Optical and radiative properties of aerosols over Abu Dhabi in the United Arab Emirates

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The present study is on the aerosol optical and radiative properties in the short-wave radiation and its climate implications at the arid city of Abu Dhabi (24.42°N, 54.61°E, 4.5 m MSL), in the United Arab Emirates. The direct aerosol radiative forcings (ARF) in the short-wave region at the top (TOA) and bottom of the atmosphere (BOA) are estimated using a *hybrid* approach, making use of discrete ordinate radiative transfer method in conjunction with the short-wave flux and spectral aerosol optical depth (AOD) measurements, over a period of 3 years (June 2012–July 2015), at Abu Dhabi located at the southwest coast of the Arabian Gulf. The inferred microphysical properties of aerosols at the measurement site indicate strong seasonal variations from the dominance of coarse mode mineral dust aerosols during spring (March–May) and summer (June–September), to the abundance of fine/accumulation mode aerosols mainly from combustion of fossil-fuel and bio-fuel during autumn (October–November) and winter (December–February) seasons. The monthly mean diurnally averaged ARF at the BOA (TOA) varies from -13.2 W m^{-2} ($\sim -0.96 \text{ W m}^{-2}$) in November to -39.4 W m^{-2} (-11.4 W m^{-2}) in August with higher magnitudes of the forcing values during spring/summer seasons and lower values during autumn/winter seasons. The atmospheric aerosol forcing varies from $+12.2 \text{ W m}^{-2}$ (November) to 28.2 W m^{-2} (June) with higher values throughout the spring and summer seasons, suggesting the importance of mineral dust aerosols towards the solar dimming. Seasonally, highest values of the forcing efficiency at the surface are observed in spring ($-85.0 \pm 4.1 \text{ W m}^{-2} \tau^{-1}$) followed closely by winter ($-79.2 \pm 7.1 \text{ W m}^{-2} \tau^{-1}$) and the lowest values during autumn season ($-54 \pm 4.3 \text{ W m}^{-2} \tau^{-1}$). The study concludes with the variations of the atmospheric heating rates induced by the forcing. Highest heating rate is observed in June (0.39 K day^{-1}) and the lowest in November (0.17 K day^{-1}) and the temporal variability of this parameter is linearly associated with the aerosol absorption index.

1. Introduction

Atmospheric aerosols strongly influence the radiation budget of the earth–atmosphere system through

direct, indirect and semi-direct effects (Charlson *et al.* 1992; Haywood *et al.* 1997; Ackerman *et al.* 2000). Any change in the radiative equilibrium results in the alteration of the ambient temperature

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and/or the dynamical processes of the atmosphere. The term *radiative forcing* refers to the changes in the radiative balance of the earth–atmosphere system due to variations in specific components, such as greenhouse gases, aerosols, clouds, etc. While the radiative forcing by greenhouse gases is fairly uniform throughout the troposphere (Hansen *et al.* 1997), large spatio-temporal heterogeneity and short aerosol lifetime makes the aerosol properties highly complex and thereby induce substantial uncertainties in the forcings (IPCC 2013). In the global aerosol system, mineral dust is one of the major contributors, which accounts for ~ 30 to 50% of the total global aerosol burden (Satheesh and Krishnamoorthy 2005; Textor *et al.* 2006). Among the global dust sources, the deserts of North Africa and Middle East (MENA) contributes more than 50% of the total dust emission (Prospero *et al.* 2002). Though the dust loading in the atmosphere is largest over deserts, these particles can travel thousands of kilometers downwind of their source regions and their presence is observed even over the pristine oceanic regions (Prospero *et al.* 1983). According to IPCC (2013) estimates, the direct radiative forcing by dust aerosols varies between -0.3 and $+0.1 \text{ W m}^{-2}$ globally. Adequate knowledge on the direct radiative forcings of aerosols is needed to address their indirect and semi-direct radiative effects (García *et al.* 2012). Furthermore, the radiative effects of aerosols depend not only on mass loading, but also on their size distributions and chemical composition. Accurate information on aerosol radiative properties on regional scale is essential for a better understanding of their effects on weather and climate, especially over little-explored dust laden regions such as the Arabian peninsula. Such information is important for solar resources modelling for photo voltaic (PV) and concentrating solar power (CSP) applications too.

In this paper, we present our results on the optical and radiative properties of aerosols from measurements and radiative transfer modelling. The method we have employed for the estimation of aerosol direct radiative forcing is called the *hybrid approach*, since it combines both the conventional method using radiative transfer models and the *differential* method using measured irradiance. The analysis is based on the continuous and collocated measurements of spectral aerosol optical depths (AODs), and *in-situ* measured short-wave irradiance at surface and space-borne observations of the top of the atmospheric flux at the coastal location, Abu Dhabi, located at the south-west coast of the Arabian Gulf, for a continuous period of 3 years (from June 2012 to July 2015). The climate implications of the aerosol-induced atmospheric forcings are assessed in terms of the heating rate estimations.

2. Methodology

2.1 Data

2.1.1 Aerosol inversion products

AOD measurements at the AERONET station Masdar Institute (MI) at Abu Dhabi (24.42°N , 54.61°E , 4.5 m MSL) are carried out using a CIMEL Sunphotometer. The Sunphotometer measures spectral direct beam and diffuse irradiance (solar almucantar measurements of sky irradiances) in 8 channels from 340 to 1640 nm with full width at half maximum (FWHM) of 10 nm over 15-min intervals throughout the daylight hours (Holben *et al.* 1998). The direct Sun irradiance data are used to infer the AOD and total water vapour column with an uncertainty of ~ 0.01 (Holben *et al.* 1998; Smirnov *et al.* 2000) and 5–6%, respectively (Pérez-Ramírez *et al.* 2014). The spectral AODs along with almucantar measurements over a large range of scattering angles are used in the inversion algorithm developed by Dubovik and King (2000) and Dubovik *et al.* (2006) to retrieve the volume size distribution of aerosols, $(dV(r)/d)\ln(r)$ ($\mu\text{m}^3/\mu\text{m}^2$), in 22 logarithmically spaced bins with radii r from 0.05 to 15 μm , and spectral complex refractive index at 440, 675, 870, and 1020 nm (Dubovik and King 2000; Dubovik *et al.* 2006). The retrieval algorithm is based on realistic aerosol mixture including the spherical, non-spherical and mixed aerosols, with shape parameter ranging between 0.3 and 3.0, rather than the simple spheroid model. From the microphysical parameters, the algorithm computes the single scattering albedo (SSA) and the phase function of aerosols. The current retrieval algorithm consists of two independent, but complementary components of forward modelling and numerical inversion techniques. The forward modelling includes the linearized vector radiative transfer model (VLIDORT, Spurr 2006), a linearized Mie code and a linearized T-matrix code for calculating aerosol single scattering properties (Spurr *et al.* 2012), modules for calculating Rayleigh scattering and gaseous absorption, and a surface model for computing bi-directional reflectance and polarization distribution function (Spurr 2004). The algorithm has a provision to specify the bimodal aerosol size distribution and each of the modes are characterized by the total particle number or volume distribution, vertical profile, size distribution, and complex refractive index. Here a bi-modal lognormal function is assumed for the size distribution (Schuster *et al.* 2006) and size-independent refractive indices for three different aerosol models, adopted from Tanre' *et al.* (1999) are incorporated in the inversion algorithm.

The modules such as Mie, T-matrix, and VLIDORT are analytically linearized and coupled. Therefore, the forward model simulates radiance and/or polarization for a given spectrum as well as the Jacobians of the radiation fields with respect to input aerosol microphysics. The inversion model supplies these Jacobians together with observational error characterizations and *a priori* constraints to the statistical optimization procedure for the retrieval. Information content and error analysis are also included in this method along with the inversion (Dubovik *et al.* 2000; Xu and Wang 2015).

The fine mode AOD (FAOD), coarse mode AOD (CAOD), and the fine model fraction (FMF), which is defined as the ratio of fine mode AOD to the total AOD (at 500 nm), are computed from the spectral behaviour of the AOD using the spectral deconvolution algorithm (SDA) developed by O'Neill *et al.* (2001, 2003). The SDA assumes a bi-modal size distribution with pre-defined coarse and fine modes (Schuster *et al.* 2006; Waquet *et al.* 2009) as:

$$\frac{dV}{d \ln r} = \sum_{i=1}^2 \frac{V_0^i}{\ln \sigma_g^i \sqrt{2\pi}} \exp \left[-\frac{(\ln r - \ln r_v^i)^2}{2(\ln \sigma_g^i)^2} \right], \quad (1)$$

where V_0 , r_v , and σ_g are the total volume concentration, volume median radius, and geometric standard deviation, respectively. The superscript i indicates the size mode (1 for fine mode and 2 for coarse mode). The particle size ranges from 0.01 to 10 μm is assumed for the fine mode and from 0.05 to 20 μm for the coarse mode, both covering more than 99.9% of the total volume of an idealistic size range (0 to $+\infty$).

An initial guess for these quantities, in the form the AOD measurements at two spectral wavelengths, are provided to speed up the iterative inversion. Given *a priori* information on the aerosol size distribution and refractive indices, the aerosol extinction efficiency, Q_{ext} can be obtained for each of these modes using Mie code. Then, the AOD is related to the aerosol fine (V^f) and coarse (V^c) mode volumes as:

$$\tau_\lambda = \tau_\lambda^f + \tau_\lambda^c = \frac{3V_0^f Q_{\text{ext}}^f}{4r_{\text{eff}}^f} + \frac{3V_0^c Q_{\text{ext}}^c}{4r_{\text{eff}}^c}. \quad (2)$$

By applying the above equation to the AODs at any two spectral wavelengths, one can solve V_f and V_c . Considering the fact that the component fraction is more sensitive to the wavelength dependency of AOD at longer wavelengths, AODs at 870 and 1020 nm are chosen to determine the initial guesses of V_f and V_c .

The Angstrom exponent, α , is estimated using the Angstrom relation $\tau_\lambda = \beta \lambda^{-\alpha}$ in the wavelength range 440–870 nm, where τ_λ is the aerosol

optical depth at the wavelength λ , and β is the turbidity parameter (Angstrom 1964). Continuous observations of AODs and retrieved size distribution parameters for a period of 3 years from June 2012 to July 2015 are used for this study. The cloud-screened and quality-assured level 2.0 data points (Smirnov *et al.* 2000) for each day are averaged to get the corresponding daily means of AOD and volume size distribution.

Error budgeting

The AOD and other inversion products of AERONET are widely used to study the aerosol climatology (Dubovik *et al.* 2002) and development and validation of aerosol retrieval algorithms for various satellite sensors (Kaufman *et al.* 1997; Diner *et al.* 1998). However, the current inversion algorithm has many constraints in evaluating the aerosol data retrieved from other satellite sensors or chemistry transport model simulations and therefore improvement in the retrieval algorithms using multispectral polarization measurements at specific AERONET sites are still continuing (Xu and Wang 2015). The uncertainties in AOD and irradiance measurements using calibrated Cimel Sun Photometer are in the range 0.01–0.02 and 3–5%, respectively (Holben *et al.* 1998). Uncertainties in the current inversion products such as V_0 , r_{eff} and effective variance are $\sim 15\%$ (~ 10), ~ 8 –14% (8–14%), $\sim 50\%$ ($\sim 50\%$) for the fine (coarse) mode, respectively. The estimated error in the real part of refractive index is 0.025–0.05 and 0.03 for SSA (Dubovik *et al.* 2000). Therefore the products such as refractive index and SSA using the current algorithm are not recommended to use when AOD at 440 nm is lower than 0.4 (Dubovik *et al.* 2000). Since the retrieval is based on the information content in the particle optical scattering, the most relevant size parameters, regardless of the shape of the size distribution, should be effective radius and effective variance.

