRF-Chem 600 nm AOD at 11:00 UTC, (c) WRF-Chem 600 nm AOD [all except dust] at 11:00 UTC.
Figure 11. (upper row: a, b, c) Spatial distribution of daily averaged clear-sky direct DRE for SW, LW, and Net (SW+LW) radiation fluxes at the TOA [W m\(^{-2}\)]; (middle row: d, e, f) same as the upper row, but in the atmosphere; (bottom row: g, h, i) same as the upper row, but at the surface.
Figure 12. (a) Mean dry deposition fluxes over the model domain (g m\(^{-2}\) s\(^{-1}\)), (b) Mean wet deposition fluxes over the model domain (g m\(^{-2}\) s\(^{-1}\)), (c) Time series of daily dust deposition rates (Tg day\(^{-1}\)).