2.1.2 Irradiance measurements

In-situ observations

The global irradiance measurements at MI are carried out using an EKO Pyranometer (MS602) for the period June 2012 to April 2013 and using a Kipp and Zonen (CMP22) Pyranometer for the later period through July 2015. The EKO pyranometer is a high precision thermopile sensor with a flat spectral response over 285–3000 nm (95% transmission), while the CMP is flat over 200–3600 nm (95% transmission). The thermopile sensor is equipped with glass domes to block radiation

beyond 3 μm and a double dome construction effectively eliminates the infrared radiative loss as well as reducing changes in thermal resistance from sensor to ambient air with wind speed. The directional response of the sensor is $< \pm 10 \text{ W m}^{-2}$ (McArthur 2005). CMP22 Pyranometer sensor has a very high quality quartz domes for a wider spectral range with improved directional response and reduced thermal offsets. Because of the high optical quality of these domes, the directional error is reduced below 0.5% up to 80° solar zenith angle (Instruction Manual, Kipp and Zonen). In order to improve the reliability and accuracy of the measurement by reducing dust, raindrops and dew on the dome, the sensors are ventilated. With thermopile-based instruments, ventilation stabilises the temperature and suppresses thermal offsets.

Since we use two different pyranometers, an inter-comparison between the measurements has been performed for the period when simultaneous data are available (2015–2016). The cumulative frequency distribution (CFD) diagram for the difference between the fluxes (MS602 – CMP22) is shown in figure 1, along with the scatter plot between the two pyranometer measurements. The tails in the CDF curve result from semi-weekly cleaning of the dome when one or other instrument is shaded for one or two records. The scatter plot (figure 1b) shows excellent mutual correlation with R^2 of 0.99 and the slope of ~ 1.0 .

Satellite observations of irradiance

Clouds and the earth's radiant energy system (CERES) sensor, onboard the Terra satellite, observations of the top of the atmosphere short-wave flux data are used to estimate the top of the atmosphere radiative forcing. Terra has a sun-synchronous orbit with the orbit period of 98 min, and a ground-track repeat cycle of 16 days. The CERES sensor has three channels:

- (i) a short-wave (SW) channel to measure radiances in the SW region (wavelengths between 0.3 and 5 μm),
- (ii) a long-wave channel to measure earth-emitted thermal radiation, in the 8–12 μm , the *window* region, and
- (iii) a total channel to measure all wavelengths of radiation (Wielicki *et al.* 1996).

The resolution of CERES-Terra is 20 km at nadir. The Single Scanner Footprint TOA/Surface Fluxes and Clouds (SSF) product ($1^\circ \times 1^\circ$) are considered for this study. As detailed in Loeb *et al.* (2003, 2005) and Geier *et al.* (2001), the CERES SSF product combines CERES radiances and fluxes with scene information (e.g., cloud and aerosol properties) from coincident high spatial and spectral

resolution MODIS measurements, and meteorological information from the Global Modeling and Assimilation Office (GMAO)'s Goddard Earth Observing System (GEOS) Data Assimilation System (DAS) (Suarez 2005). The uncertainty in SW radiance is approximately 2.75% for $1^\circ \times 1^\circ$ gridded data, which corresponds to a 24-hr averaged regional SW TOA flux uncertainty of $\sim 1 \text{ Wm}^{-2}$ (Loeb *et al.* 2005).

2.1.3 Columnar ozone

The ozone monitoring instrument (OMI) onboard the AURA satellite is a nadir-viewing, wide-swath, ultraviolet-visible imaging spectrometer that provides the columnar ozone (O_3) concentrations along with other trace gases with global coverage of once per day and spatial resolution of $13 \text{ km} \times 24 \text{ km}$. OMI uses the data retrieval algorithm developed for the total ozone monitoring spectrometer (TOMS) developed by NASA, but with a differential optical absorption spectroscopy (DOAS) technique, in order to ensure continuity of ozone trends (Levelt *et al.* 2006; Koukouli *et al.* 2012). Experience with TOMS and DOAS suggests that the algorithms are capable of producing total ozone with a root-mean-square error (RMSE) of about 2%, depending on global coordinates (Levelt *et al.* 2006). The uncertainty in the OMI measured ozone data is reported as 2% (Veefkind *et al.* 2006).

2.1.4 Aerosol absorption index

Aura OMI aerosol absorption index (AAI) (collection 3, $1^\circ \times 1^\circ$ gridded data) data are used in this study to quantify the amount of absorbing aerosols at the measurement site. The near UV method of aerosol characterization uses space measurements at two channels in the near UV region to detect aerosol absorption (Torres *et al.* 1998). The OMI near UV aerosol algorithm (OMAERUV) uses the radiances measured at 354 and 388 nm to calculate the AAI (Torres *et al.* 1998). The algorithm uses pre-calculated reflectances at TOA for a set of 21 aerosol models composed of three different aerosol types (dust, carbonaceous aerosols and sulfate-based aerosols) to retrieve AAI. Since the algorithm is sensitive to the elevated aerosol layers, CALIPSO observed climatological vertical profiles are used (Torres *et al.* 2013).

Near-zero values of AAI represent nearly aerosol-free atmosphere or the abundance of non-absorbing type of aerosols and clouds. Presence of UV-absorbing aerosols such as soot, desert dust, volcanic ash give rise to positive AAI values. More details of the algorithm are given elsewhere also (e.g., Torres *et al.* 2007, 2013).

Figure 1. The cumulative frequency distribution function of the difference between the fluxes (MS602-CMP22) and the scatter plot between the two pyranometer measured fluxes.

2.1.5 Meteorological data

Surface level meteorological data

Near-surface meteorological parameters such as wind speed, wind direction, temperature, relative humidity and precipitation are measured using meteorological sensors installed at 10 m height from the surface. The winds are measured by a two-axis sonic anemometer (Gill instruments-Wind Sonic) with an accuracy $\pm 0.2\%$ of the reading with a resolution of 0.01 m s^{-1} for wind speed, and with an error of $\pm 3^\circ$ and resolution of 1° for wind direction (Instruction Manual Campbell Scientific Inc.). Relative humidity and ambient temperature are measured using Vaisala humidity and temperature probe (HMP60), respectively.

NCEP/NACR reanalysis datasets

The climatology (averaged for the period 1981–2010) of monthly wind vectors from the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) global reanalysis datasets (Kalnay *et al.* 1996), are used for examining the synoptic scale wind vectors over the Arabian peninsula.

2.2 Aerosol radiative forcing estimation

The short-wave direct aerosol radiative forcing (ARF) at the bottom (BOA) and top of the atmosphere (TOA), and forcing efficiencies (RFE) are estimated using a *hybrid* approach, which combines the measurements of irradiance, at the surface

(using pyranometers) and TOA (using CERES satellite sensor), and the radiative transfer models for aerosol-free flux simulations (Conant 2000; Satheesh and Ramanathan 2000). In the conventional method, using radiative transfer models for the estimation of clear-sky flux, by giving aerosol parameters as input, the estimates are sensitive to the model assumptions of aerosol optical properties such as asymmetry parameter and single scattering albedo. In the present study, we use more accurate ground-based observations of irradiance data. To get the aerosol-free flux, we have employed the radiative transfer model as well. However, there are studies on aerosol forcing estimations using ground measured flux (in a spectral band) alone, by considering the lowest flux from each month of the data as a reference level for that particular month, and assumes the day-to-day variation of the irradiance measured within the month are due to the changes in aerosol optical depth (Conant 2000). This approach is referred as *differential method* and is more suited for the pristine oceanic regions (e.g., Conant 2000), as it underestimates forcing due to the background aerosols. Hence, we have merged the two approaches (termed as the *hybrid method*), in order to minimize the errors due to model assumptions and any missing information on background aerosols.

At first, the pyranometer observations of global irradiance data at the measurement site are corrected for the changes in the eccentricity parameter and declination angle using the empirical relation:

$$I_{\text{corr}} = I_{\text{mea}} \left[\left(1 + 0.033 \cos \left(360 * \frac{n}{365} \right) \right) * \cos \vartheta_z \right]^{-1} \quad (3)$$

where I_{corr} is the corrected irradiance flux, I_{mea} is the measured flux, n is the day number and ϑ_z is the solar zenith angle. The corrected flux data for the solar geometry are averaged over 24 hr, to get the diurnal mean flux at the bottom of the atmosphere ($\langle F_{\text{BOA}\downarrow} \rangle$). The analyses are carried out strictly for cloudless days, in which the level-2 AOD data (cloud cleaned and quality assured) are available. Further, to avoid the contamination due to sub-visible cirrus clouds, the outliers lying beyond 2s level (s being the standard deviation) have been removed from individual day's AOD observations and the corresponding observations are eliminated from the flux measurements too. Furthermore, the days are included for the analysis only when AOD data are available throughout the entire sunshine hours (about 09:00–16:00 hrs). A total of 395 days of such clear sky day's data are used to estimate forcing. The clear sky *net flux* ($F_{\text{BOA}\downarrow} - F_{\text{BOA}\uparrow}$), with aerosols, at the surface are estimated from the mean-measured down-welling irradiance ($\langle F_{\text{BOA}\downarrow} \rangle$), using the relation $\langle F_{\text{net}}^{\text{aer}}_{\text{BOA}} \rangle = \langle F_{\text{BOA}\downarrow} \rangle (1 - \varepsilon)$,

where ε is the surface albedo. An earlier study on surface albedo in the short-wave region, based on the ASTER satellite observations in 15 m spatial resolution at Abu Dhabi revealed that the values vary between 0.3 and 0.35 for different terrains with higher values over sandy areas in comparison with vegetated and/or residential areas (Frey *et al.* 2005). In addition, based on the MODIS satellite observations, Satheesh *et al.* (2006) have reported that the surface albedo are in the range of 0.3–0.4 for the arid regions of northern Africa and 0.25–0.35 for Saudi Arabia (Satheesh *et al.* 2006). Since the measurement site is located in a sandy area with little vegetation, a default surface albedo of 0.35 is used for the measurement site, Masdar Institute, to carry out SBDART simulations.

The Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) model (Ricchiazzi *et al.* 1988) is used to estimate the aerosol-free flux. The model is based on the widely used and versatile radiative transfer codes, Discrete Ordinate Radiative Transfer (DISORT) (Stamnes *et al.* 1988), LOWTRAN (Kneizys *et al.* 1983) and MODTRAN (Berk *et al.* 1989). In the present study, the model simulations have been carried out for the measurement site to estimate the aerosol-free short-wave flux at surface, throughout the day at hourly intervals, for all the cloudless days during the study period. In order to perform the flux simulations using SBDART, atmospheric profiles of temperature, pressure, columnar ozone, water vapour and surface reflectance are required. Rayleigh scattering coefficient and Rayleigh optical depth estimations are based on the methodology of Liou (1980) and Shettle *et al.* (1980), respectively. The standard tropical atmospheric profiles of temperature, pressure and molecular number density are used for the estimations McClatchey *et al.* (1972). Spectrally uniform surface albedo of 0.35 is considered for the simulations. Since water vapour and columnar ozone indicated significant temporal variability during the course of a year (0.5–5.0 cm for water vapour and 221–315 DU for ozone at the measurement site as shown in figure 2) and are sensitive to the flux simulations, the aerosol-free flux estimations are carried out with the observed water-vapour and OMI observations of columnar ozone.

From the simulated hourly flux, the corresponding daily averages are estimated to get the daily mean aerosol-free flux at the BOA ($\langle F_{\text{net}}^{\text{no}}_{\text{BOA}} \rangle$). Thus, the direct short-wave aerosol radiative forcing at the surface is estimated by subtracting the mean modelled flux for aerosol free atmosphere ($\langle F_{\text{net}}^{\text{no}}_{\text{BOA}} \rangle$) from the corresponding daily mean net flux with aerosols at BOA ($\langle F_{\text{net}}^{\text{aer}}_{\text{BOA}} \rangle$) as:

$$\text{ARF}_{\text{BOA}} = \langle F_{\text{net}}^{\text{aer}}_{\text{BOA}} \rangle - \langle F_{\text{net}}^{\text{no}}_{\text{BOA}} \rangle. \quad (4)$$

The uncertainties in the aerosol-free flux estimations using discrete ordinate radiative transfer code arises from the inaccuracies in column water vapour, ozone and surface albedo. The uncertainty of $\sim 100\%$ in the column water vapour content is expected to induce an error smaller than $\pm 2\%$ on the surface short-wave irradiance (Gautier and Landsfeld 1997). An uncertainty of 2% (± 10 DU) in the column ozone will induce an error of $0.17 \pm 0.04 \text{ Wm}^{-2}$ (Lee *et al.* 2007) in the short-wave flux estimations.

In order to examine the accuracy of the flux estimates using SBDART, the global horizontal irradiance (GHI) modelled using SBDART in hourly intervals for relatively low turbid days ($\text{AOD} \leq 0.1$ at 500 nm) are compared with the observed GHI. AOD values lower than 0.05 at 440 nm can be considered as clean background conditions (Toledano *et al.* 2007) and when AODs between 0.04 and 0.06 (at 400 nm) indicate the pure atmospheric conditions (Smirnov *et al.* 1995). Since our measurement site is located in the Middle East with significant impact of dust aerosols with comparatively higher values of AOD, our observations are mostly between ~ 0.1 and 1.0 at 500 nm . However, we observe a few low values of AOD (≤ 0.1) during November–January period. One of such comparative study for December 19, 2013, when the AOD at 500 nm is ~ 0.07 , as a representative day is shown in figure 3. Agreement between the two fluxes is observed with a root mean error of $\sim 2 \text{ W m}^{-2}$ (1%).

Then, the net radiative flux at the top of the atmosphere are estimated from the difference between the outgoing short-wave flux without aerosols and the flux with aerosols under cloud-free conditions. Therefore, the short-wave aerosol direct

radiative forcing at the top of the atmosphere (ARF_{TOA}) can be expressed as:

$$\begin{aligned} ARF_{TOA} &= \langle F_{net_{TOA}}^{aer} \rangle - \langle F_{net_{TOA}}^{no} \rangle \\ &= (F_{TOA\downarrow} - F_{TOA\uparrow}^{aer}) - (F_{TOA\downarrow} - F_{TOA\uparrow}^{no}) \\ &= F_{TOA\uparrow}^{no} - F_{TOA\uparrow}^{aer}. \end{aligned} \quad (5)$$

Here, $F_{TOA\uparrow}^{no}$ and $F_{TOA\uparrow}^{aer}$ are the CERES measured outgoing solar irradiance without aerosols and the flux with aerosols under cloud-free conditions at top of the atmosphere in the SW region. From the radiative forcing estimations from the top and bottom of the atmosphere, the net atmospheric forcing of aerosols, which is defined as the difference

Figure 3. Diurnal variation of the measured and modelled (SBDART) GHI on a clean day (19 December 2013) with very low AOD (< 0.1). The vertical lines represent the local sunrise and sunset times.

Figure 2. The day-to-day variation of (a) columnar water vapour and (b) total column ozone at the measurement site.

in forcings between the top and bottom of the atmosphere, is estimated as:

$$ARF_{ATM} = ARF_{TOA} - ARF_{BOA}. \quad (6)$$

The atmospheric forcing indicates the amount of radiative flux absorbed by aerosols. This amount of flux is trapped in the atmosphere and will induce perturbations in the climate. The climate implications of aerosols are assessed in terms of the atmospheric heating rate. The heating rate is estimated from ARF_{ATM} , following Liou (2002), as:

$$\frac{\partial T}{\partial t} = \frac{g}{C_p} \frac{ARF_{ATM}}{\Delta P}, \quad (7)$$

where $\partial T/\partial t$ is the heating rate, C_p is the specific heat capacity of air at constant pressure, g is the acceleration due to gravity and P is the atmospheric pressure. Since the major amount of aerosols is present in the lower tropospheric level and contributes to atmospheric heating, the altitude is taken as 3 km, following Ramanathan *et al.* (2001) and Lubin *et al.* (2002). Therefore, ΔP is considered as 300 hPa, which is equal to the pressure difference between surface and 3 km.

3. Prevailing meteorology

The prevailing meteorological conditions have an important role in the production, size transformation, transport and deposition of aerosols. In view

of this, the prevailing weather parameters at the measurement site and the synoptic scale meteorology over the Arabian peninsula are reviewed.

3.1 Variation of weather parameters

Surface meteorological parameters measured using an automatic weather station collocated at the measurement site include temperature, humidity, wind speed and wind direction. These parameters are plotted to show the day-to-day variation for the period January 2012–July 2015, and are shown in figure 4. During the study period, the daily mean temperature varies from 13.6°C/56°F (January 23, 2012) to 40°C/104°F (July 18, 2012). The variation of mean daily temperature follows a periodic pattern with a summer high value of 35.5°C in July and a low winter value of 18°C in January. A near inverse pattern of variability is observed for relative humidity with highest value during January (63%) and the lowest during July (47%). Since relative humidity is a function of ambient temperature (Givoni 1969), this observed pattern is quite expected. Moderate surface wind speeds are observed throughout the study period (2–3 m/s) with comparatively lower speeds during winter months. The direction of the prevailing winds is found to shift from south-easterly to south-westerly as the season changes from winter to summer. Based on the prevailing meteorological conditions, the year has been divided into four

Figure 4. The near-surface meteorological parameters, such as (a) T, temperature; (b) RH, relative humidity; (c) WS, wind speed; and (d) WD, wind direction; averaged for each day, measured at the site during January 2012–July 2015.

seasons, namely, summer (June–September), autumn (October–November), winter (December–February) and spring (March–May).

3.2 Synoptic meteorology

The prevailing synoptic meteorological conditions have an important role in the temporal variability of local aerosol properties. Arabian peninsula is among the hottest and driest climatic zones in the world (Oliver and Hidore 1984; Ouarda *et al.* 2014). The hot desert conditions induce strong seasonal changes in the wind pattern (Parajuli *et al.* 2013). Climatological monthly mean wind vectors at 850 hPa obtained from NCEP/NCAR reanalysis datasets are shown in figure 5. The measurement location is denoted by the black dot in the figure. Seasonal changes in the wind pattern are evident: from northerlies/north-easterlies to easterlies/south-easterlies over the Arabian peninsula, as the season progresses from summer to winter. In the area surrounding the measurement site, the prevailing north-westerlies during June/July change to north-easterlies by August. Although, the wind speed gradually decreases towards September, the wind direction remains more or less the same. By October, the pattern changes drastically, and an anti-cyclonic circulation centered on the measurement location appears during November. The center of the high-pressure system moves southwards and it reaches the center of the Arabian peninsula by January. The weakening pattern persists through February, and thereafter, the circulation disappears by March. The prevailing weak southerlies of April change by May and strong north-westerlies are re-established by June.

3.3 Transport and advection

3.3.1 AOD weighed trajectory analysis

The changing synoptic scale meteorology has a significant impact in the aerosol loading, as the local meteorological parameters affect ambient aerosols through the production, transport and removal mechanisms (e.g., Megaritis *et al.* 2014). Among these aforementioned processes, long-range transport of aerosols are important, particularly at locations away from active source regions. Therefore, the potential advection pathways of aerosol and its role in modulating the aerosols at the measurement site are identified using the AOD weighed trajectory (AWT) analysis. The AWT method has been used to quantify the spatial distribution of the potential sources contributing to the observed aerosol loading at a given receptor location (Hsu *et al.* 2003; Gogoi *et al.* 2009; Vinoj *et al.* 2010). In AWT technique, air-mass back-trajectories

reaching the receptor site are weighted on the basis of the mean value of AOD (at 500 nm in this case) at the measurement site. The area around the measurement site is divided into grids, where each grid point gets a weighed AOD value (τ_{ij}) obtained by averaging sample data observed at the receptor site when the associated trajectory crosses the grid area (i, j), as described in the following equation.

$$\tau_{ij} = \frac{1}{\sum_{l=1}^M t_{ijl}} \sum_{l=1}^M \tau_l t_{ijl}, \quad (8)$$

where M is the total number of trajectories, l is the trajectory index, τ_l is the measured AOD at the receptor location (Abu Dhabi, in this study) associated with the trajectory l , and t_{ijl} is the residence time of AOD in trajectory l in grid cell (i, j). The resulting weighed AOD values at each grid represent the AOD that can be expected at the receptor site at a given time if the trajectories are passed through the spatial grid. This technique assumes that the local contribution at the receptor site is very small, or in other words, the receptor site is not the source point.

The seasonal variation of the AOD weighed trajectory patterns using isentropic air-mass back-trajectories arriving at 1800 m (mostly above the boundary layer, where the aerosols can be lifted by convection and transported over long distance) is shown in figure 6. Though the advection pattern remains more or less the same over the entire lower atmospheric height range, the aerosols within the boundary layer contribute most to the AOD.

The AWT pattern show the following features. During summer, the observed AODs are mainly affected by the contribution of the cluster of trajectories from the north-western continental locations. As these locations are mostly arid and semi-arid with weather conditions favourable to mineral dust production, such trajectories would be expected to carry dust aerosols to the measurement site. Towards autumn, the pattern changes drastically, with highest contribution to AOD by the local sources (~ 0.2), which are confined to the north-eastern part of the Middle East and the Arabian Gulf. During November, most of the trajectories have long traverses through north/northwest Asian regions such as Iran, Iraq, Turkey, Syria, and some of those trajectories even originate from far away locations such as Western Europe or north-west Africa. The AOD contribution of the aforementioned long trajectories to the measured AOD is also high (>0.05). During spring season, advection from North Africa along the southern Arabian peninsula is seen to contribute significantly to the seasonal pattern.

Figure 5. Long-term monthly mean climatology of wind vectors (NCEP derived) at 850 hPa over the Arabian peninsula. The measurement location, Abu Dhabi, is denoted by the black dot.

Figure 6. Seasonal variation of the AOD concentration weighed trajectory analysis (AWT) at the measurement site showing the potential advection pathways of aerosols at the measurement site.

4. Results and discussion

4.1 Optical properties

4.1.1 AOD and Angstrom parameters

The temporal variation of the daily mean AOD (at 500 nm), Angstrom wavelength exponent, α and turbidity parameter, β are shown in figure 7. Among the daily mean values, the lowest AOD value is observed on January 18, 2013 (~ 0.07 at 500 nm) and the highest value is observed on September 19th (1.56 at 500 nm), the latter being associated with a dust event. Dust storm episodes are more frequent during the period April–September, and this is clear from the sporadic increases of AOD in the time series data. The values of AODs are significantly higher during spring and summer seasons compared to the other two seasons.

The parameters such as fine mode AOD (FAOD) and coarse mode AOD (CAOD) are also examined to understand the variability of particle size distribution. The CAOD shows the same pattern of variability as that of the AOD (at 500 nm), with slightly lower magnitudes. Whereas a nearly reversed pattern (with higher values during winter/autumn and lower values during summer/spring

seasons), is observed for the FAOD. The FAOD (CAOD) varies from 0.02 to 1.0 (0.1 to 0.5) with values close to 0.5 (1.0) during winter/autumn (spring/summer). In general, the size distributions of aerosols are expressed in terms of α and FMF. The calculations of α using all the seven wavelengths (340–1020 nm) are sensitive to the fraction of aerosols in the fine mode, but not to the fine mode effective radius. Therefore, this multi-wavelength α is not such a rigorous indicator of the fine mode. Rather, $\alpha \geq 2$ correspond to fine mode fractions between 0.5 and 1, and $\alpha \leq 1$ correspond to fine mode fractions ≤ 0.5 (Schuster *et al.* 2006).

Figure 7 shows similar pattern of variability for FMF and α . In general, the values of α are less than 1.0 during spring/summer seasons and greater than 1.0 during the autumn/winter seasons. Similarly, FMF shows higher values during winter and autumn seasons ($>50\%$) and lower values during the dust-dominated summer and spring seasons ($<50\%$). Higher values of α are associated with the higher FMF and show a linear relationship with a correlation coefficient of ~ 0.90 (figure not shown). Occasionally the values of α (FMF) exceed 1.5 (90%). The Angstrom exponent, α , is a qualitative indicator

Figure 7. The day-to-day variation of (a) the aerosol optical properties at 500 nm (AOD), (b) fine mode aerosol optical depth at 500 nm (FAOD), (c) coarse mode aerosol optical depth at 500 nm (CAOD), and (d) Angstrom wavelength exponent (α) at the AERONET station, Masdar Institute, during the period June 2012–July 2015.

of aerosol particle size, with values > 2 indicating the abundance of small particles, mostly anthropogenic in origin, and values < 1 indicating large particles like sea salt and dust (Eck *et al.* 1999). The values in between 1 and 1.5 suggest significant contributions from both categories of particles. In this case, the majority of α values, in any year, indicate coarse mode dominance (more than 75% of the values of $\alpha \leq 1$). However, the remaining 25% depict higher α values (mostly between 1 and 2) occur during November–January, indicates significant contribution from anthropogenic sources. Although, the estimation of the spectral curvature of α (derivative of α) provides additional information on the size distribution of aerosols, it does not convey any extra information when $\alpha \leq 1$ and $\alpha \geq 2$, as in this case. Therefore, investigations on spectral curvature are not performed in this study.

From the daily mean values, the corresponding climatological monthly mean values are estimated and are presented in figure 8. On an average, AOD exhibits higher values during summer (0.6 for July/August) and lower during winter (0.2 for December). A similar pattern is observed for CAOD with slightly lower magnitude of the values.

FAOD shows a slightly different pattern with relatively higher values during the latter part of summer and autumn. The values remain more or less the same even during winter. However, the values are lowest towards the spring and early summer. The variations of Angstrom exponent, α and fine mode fraction show a continuous increase from their lowest values in April (0.5 for α and 30% for FMF) to highest values (1.2 for α and 61% for FMF) during November, and decreasing thereafter. A similar pattern of variability is observed for the microphysical properties of aerosols at other locations over the Middle East, (e.g., Sabbah and Hasan 2008). However, the values of α during winter/autumn seasons are comparable or even higher than those observed at various urban locations across Asia and Africa with similar seasonal variations (Nakajima *et al.* 1996; Moorthy *et al.* 2005, 2007; Satheesh and Krishnamoorthy 2005; Satheesh *et al.* 2006; Beegum *et al.* 2008; Wang *et al.* 2008; Alam *et al.* 2011; Lodhi *et al.* 2013; Khoshshima *et al.* 2014), with significant amount of fine mode aerosols from fossil-fuel and bio-fuel combustion. This indicates an anthropogenic aerosol dominance at the measurement site during the two

Figure 8. Monthly mean aerosol optical parameters estimated from the daily mean values (a) AOD, (b) fine mode optical depth (FMAOD), (c) coarse mode optical depth (CMAOD) and (d) Angstrom wavelength exponent, α .

seasons. The pattern of variability of the coarse mode AOD is similar to that of AOD at 500 nm with higher values during summer and spring seasons and lower values during winter and autumn seasons. Similar observations are also reported from many continental and island locations when the aerosols are dominated by coarse mode dust particles or sea-salt (Prospero *et al.* 2002; Dey *et al.* 2004; Moorthy *et al.* 2007; Beegum *et al.* 2008). The measurement site is being surrounded by deserts (Rub Al Khali in the south, Ad Dahna and Nafud deserts in the north/north-west), the dust aerosol contribution is significant. The observations indicate that aerosol microphysics undergo large seasonal variations with coarse mode natural aerosol dominance (mainly dust) during spring and summer seasons and fine/accumulation mode aerosols (from bio-fuel and fossil-fuel combustions) during autumn and spring seasons.

4.2 Physical properties

The contour map showing the monthly mean volume size distributions of the columnar aerosols (averaged over the study period) at the measurement site is given in figure 9 (top panel). In general, a bi-modal distribution is observed throughout the year with varying amplitudes for the modes. The primary mode is observed in the fine size range ($r_{m1} \sim 0.1\text{--}0.2 \mu\text{m}$) and secondary mode in the coarse regime ($r_{m2} \sim 2\text{--}3 \mu\text{m}$). Relatively high variability is observed for the coarse mode volume distribution (from $0.04 \mu\text{m}^3 \mu\text{m}^{-2}$ in December up to $0.24 \mu\text{m}^3 \mu\text{m}^{-2}$ in August) in comparison with

the fine mode (between 0.01 and $0.04 \mu\text{m}^3 \mu\text{m}^{-2}$). The relative dominance of coarse mode aerosol volume during late summer months (August–September) are even higher than that of the spring and early summer months. Whereas the fine mode volume distributions remain high for summer and autumn seasons and decreases towards winter. The modal radius of the fine mode is not constant throughout the year, but a shift towards the lower size regime ($0.2\text{--}0.1 \mu\text{m}$) is observed from winter to spring with a lessening of the standard deviation of the mode. Conversely, the secondary mode width increases from winter to spring. This explains why the observed values of α and FMF are lower in April. The feature is persistent, with a further increase in the width and amplitude of the secondary mode to reach the highest values in August. The pattern changes drastically towards autumn (October and November) with a sharp decrease in the coarse mode volume distribution and enhancement in the fine mode width. The mode radius of the primary mode shifts back to $0.2 \mu\text{m}$. This obviously results in low aerosol loading in the vertical column with fine/accumulation mode dominated size distribution with highest value of α and FMF.

The effective radius (R_{eff}), which is defined as the ratio of the volume concentration (third moment) to the area (second moment) of the aerosol size distribution, is estimated using the relation:

$$R_{\text{eff}} = \frac{\int_r^{r_2} r^3 \frac{dN(r)}{d \ln r} d \ln r}{\int_r^{r_2} r^2 \frac{dN(r)}{d \ln r} d \ln r}. \quad (9)$$

Figure 9. Contour map showing the monthly mean aerosol size distributions (a) and the corresponding effective radius (R_{eff}) at the site (b), averaged over the study period.

Here $\frac{dN(r)}{d \ln r}$ is the aerosol number size distribution, which is related to the volume size distribution as:

$$\frac{dV(r)}{d \ln r} = V(r) \frac{dN(r)}{d \ln r} = \frac{dN(r)}{d \ln r} \frac{4}{3} \pi r^3. \quad (10)$$

The variation of the effective radius (figure 9, bottom panel) illustrates the variation of the seasonal changes in the size distribution of the columnar aerosols. Higher values of R_{eff} are observed during spring/summer and lower values are observed during winter/autumn seasons with lowest value in November. This seasonal pattern of effective radius further corroborates the changes in the size distribution and subsequent changes in the types of aerosols at the measurement site. It should be noted that as these parameters are derived from the columnar aerosol optical depth, therefore they cannot provide any information on the vertical heterogeneity.

4.3 Radiative properties

4.3.1 Direct radiative forcing

The 24-hr average short-wave radiative forcings at the top (TOA) and bottom (BOA) of the atmosphere

are estimated concurrently with the measured short-wave flux from both *in-situ* (at surface) and satellite (top of the atmosphere) observations and SBDART radiative transfer model simulations on cloud-free days. The ARF at the surface (ARF_{BOA}) are estimated from the difference in the flux between the net flux from the surface observations and SBDART simulated aerosol-free flux. Whereas for the ARF at the top of the atmosphere (ARF_{TOA}), CERES sensor onboard the TERRA satellite observed flux are used. From the forcing values at the top and bottom of the atmosphere, the atmospheric radiative forcings of the aerosols are estimated (ARF_{TOA}). The climate implications of these forcings are assessed in terms of the heating rate estimations (Liou 2002). Details are given in the methodology section. The radiative forcing efficiency, which is defined as the forcing per unit change in AOD, is estimated by dividing the ARF by the daily mean AOD (Meywerk and Ramanathan 1999; Satheesh *et al.* 1999).

Figure 10 shows the variation of the monthly mean forcing values at different regions of the atmosphere, along with the AOD at 500 nm.

Figure 10. The variation of monthly mean direct radiative forcing at top (TOA), bottom (BOA) of the atmosphere as well as the atmospheric forcings (ATM), averaged over the period June 2012–July 2015, along with AOD at 500 nm.

ARF_{BOA} is always negative due to the attenuation of solar radiation at the surface. A negative forcing at the TOA indicates the enhancement in back-scattering by scattering type of aerosols, leading to cool the earth–atmosphere system. Whereas a positive forcing by absorbing aerosols contribute to warming the atmosphere. The difference between TOA and surface radiative forcing introduces the atmospheric radiative forcing. The estimated BOA and TOA short-wave radiative forcing are found to be negative throughout the year: varies between 13.2 ± 2.2 (November) and $-39.4 \pm 3.9 \text{ W m}^{-2}$ (August) for BOA and -1 ± 1.0 and $-11.4 \pm 3.0 \text{ W m}^{-2}$ for TOA. In general, the ARF (both at the surface and TOA) is found to be higher (in absolute magnitudes) throughout the spring and summer seasons, especially from April to August ($< -30 \text{ W m}^{-2}$) and showed a sharp decrease towards autumn and thereafter a weak increase towards winter seasons. The higher forcing values indicate higher scattering of solar radiation due to high dust load in the atmosphere. The atmospheric forcing (ARF_{ATM}) due to the aerosols is found to be highest during June ($28.1 \pm 5.5 \text{ W m}^{-2}$) and lowest during November ($12.2 \pm 2.2 \text{ W m}^{-2}$). The highest dust load leads to the maximum positive atmospheric radiative forcing during June. However, the atmospheric forcing remains high throughout the spring and summer months. The short-wave radiative forcing at the BOA, TOA and at ATM over Abu Dhabi during spring and summer seasons are comparable to those reported from other continental locations with frequent dust storm occurrence. Previous findings on short-wave

radiative forcing values at BOA, TOA and ATM respectively for a few locations where the influence of dust aerosols (either local production or through long-range transport) are significant are given in table 1, along with our present study results. Highest seasonal mean forcing values (both at the surface, TOA and ATM in terms of absolute magnitudes) at Abu Dhabi are observed during the spring and summer seasons and the values are comparable to those from the other dust-laden atmospheres. The magnitude of forcing at the TOA is smaller than that at the surface, as the cooling due to the scattering of solar flux is greatly offset by warming due to the absorption of solar and thermal radiations (Miller and Tegen 1998). The cooling/warming effect in the earth's atmosphere potentially depends upon the absorbing properties of aerosols as well as underlying reflecting earth's surface Satheesh (2002). Based on AERONET and MODIS data, Satheesh *et al.* (2006) have observed similar seasonal variations in the surface forcing at Solar Village in Saudi Arabia. Furthermore, Markowicz *et al.* (2008) have reported an average surface forcing of -20 W m^{-2} during the period August–September 2004, based on the measurements from Arabian Gulf region during the UAE2 field campaign. The observed annual average forcing at the surface from the present study ($\sim -25 \text{ W m}^{-2}$) is compared to forcings observed in adjacent mainland locations with significant impact of dust aerosols (Mallet *et al.* 2009; Zhao *et al.* 2010; Valenzuela *et al.* 2012).

The forcing efficiency, defined as the derivative of the flux with respect to AOD (Satheesh and

Table 1. *The radiative forcing and forcing efficiencies for different locations.*

Study location	Study period	Radiative forcing (W m^{-2})			Forcing efficiency ($\text{W m}^{-2} \tau^{-1}$)		Heating rate (K day^{-1})	Reference
		TOA	BOA	ATM	TOA	BOA		
M'Bour, Senegal	2006	-8.1	-29.1	21.0	-15.7	-56.4		Derimian <i>et al.</i> (2008)
MENA		-12	-20	8				Thomas <i>et al.</i> (2013)
Negev, Israel		-2.1	-6.4	4.3	-22	-65		Derimian <i>et al.</i> (2008)
North Africa		+2 to +4	-15 to -45				-0.4 to 1.2	Satheesh <i>et al.</i> (2006)
Saudi Arabia	June-August 2001	+2 to +4	-30 to -70		-35.3 $\pm$ 3	-65	0.8 to 2.2	Satheesh <i>et al.</i> (2006)
West African coast			10 to 15					Li <i>et al.</i> (2003)
Afro-Asian region		-1.4 to 21	-46 to -110	46 to 115			0.3 to 0.5	Satheesh <i>et al.</i> (2007)
Delhi, India		-2.9 to -26	-29.5 to 087.5		-17 $\pm$ 3	-46 $\pm$ 3		Singh <i>et al.</i> (2010)
Kanpur, India	2015-2010	-7 to -5	-18 to -21	12 to 15	-17 to 22	-65 to -74	0.05 to 0.19	Prasad <i>et al.</i> (2007)
Granada, Spain	June-September 2000-2003	-1.4 to -2.6	-26.9 to -24.4	25.5 to 21.8			0.6 to 0.7	Valenzuela <i>et al.</i> (2012)
Trivandrum, India								Babu <i>et al.</i> (2007)
Arabian Sea	March-June 2005				-26.1 to -32.4	-52.8 to -62.8		Babu <i>et al.</i> (2008)
Arabian Sea	March-May 2006				-41 to -28	-72 to -36		Moorthy <i>et al.</i> (2009)
Lampedusa, Italy			-69.9 $\pm$ 3.4			-59.9 $\pm$ 2.6		di Sarra <i>et al.</i> (2011)
Taklimakan, China		14.11	-69.72	78.8	-	-		Huang <i>et al.</i> (2009)
Zhangye, China		-58 $\pm$ 25	-114 $\pm$ 40					Alam <i>et al.</i> (2014)
Toulon, France, 2008	April-June 2008	0.52 $\pm$ 1.69	-22.4 $\pm$ 8.9	21.8		-95.1 $\pm$ 10.3		Ge <i>et al.</i> (2010)
Abu Dhabi, UAE	December-February 2012-2015	-7.7 to -9.8	-61.8 to -64.4	54.1 to 54.6	-9.7 to -12.4	-78.2 to -81.5		Saha <i>et al.</i> (2008)
Abu Dhabi, UAE		-1.9 $\pm$ 0.22	-20 $\pm$ 4.3	18.4 $\pm$ 4.2	-7.7 $\pm$ 1.1	-79.2 $\pm$ 7.1	0.25 $\pm$ 0.06	Present study
Abu Dhabi, UAE		-6.3 $\pm$ 3.3	-32.2 $\pm$ 5.2	25.9 $\pm$ 2.2	-15.8 $\pm$ 6.6	-85.0 $\pm$ 4.1	0.35 $\pm$ 0.03	Present study
Abu Dhabi, UAE		-7.1 $\pm$ 2.7	-33.2 $\pm$ 5.6	26.1 $\pm$ 3.2	-13.3 $\pm$ 3.4	-63.7 $\pm$ 7.0	0.36 $\pm$ 0.04	Present study
Abu Dhabi, UAE	October-November 2012-2015	-1.3 $\pm$ 0.5	-16.0 $\pm$ 13.1	13.2 $\pm$ 1.4	-5.0 $\pm$ 1.7	-54 $\pm$ 4.3	0.18 $\pm$ 0.02	Present study

Ramanathan 2000; Conant 2000; Rajeev and Ramanathan 2001) are estimated at BOA, TOA and ATM and are shown in figure 11. The forcing efficiencies at surface and atmosphere show higher values (in absolute magnitudes) during winter and early spring and comparatively lower values during summer and autumn seasons. The highest values of RFE at the BOA (ATM) is observed during March with the values as high as -88 ± 7.0 (80 ± 5.7) W m^{-2} per AOD at 500 nm. The lowest value observed is during November and values are -51.1 ± 4.8 and -47.4 ± 3.9 $\text{W m}^{-2}\tau_{500}^{-1}$, respectively for surface and atmosphere. Larger negative surface forcing efficiencies indicate higher dust loading with higher absorption of solar radiation at the surface. The forcing efficiency at the TOA varies between -19.3 ± 2.9 $\text{W m}^{-2}\tau_{500}^{-1}$ (August) and -3.7 ± 1.3 $\text{W m}^{-2}\tau_{500}^{-1}$ (November). Seasonally, highest values of the forcing efficiency at the surface is observed in spring (-85.0 ± 4.1 $\text{W m}^{-2}\tau^{-1}$) followed closely by winter (-79.2 ± 7.1 $\text{W m}^{-2}\tau^{-1}$) and the lowest value is observed during autumn (-54 ± 4.3 $\text{W m}^{-2}\tau^{-1}$). At the TOA, the higher values of forcing efficiencies are observed during spring and summer seasons (-15.8 ± 6.6 and -13.3 ± 3.4 $\text{W m}^{-2}\tau^{-1}$) compared to those observed during winter and autumn seasons (-7.7 ± 1.1 and -5.0 ± 1.7 $\text{W m}^{-2}\tau^{-1}$). But the highest atmospheric forcing efficiency is observed during winter, followed by spring and lowest during autumn. Seasonal average forcing efficiencies at BOA during winter/autumn seasons are comparable with those from other urban/industrial locations with significant anthropogenic activities

(Ramanathan *et al.* 2007; Pandithurai *et al.* 2008; Srivastava *et al.* 2011; García *et al.* 2012; Esteve *et al.* 2014) and those during spring/summer are comparable to those of the mineral dust-dominated sites.

The climate response of the aerosol forcing is analyzed in terms of the heating rate and is found that the heating rate varies from 0.17 (in November) to 0.39 K day^{-1} (in June). In general, the dust loading in the atmosphere can lead to the enhancement of heating rate (Kim *et al.* 2006). Also, recent studies showed that the absorbing aerosols like soot can have significant warming effect, even if its amount is very low (Bond and Bergstrom 2006). Since the atmospheric forcing and heating rates depend on the scattering and absorbing properties of aerosols, aerosol absorption index (AAI) is a useful tool to infer the types of aerosols. The AAI is an index that detects the presence of UV-absorbing aerosols such as dust and soot. Higher or positive values of AI indicate UV-absorbing aerosols mainly mineral dust, smoke and volcanic aerosols and negative values are associated with non-absorbing aerosols such as sulfate and sea-salt aerosols (Torres *et al.* 2007). Thus, over the desert regions, the positive values represent dust sources and dust loading in the atmosphere.

The climatological monthly mean AAI, for the study period, obtained from OMI are shown in figure 12. Interestingly, the higher heating rates coincide with high AAI and a correlation coefficient of 0.89 (figure not shown) is obtained between the two. Highest values of AAI are observed during summer, followed closely by summer and lowest values during autumn. The lower values of

Figure 11. The variation of monthly mean aerosol radiative forcing efficiencies at TOA, BOA and ATM along with the heating rate.

Figure 12. Monthly mean aerosol absorbing index (AAI) derived from OMI, averaged for the period June 2012–July 2015 at Masdar Institute.

the heating rates and AAI (<1) are observed during autumn season (<1), corresponding to background aerosols with the less amount of absorbing types of aerosols (Torres *et al.* 2007). The highest value of heating rate during summer is a clear indicator of the prevalence of absorbing aerosols, mainly mineral dust during the seasons. However, the heating rates and forcing efficiencies are lower compared to those reported from East Asian or African deserts. The single scattering albedo of the mineral dust over Middle East is found to be higher in comparison with those reported from Africa or Asia, which indicates a lower forcing efficiency for the mineral dust over this region (Kubilay *et al.* 2003; Kim *et al.* 2005; García *et al.* 2012). This could be due to the difference in the dust mineralogy of the Middle East from African or Asian deserts (Sokolik and Toon 1999; Caquineau *et al.* 2002).

5. Discussions

The present study shows that the optical and radiative properties of aerosols at Abu Dhabi exhibit large seasonal variations. These variations are associated with the changes in the size distribution and types of aerosols during the course of a year with an overall dominance of coarse mode natural aerosols during summer/spring and fine/accumulation mode anthropogenic aerosols during winter/autumn seasons.

Although the Arabian peninsular region is prone to dust due to its arid nature, the present study indicates significant contributions from anthropogenic aerosols as well. Until 1950s, the most of the Arabian peninsula was barren desert land. The discovery of oil and gas in 1950s spurred rapid growth and industrialization turning the UAE, Qatar, Bahrain, Kuwait, Saudi Arabia, and Oman into some of the world's most affluent countries in just a few decades. The region experienced a very drastic change in the infrastructure developments such as roads, ports, airports, water and

electricity supplies, and other oil-based industries. This resulted in an exponential increase in the population, which in turn increased the urban and industrial activities as well (Shihab 2001). Vehicular emissions, industrial activities such as mining, refineries, manufacturing chemicals and plastics, and construction activities are the main sources of local pollution. Furthermore, marine aerosol may be significant, as the measurement location is < 1 km from the coastline. Thus the observed annual variability in the aerosol properties arises by the uneven contributions from the aforementioned source activities along with the long-range transport. The role of long-range transport (described in section 3.3) is found to have a significant role in modulating the aerosol properties.

Dust storms over the Middle East and North African locations are frequent during spring and summer seasons. The extremely hot and dry weather conditions are conducive to production of dust over the entire Arabian peninsula and North-African regions. Over the Middle East and west Asia, dust activity increases in March/April, peaks during the summer months of June and July, and weakens towards September (Shao 2009). Severe dust storms occurring in summer are associated with *Shamal* winds over the Middle East including the arid plains of west Asia and the Arabian peninsula, southern Iraq, Iran, Kuwait, the northwestern Persian Gulf, and the northeastern Arabian peninsula (Hamidi *et al.* 2013; Gherboudj and Ghedira 2014). The measurement site being located close to the deserts such as *Rub' Al Khali* (Empty Quarter), *Ad Dahana* and *Nafud* desert in the Arabian peninsula, local production of mineral dust along with air-mass trajectories that pass through other arid locations in the Middle East and North Africa (figure 6) would contribute a significant amount of airborne dust to the measurement location leading to highest aerosol loading and lowest α (and FMF) for the seasons of spring and summer. Prospero *et al.* (2002) identified two major advection pathways of dust aerosols over the Arabian peninsula,

the first one is from the valley between the Tigris and Euphrates rivers through the deserts of Iraq, Saudi Arabia and Iran, and the second from the coast of Oman. The present study shows that the high aerosol loading during summer is mostly modulated by the transported dust from the first source area. The increase in the volume distribution of both modes of aerosols observed (figure 9) could be attributed to the transported mineral dust. The resulting highest dust loading is obvious from the highest seasonal value of AOD and AAI. Heavy dust loading in the atmosphere during summer would lead to large reduction of the ground solar flux leading to highest aerosol forcing at the surface and highest heating rate in the atmosphere.

During autumn, the prevailing meteorological conditions, i.e., anti-cyclonic circulation centered on the measurement site (figure 5) with low surface winds and significant reduction in the surface temperature (figure 4), are unfavourable for the production of mineral dust. This would eventually remove the coarse mode regime from the size distribution, leading to low background aerosol concentrations with a Junge power-law aerosol size distribution with fine/accumulation mode abundance. The observed high values of α and FMF along with lower values of AOD are consistent with this finding. The AWT pattern (figure 6) also indicates the major share from local sources. However, the seasonal mean AOD is lowest during the season (~ 0.15) and the lowest value of AAI, representative of background aerosols with net scattering, results in lowest values of radiative forcing, forcing efficiencies and heating rate. In winter, local urban activities along with the advection from the polluted north-west Asian locations result in high levels of anthropogenic air pollution. A recent report published by the United Nations Economic and Social Commission for Western Asia concluded that the anthropogenic air pollution due to urban and industrial activities has been increasing for the past few decades over the north-west Asian countries. Tehran is one of the most heavily polluted cities in the west Asian regions where the level of pollution is highest during winter season. According to recent estimates, $\sim 45,000$ people die per year due to various diseases associated with poor air quality (Naddafi *et al.* 2012). Based on the satellite observations and modelling studies, Lelieveld *et al.* (2009) have reported that the Arabian Gulf region is increasingly plagued by photochemical smog, where the tropospheric O_3 concentrations exceed the air quality standards throughout the year. Long-range transport from Europe and west Asia is also implicated in the high levels of air pollution (Lelieveld *et al.* 2009). Thus, the local pollution along with the advected component from

other polluted west Asian locations, play a significant role in the observed high levels of anthropogenic aerosols at the measurement site during winter. However, it is worth noting that the values of α and FMF observed during January and February are lower than those observed during November. Since the December–February period is the wet season over the Arabian peninsula with relatively high rainfall amounts, an additional removal mechanism of aerosols is active during that season. Moreover, relative humidity reaches its highest values during winter and this in turn may lead to the hygroscopic growth of aerosols (Jeong *et al.* 2007; Eck *et al.* 2012). Thus, the particles in the accumulation mode regime can grow until reaching sufficient size to enter the coarse mode regime leading to lower values of FMF and α in comparison with those recorded in autumn season. The prevailing urban pollution leads to higher forcing efficiencies and heating rate. During long-range transport, aerosol properties, such as hygroscopicity and homogenous and/or heterogeneous mixing with other continental aerosols may change through chemical reactions and thus in turn leading to high atmospheric forcing and heating rate. Towards spring, the dust emission initiates from the arid/semi-arid areas over the Middle East and eventually get mixed up with prevailing urban pollution. Deepshikha *et al.* (2005) and Rodríguez *et al.* (2011) have showed that mixing of mineral dust with local and/or long-range transported pollution enhances the absorption efficiency of dust. In this study, we observe highest forcing efficiencies during March could be due to this mixing of the dust emitted during the seasons with the prevailing urban pollution.

Highest atmospheric forcing and heating rate are observed in the case of intense dust loading in the atmosphere (summer), followed closely by the dust mixed with urban pollution (spring) and then polluted continental (winter) and lowest for the clean background aerosols (autumn). Thus, it can be concluded that the observed temporal variations in the radiative properties of aerosols are strongly governed by the seasonal changes in aerosol size, type and abundance in the vertical column, and this in turn has many local and global climate implications. Satheesh and Ramanathan (2000) have reported that the heating can affect local vertical circulations and suggested modifications in large-scale monsoonal circulations. Heating the atmosphere and cooling the surface can change the boundary layer vertical temperature gradient and may be expected to reduce evaporation and cloud formation (Hansen *et al.* 1997; Ackerman *et al.* 2000) and weaken the hydrological cycle (Ramanathan *et al.* 2001).

6. Conclusions

Estimates of direct aerosol radiative forcing at the surface based on the collocated measurements of spectral aerosol optical depth and short-wave irradiances have been presented for a period of 3 years from June 2012 to July 2015 in Abu Dhabi, a north-eastern coastal location in the Arabian peninsula. Analysis of the inferred microphysical and radiative properties of aerosols revealed the following findings.

- The spectral aerosol optical depth (AOD) and the steepness of the AOD spectra were found to vary significantly throughout the year with higher aerosol loading and flatter spectra during spring/summer seasons and comparatively steeper spectra and lower values of AOD during autumn/winter seasons.
- The Angstrom wavelength exponent (α) and the fine mode fraction (FMF) attained their highest values during November (1.2 for α and 61% for FMF) and lowest values during April (0.5 and 30%), indicating significant seasonal variation of aerosol size and types.
- The volume size distribution of aerosols indicate coarse mode dominated size distribution during spring/summer seasons and a distinct fine/accumulation mode dominance during autumn/winter.
- The inferred aerosols at the measurement site indicate strong seasonal variations from the dominance of coarse mode mineral dust aerosols during spring/summer seasons, to the abundance of fine/accumulation mode aerosols.
- The observed diurnally averaged aerosol radiative forcing (ARF) at bottom (top) of the atmosphere varies from -13.2 W m^{-2} ($\sim -0.96 \text{ W m}^{-2}$) in November to -39.4 W m^{-2} (-11.4 W m^{-2}) in August with higher magnitudes of the forcing values during spring and summer seasons and lower values during autumn and winter seasons.
- Highest value of the forcing efficiency at the surface is observed in spring ($-85.0 \pm 4.1 \text{ W m}^{-2} \tau^{-1}$) followed closely by winter ($-79.2 \pm 7.1 \text{ W m}^{-2} \tau^{-1}$) and the lowest value is observed during autumn ($-54 \pm 4.3 \text{ W m}^{-2} \tau^{-1}$).
- The highest atmospheric heating rate is occurring during June (0.39 K day^{-1}) and the lowest in November (0.17 K day^{-1}) and the parameter is closely associated with the aerosol absorption index.

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References

- Ackerman A S, Toon O B, Stevens D E, Heymsfield A J, Ramanathan V and Welton E J 2000 Reduction of tropical cloudiness by soot; *Science* **288** 1042–1047.
- Alam K, Trautmann T and Blaschke T 2011 Aerosol optical properties and radiative forcing over mega-city Karachi; *Atmos. Res.* **101** 773–782.
- Alam K, Trautmann T, Blaschke T and Subhan F 2014 Changes in aerosol optical properties due to dust storms in the Middle East and southwest Asia; *Remot. Sens. Environ.* **143** 216–227.
- Angstrom A 1964 The parameters of atmospheric turbidity; *Tellus* **16** 64–75.
- Babu S S, Moorthy K K and Satheesh S K 2007 Temporal heterogeneity in aerosol characteristics and the resulting radiative impacts at a tropical coastal station. Part 2: Direct short wave radiative forcing; *Ann. Geophys.* **25** 2309–2320.
- Babu S S, Nair V S and Moorthy K K 2008 Seasonal changes in aerosol characteristics over Arabian Sea and their consequence on aerosol short-wave radiative forcing: Results from ARMEX field campaign; *J. Atmos. Sol.-Terr. Phys.* **70** 820–834, doi: 10.1016/j.jastp.2007.10.005.
- Beegum S N et al. 2008 Characteristics of spectral aerosol optical depths over India during ICARB; *J. Earth Syst. Sci.* **117**(S1) 303–313.
- Berk A, Bernstein L S and Robertson D C 1989 MODTRAN: A Moderate Resolution Model for LOWTRAN 7, GL-TR-89-C122, AD-A214 337.
- Bond T C and Bergstrom R W 2006 Light absorption by carbonaceous particles: An investigative review; *Aerosol Sci. Technol.* **41**(1) 27–47.
- Caqueneau S, Gaudichet A, Gomes L and Legrand M 2002 Mineralogy of Saharan dust transported over northwestern tropical Atlantic Ocean in relation to source regions; *J. Geophys. Res.* **107**, doi: 10.1029/2000JD000247.
- Charlson R J, Schwartz S E, Hales J M, Cess J A, Coakley Jr, Hansen J E and Hofmann D J 1992 Climate forcing by anthropogenic aerosols; *Science* **255** 423–430.
- Conant W C 2000 An observational approach for determining aerosol surface radiative forcing: Results from the first field phase of INDOEX; *J. Geophys. Res.* **105** 15,347–15,360.
- Deepshikha S, Satheesh S K and Srinivasan J 2005 Regional distribution of absorbing efficiency of dust aerosols over India and adjacent continents inferred using satellite remote sensing; *Geophys. Res. Lett.* **32** L03811, doi: 10.1029/2004GL022091.
- Derimian Y, Leon J F, Dubovik O, Chiapello I, Tanre D, Sinyuk A, Auriol F, Podvin T, Brogniez G and Holben B N 2008 Radiative properties of aerosol mixture observed during the dry season 2006 over M'Bour, Senegal (African Monsoon Multidisciplinary Analysis campaign); *J. Geophys. Res.* **113** D00C09, doi: 10.1029/2008JD009904.

- Dey S, Tripathi S N and Singh R P 2004 Influence of dust storm on the aerosol optical properties over Indo-Gangetic basin; *J. Geophys. Res.* **109** D20211, doi: 10.1029/2004JD004924.
- Diner D J, Beckert J C, Reilly T H, Bruegge C J, Conel J E, Kahn R A, Martonchik J V, Ackerman T P, Davies R, Gerstl S A W, Gordon H R, Muller J P, Myneni R B, Sellers P J, Pinty B and Verstraete M M V 1998 Multi-angle Imaging SpectroRadiometer (MISR) – instrument description and experiment overview; *IEEE Trans. Geosci. Remote Sens.* **36** 1072–1087.
- di Sarra A, Di Biagio C, Meloni D, Monteleone F, Pace G, Pugnaghi S and Sferlazzo D 2011 Short-wave and longwave radiative effects of the intense Saharan dust event of 25–26 March, 2010, at Lampedusa (Mediterranean Sea); *J. Geophys. Res.* **116** D23209, doi: 10.1029/2011JD016238.
- Dubovik O and King M D 2000 A flexible inversion algorithm for the retrieval of aerosol optical properties from Sun and sky radiance measurements; *J. Geophys. Res.* **105** 20,673–20,696.
- Dubovik O, Smirnov A and Holben B N 2000 Accuracy assessments of aerosol optical properties retrieved from AERONET sun and sky radiance measurements; *J. Geophys. Res.* **105** 9791–9806.
- Dubovik O, Holben B N, Eck T F, Smirnov A, Kaufman Y J, King M D, Tanré D and Slutsker I 2002 Variability of absorption and optical properties of key aerosol types observed in worldwide locations; *J. Atmos. Sci.* **59** 590–608.
- Dubovik O *et al.* 2006 Application of spheroid models to account for aerosol particle nonsphericity in remote sensing of desert dust; *J. Geophys. Res.* **111** D11208, doi: 10.1029/2005JD006619.
- Eck T F, Holben B N, Reid J S, Dubovik O, Smirnov A, O'Neill N T, Slutsker I and Kinne S 1999 Wavelength dependence of the optical depth of biomass burning, urban, and desert dust aerosols; *J. Geophys. Res.* **104** 31,333–31,349.
- Eck T F *et al.* 2012 Fog- and cloud-induced aerosol modification observed by the aerosol robotic network (AERONET); *J. Geophys. Res.* **117** D07206, doi: 10.1029/2011JD016839.
- Esteve A R, Estellés V, Utrillas M P and Martínez-Lozano J A 2014 Analysis of the aerosol radiative forcing over a Mediterranean urban coastal site; *Atmos. Res.* **137** 195–204.
- Frey C M, Rigo G and Parlow E 2005 Investigation of the Daily Urban Cool Island (UCI) in two coastal cities in an arid environment: Dubai and Abu Dhabi (UAE); *Remote Sens. Spat. Info. Sci.* **36**, Part 8/W27, 5S.
- García O E *et al.* 2012 Validation of AERONET estimates of atmospheric solar fluxes and aerosol radiative forcing by ground-based broadband measurements; *J. Geophys. Res.* **113** D21207, doi: 10.1029/2008JD010211.
- Gautier C and Landsfeld M 1997 Surface solar radiation flux and cloud radiative forcing for the atmospheric radiation measurement (ARM) Southern Great Plains (SGP): A satellite, surface observations, and radiative transfer model study; *J. Atmos. Sci.* **54**(10) 1290–1307.
- Ge J M, Su J, Ackerman T P, Fu Q, Huang J P and Shi J S 2010 Dust aerosol optical properties retrieval and radiative forcing over northwestern China during the 2008 China–U.S. joint field experiment; *J. Geophys. Res.* **115** D00K12.
- Geier E B, Green R N, Kratz D P, Minnis P, Miller W F, Nolan S K and Franklin C B 2001 Single satellite footprint TOA/surface fluxes and clouds (SSF) collection document; Geophysics Series No. 84, Academic Press.
- Gherboudj I and Ghedira H 2014 Spatiotemporal assessment of dust loading over the United Arab Emirates; *Int. J. Climatol.* **34**(12) 3321–3335.
- Givoni B 1969 *Man, Climate and Architecture*; Elsevier, Amsterdam, London.
- Gogoi M M, Moorthy K K, Babu S S and Bhuyan P K 2009 Climatology of columnar aerosol properties and the influence of synoptic conditions: First-time results from the northeastern region of India; *J. Geophys. Res.* **114** D08202.
- Hamidi M, Kavianpour M R and Shao Y 2013 Synoptic analysis of dust storms in the Middle East; *Asia-Pac. J. Atmos. Sci.* **49** 279–286.
- Hansen J, Sato M and Ruedy R 1997 Radiative forcing and climate response; *J. Geophys. Res.* **102** 6831–6864, doi: 10.1029/96JD03436.
- Haywood J M, Roberts D L, Slingo A, Edwards J M and Shine K P 1997 General circulation model calculations of the direct radiative forcing by anthropogenic sulfate and fossil-fuel soot aerosol; *J. Climate* **10** 1562–1577.
- Holben B N *et al.* 1998 AERONET – A federated instrument network and data archive for aerosol characterization; *Remote Sens. Environ.* **66** 1–16.
- Hsu N C, Herman J R and Tsay S C 2003 Radiative impacts from biomass burning in the presence of clouds during boreal spring in southeast Asia; *Geophys. Res. Lett.* **30** 1224.
- Huang J, Minnis P, Yi Y, Tang Q, Wang X, Hu Y, Liu Z, Ayers K, Trepte C and Winker D 2007 Summer dust aerosols detected from CALIPSO over the Tibetan Plateau; *Geophys. Res. Lett.* **34** L18805, doi: 10.1029/2007GL029938.
- IPCC 2013 Climate Change 2013: The Physical Science Basis (eds) Stocker T F, Qin D, Plattner G K, Tignor M, Allen S K, Boschung J, Nauels A and Xia Y, Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change.
- Jeong M J, Li Z, Andrews E and Tsay S C 2007 Effect of aerosol humidification on the column aerosol optical thickness over the atmospheric radiation measurement Southern Great Plains site; *J. Geophys. Res.* **112** D10202.
- Kalnay E *et al.* 1996 The NCEP/NCAR 40-year reanalysis project; *B. Am. Meteorol. Soc.* **77**(3) 437–471.
- Kaufman Y J, Tanré D, Remer L A, Vermote E F, Chu A and Holben B N 1997 Operational remote sensing of tropospheric aerosol over land from EOS moderate resolution imaging spectroradiometer; *J. Geophys. Res.* **102** 17051–17067.
- Khoshsima M, Bidokhi A A and Ahmadi-Gavi F 2014 Variations of aerosol optical depth and Angstrom parameters at a sub-urban location in Iran during 2009–2010; *J. Earth Syst. Sci.* **123**(1) 187–199.
- Kim D H, Sohn B J, Nakajima T and Takamura T 2005 Aerosol radiative forcing over east Asia determined from ground-based solar radiation measurements; *J. Geophys. Res.* **110** D10S22.
- Kim M K, Lau K M, Chin M, Kim K M, Sud Y C and Walker G K 2006 Atmospheric teleconnection over Eurasia induced by aerosol radiative forcing during boreal spring; *J. Climate* **19** 4700–4718.
- Kneizys F X, Shettle E P, Gallery W O, Chetwynd J H Jr and Abreu L W 1983 Atmospheric transmittance/radiance: Computer code LOWTRAN6; Report No. AFGL-TR-83-0187, Air Force Geophys. Lab. Hanscom AFB, MA, 200p.
- Koukoulis M E, Balis D S, Loyola D, Valks P, Zimmer W, Hao N, Lambert J-C, Van Roozendael M, Lerot C and

- Spurr 2012 Geophysical validation and long-term consistency between GOME-2/MetOp-A total T2169–2181, doi: 10.5194/amt-5-2169-2012.
- Kubilay N, Cokacar T and Temel O 2003 Optical properties of mineral dust outbreaks over the northeastern Mediterranean; *J. Geophys. Res.* **108** 4666.
- Lee K H, Li Z, Wong M S, Xin J, Wang Y, Hao W M and Zhao F 2007 Aerosol single-scattering albedo estimated across China from a combination of ground and satellite measurements; *J. Geophys. Res.* **112** D22S15, doi: 10.1029/2007JD009077.
- Lelieveld J, Hoor P, Jöckel P, Pozzer A, Hadjinicolaou P, Cammas J P and Beirle S 2009 Severe ozone air pollution in the Persian Gulf region; *Atmos. Chem. Phys.* **9** 1393–1406.
- Levelt P F, van den Oord G H I, Dobber M R, Malkki A, Visser H, de Vries J, Stammes P, Lundell J O V and Saari H 2006 The ozone monitoring instrument; *IEEE T. Geosci. Remote* **44** 1093–1101.
- Li F, Vogelmann A M and Ramanathan V 2003 Saharan dust aerosol radiative forcing measured from space; *J. Climate* **17** 2558–2571.
- Liou K N 1980 *An Introduction to Atmospheric Radiation*; Academic Press, New York, 392p.
- Liou K N 2002 *An Introduction to Atmospheric Radiation*; second edn, Academic Press, San Diego, 583p.
- Lodhi N K, Beegum S N, Singh S and Kumar K 2013 Aerosol climatology at Delhi in the western Indo-Gangetic Plain: Microphysics, long-term trends, and source strengths; *J. Geophys. Res.* **118** 1–15.
- Loeb N G, Smith N M, Kato S, Miller W F, Gupta S K, Minnis P and Wielicki B A 2003 Angular distribution models for top-of-atmosphere radiative flux estimation from the Clouds and the Earth's Radiant Energy System instrument on the Tropical Rainfall Measuring Mission Satellite. Part I: Methodology; *J. Appl. Meteor.* **42** 240–265.
- Loeb N G, Kato S, Loukachine K and Smith N M 2005 Angular distribution models for top-of-atmosphere radiative flux estimation from the clouds and the Earth's Radiant Energy System instrument on the Terra satellite. Part I: Methodology; *J. Atmos. Oceanic Technol.* **22** 338–351.
- Lubin D, Satheesh S K, Macfarquar G and Heymsfield A 2002 The longwave radiative forcing of Indian Ocean tropospheric aerosol; *J. Geophys. Res.* **107** 8004, doi: 10.1029/2001JD001183.
- Mallet M, Tulet P and Serc D 2009 Impact of dust aerosols on the radiative budget, surface heat fluxes, heating rate profiles and convective activity over West Africa during March 2006; *Atmos. Chem. Phys.* **9** 7143–7160.
- Markowicz K M, Flatau P J, Remiszewska J, Witek M, Reid E A, Reid J S, Bucholtz A and Holben B 2008 Observations and modeling of the surface aerosol radiative forcing during UAE 2; *J. Atmos. Sci.* **65** 2877–2891.
- McArthur L J B 2005 World Climate Research Programme, Baseline Surface Radiation Network (BSRN); WCRP-121 WMO/TD-NO. 1274, <http://www.bsrn.awi.de/en/other/publications/>.
- McClatchey R A, Fenn R W, Selby E G A, Volz F E and Garing J S 1972 Optical properties of the atmosphere; Rep. AFCRL-72-0497, Air Force Cambridge Res. Lab., Bedford, Mass.
- Megaritis A G, Fountoukis C, Charalampidis P E, Denier van der Gon H A C, Pilinis C and Pandis S N 2014 Linking climate and air quality over Europe: Effects of meteorology on PM_{2.5} concentration; *Atmos. Chem. Phys.* **14**(18) 10,283–10,298, doi: 10.5194/acp-14-10283-2014.
- Meywerk J and Ramanathan V 1999 Observations of the spectral clear-sky aerosol forcing over the tropical Indian Ocean; *J. Geophys. Res.* **104** 24,359–24,370.
- Miller R L and Tegen I 1998 Climate response to soil dust aerosols; *J. Climate* **11** 3247–3267.
- Moorthy K K, Babu S S and Satheesh S K 2005 Aerosol characteristics and radiative impacts over the Arabian Sea during the intermonsoon season results from ARMEX field campaign; *J. Atmos. Sci.* **62** 192–206.
- Moorthy K K, Babu S S and Satheesh S K 2007 Temporal heterogeneity in aerosol characteristics and the resulting radiative impact at a tropical coastal station – Part 1: Microphysical and optical properties; *Ann. Geophys.* **25** 2293–2308.
- Moorthy K K, Nair V S, Babu S S and Satheesh S K 2009 Spatial and vertical heterogeneities in aerosol properties over oceanic regions around India: implications for radiative forcing; *Quart. J. Roy. Meteorol. Soc.* **135** 2131–2145.
- Naddafi K, Hassanvand M S, Yunesian M and Momeniha F 2012 Environmentl health impact assessment of air pollution in megacity of Tehran, Iran; *J. Environ. Health Sci. Eng.* **9** 1–7.
- Nakajima T, Hayasaka T, Higurashi A, Hashida G, Moharram-Nejab N, Najafi Y and Valavi H 1996 Aerosol optical properties in the Iranian region obtained by ground-based solar radiation measurements in the summer of 1991; *J. Appl. Meteorol.* **35** 1265–1278.
- O'Neill N T, Dubovic O and Eck T F 2001 Modified Angstrom exponent for the characterization of submicrometer aerosols; *Appl. Optics* **40**(15) 2368–2375.
- O'Neill N T, Eck T F, Smirnov A, Holben B N and Thulasiraman S 2003 Spectral discrimination of coarse and fine mode optical depth; *J. Geophys. Res.* **108** 4559.
- Oliver J E and Hidore J J 1984 Climatology: An introduction; *Nature* 381p.
- Ouarda T B M J, Charron C, Niranjana Kumar K, Marpu P R, Ghedira H, Molini A and Khayal I 2014 Evolution of the rainfall regime in the United Arab Emirates; *J. Hydrol.* **514** 258–270.
- Pandithurai G, Dipu S, Dani K K, Tiwari S, Bisht D S, Devara P C S and Pinker R T 2008 Aerosol radiative forcing during dust events over New Delhi, India; *J. Geophys. Res.* **113** D13209.
- Parajuli S, Gherboudj I and Ghedira H 2013 The effect of soil moisture and wind speed on aerosol optical thickness retrieval in a desert environment using SEVIRI thermal channels; *Int. J. Remote Sens.* **34**(14) 5054–5071.
- Pérez-Ramírez D, Whiteman D N, Smirnov A, Lyamani H, Holben B N, Pinker R, Andrade M and Alados-Arboledas L 2014 Evaluation of AERONET precipitable water vapour versus micro-wave radiometry, GPS, and radiosondes at ARM sites; *J. Geophys. Res. Atmos.* **119** 9596–9613, doi: 10.1002/2014JD021730.
- Prasad A K, Singh S, Chauhan S S, Srivastava M K, Singh R P and Singh R 2007 Aerosol radiative forcing over the Indo-Gangetic plains during major dust storms; *Atmos. Environ.* **41**(29) 6289–6301.
- Prospero J M, Charlson R J, Mohnen V, Jaenicke R, Delany A C, Moyers J, Zoller W and Rahn K 1983 The atmospheric aerosol system: An overview; *Rev. Geophys.* **21** 1607, doi: 10.1029/RG021i007p01607.
- Prospero J M, Ginoux P, Torres O and Nicholson S E 2002 Environmental characterization of global sources of atmospheric soil dust identified with the NIMBUS 7 total ozone mapping spectrometer (TOMS) absorbing aerosol product; *Rev. Geophys.* **40** 1–31.
- Rajeev K and Ramanathan V 2001 Direct observations of clear-sky aerosol radiative forcing from space during the Indian Ocean Experiment; *J. Geophys. Res.* **106**(D15) 17,221–17,235, doi: 10.1029/2000JD900723.

- Ramanathan V *et al.* 2001 Indian Ocean Experiment: An integrated analysis of the climate forcing and effects of the great Indo-Asian haze; *J. Geophys. Res.* **106** 28,371–28,398.
- Ramanathan V, Li F, Ramana M V, Praveen P S, Kim D, Corrigan C E, Nguyen H, Stone Schauer J J, Carmichael G R, Adhikary B and Yoon S C 2007 Atmospheric brown clouds: Hemispherical and regional variations in long-range transport, absorption, and radiative forcing; *J. Geophys. Res.* **112** D22S21.
- Ricchiazzi P, Yang S, Gautier C and Sowle D 1988 SBDART: A research and teaching software tool for plane-parallel radiative transfer in the earth's atmosphere; *B. Am. Meteorol. Soc.* **79** 2101–2114.
- Rodríguez S, Alastuey A, Alonso-Pérez S, Querol X, Cuevas E, Abreu-Afonso J, Viana M, Pandolfi M and de la Rosa J 2011 Transport of desert dust mixed with North African industrial pollutants in the subtropical Saharan Air Layer; *Atmos. Chem. Phys. Discuss.* **11** 8841–8892.
- Sabbah I and Hasan F M 2008 Remote sensing of aerosols over the Solar Village, Saudi Arabia; *Atmos. Res.* **90** 170–179.
- Saha A, Mallet M, Roger J C, Dubuisson P, Piazzola J and Despiou S 2008 One year measurements of aerosol optical properties over an urban coastal site: Effect on local direct radiative forcing; *Atmos. Res.* **90** 195–202, doi: 10.1016/j.atmosres.2008.02.003.
- Satheesh S K 2002 Radiative forcing by aerosols over Bay of Bengal region; *Geophys. Res. Lett.* **29**(22) 2083, doi: 10.1029/2002GL015334.
- Satheesh S K and Ramanathan V 2000 Large differences in tropical aerosol forcing at the top of the atmosphere and Earth's surface; *Nature* **405** 60–63.
- Satheesh S K and Krishnamoorthy K 2005 Radiative effects of natural aerosols: A review; *Atmos. Environ.* **39** 2089–2110.
- Satheesh S K, Ramanathan V, Li-Jones X, Lobert J M, Podgorny I A, Prospero J M, Holben B N and Loeb N G 1999 A model for the natural and anthropogenic aerosols over the tropical Indian Ocean derived from Indian Ocean experiment data; *J. Geophys. Res.* **104** 27,421–27,440.
- Satheesh S K, Deepshikha S and Srinivasan J 2006 Impact of dust aerosols on earth-atmosphere clear-sky albedo and its short wave radiative forcing over African and Arabian regions; *Int. J. Remote Sens.* **27** 1691–1706.
- Satheesh S K, Dutt C B S, Srinivasan J and Rao U R 2007 Atmospheric warming due to dust absorption over Afro-Asian regions; *Geophys. Res. Lett.* **34**(4) L04805, doi: 10.1029/2006GL02862.
- Schuster G L, Dubovik O and Holben B N 2006 Angstrom exponent and bimodal aerosol size distributions; *J. Geophys. Res.* **111** D07207, doi: 10.1029/2005JD006328.
- Shettle E P, Kneizys F X and Gallery W O 1980 Suggest modification to the total volume molecular scattering coefficient in LOWTRAN; *Appl. Opt.* **19** 2873–2874.
- Shao Y 2009 *Physics and Modelling of Wind Erosion*; Kluwer Academic Publishers, 105p.
- Shihab M G 2001 Economic development in the UAE; In: United Arab Emirates: A new perspective (eds) Hellyer P and Al Abed I, Trident Press, London.
- Singh S, Soni K, Bano T, Tanwar R S, Nath S and Arya B C 2010 Clear-sky direct aerosol radiative forcing variations over mega-city Delhi; *Ann. Geophys.* **28** 1157–1166.
- Smirnov A V, Villevalde Y, O'Neill N T, Royer A and Tarussov A 1995 Aerosol optical depth over the oceans: Analysis in terms of synoptic air mass types; *J. Geophys. Res.* **100** 16,639–16,650.
- Smirnov A, Holben B N and Eck T F 2000 Cloud-screening and quality control algorithms for the AERONET database; *Remote Sens. Environ.* **73**. doi: 10.1016/S0034-4257(00)00109-7.
- Sokolik I N and Toon O B 1999 Incorporation of mineralogical composition into models of the radiative properties of mineral aerosol from UV to IR wavelengths; *J. Geophys. Res.* **104** 9423.
- Spurr R J D 2004 A new approach to the retrieval of surface properties from earthshine measurements; *J. Quant. Spectrosc. Radiat. Transfer* **83**(1) 15–46.
- Spurr R 2006 VLIDORT: A linearized pseudo-spherical vector discrete ordinate radiative transfer code for forward model and retrieval studies in multilayer multiple scattering media; *J. Quant. Spectrosc. Radiat. Transfer* **102** 316–342.
- Spurr R, Wang J, Zeng J and Mishchenko M J 2012 Linearized T-matrix and Mie scattering computations; *J. Quant. Spectrosc. Radiat. Transfer* **113**(6) 425–439.
- Srivastava A K, Tiwari S, Devara P C S, Bisht D S, Srivastava M K, Tripathi S N and Goloub P 2011 Annales geophysicae pre-monsoon aerosol characteristics over the Indo-Gangetic Basin: Implications to climatic impact; *Ann. Geophys.* **29** 789–804.
- Stamnes K, Tsay S C, Wiscombe W and Jayaweera K 1988 Numerically stable algorithm for discrete-ordinate-method radiative transfer in multiple scattering and emitting layered media; *Appl. Optics* **27**(12) 2502–2509, doi: 10.1364/AO.27.002502.
- Suarez M J 2005 Technical report series on global modeling and data assimilation; Documentation and validation of the Goddard Earth Observing System (GEOS) data assimilation system - Version 4, NASA/TM-2005-104606, Washington, DC, Vol. **26**, 181 p.
- Tanre' D, Remer L A, Kaufman Y J, Mattoo S, Hobbs P V, Livingston J M, Russell P B and Smirnov A 1999 Retrieval of aerosol optical thickness and size distribution over ocean from the MODIS Airborne Simulator during TARFOX; *J. Geophys. Res.* **104** 2261–2278.
- Textor C, Schulz M, Guibert S, Kinne S and Balkanski Y *et al.* 2006 Analysis and quantification of the diversities of aerosol life cycles within AeroCom; *Atmos. Chem. Phys.* **6**(7) 1777–1813, doi: 10.5194/acp-6-1777-2006.
- Thomas G E, Chalmers N, Harris B, Grainger R G and Highwood E J 2013 Regional and monthly and clear-sky aerosol direct radiative effect (and forcing) derived from the Glob AEROSOL-AATSR satellite aerosol product; *Atmos. Chem. Phys.* **13** 393–410, doi: 10.5194/acp-13-393-2013.
- Toledano C, Cachorro V E, Sorribas M, Berjon A, de la Morena B A, de Frutos A M and Gouloub P 2007 Aerosol optical depth and Angstrom exponent climatology at El Arenosillo AERONET site (Huelva, Spain); *Quart. J. Roy. Meteor. Soc.* **133**, doi: 10.1002/qj.54.
- Torres O, Bhartia P K, Herman J R, Ahmad Z and Gleason J 1998 Derivation of aerosol properties from satellite measurements of backscattered ultraviolet radiation: theoretical basis; *J. Geophys. Res.* **103** 17,099–17,110.
- Torres O, Tanskanen A, Veihelmann B, Ahn C, Braak R, Bhartia P K and Levelt P 2007 Aerosols and surface UV products from ozone monitoring instrument observations: an overview; *J. Geophys. Res. Atmos.* **112** 1984–2012.
- Torres O, Ahn C and Chen Z 2013 Improvements to the OMI near-UV aerosol algorithm using A-train CALIOP and AIRS observations; *Atmos. Meas. Tech.* **6** 3257–3270, doi: 10.5194/amt-6-3257-2013.
- Valenzuela A, Olmo F J, Lyamani H, Antón M, Quirantes A and Alados-Arboledas L 2012 Aerosol radiative forcing

- during African desert dust events (2005–2010) over South-eastern Spain; *Atmos. Chem. Phys.* **12** 10,331–10,351.
- Veefkind J P, de Haan J F, Brinksma E J, Kroon M and Levelt P F 2006 Total ozone from the ozone monitoring instrument (OMI) using the DOAS technique; *IEEE Trans. Geosci. Remote Sens.* **44**(5) 1239–1244, doi: 10.1109/TGRS.2006.871204.
- Vinoj V, Satheesh S K and Moorthy K K 2010 Optical, radiative, and source characteristics of aerosols at Minicoy, a remote island in the southern Arabian Sea; *J. Geophys. Res.* **115** D01201, doi: 10.1029/2009JD011810.
- Wang Y, Xin J, Li Z, Wang S, Wang P, Hao W, Nordgren B L, Chen H, Wang L and Sun Y 2008 Seasonal variations in aerosol optical properties over China; *Atmos. Chem. Phys. Discuss.* **8** 8431–8453.
- Waquet F, Cairns B, Knobelspiesse K, Chowdhary J, Travis L D, Schmid B and Mishchenko M I 2009 Polarimetric remote sensing of aerosols over land; *J. Geophys. Res.* **114** D01206, doi: 10.1029/2008JD010619.
- Wielicki B A, Barkstrom B R, Harrison E F, Lee R B, Smith G L and Cooper J E 1996 Clouds and the Earth's Radiant Energy System (CERES): An Earth Observing System Experiment; *Bull. Am. Meteor. Soc.* **77** 853–868.
- Xu X and Wang J 2015 Retrieval of aerosol microphysical properties from AERONET photo-polarimetric measurements: Information content analysis; *J. Geophys. Res. Atmos.* **120**, doi: 10.1002/2015JD023108.
- Zhao C, Liu X, Leung L R, Johnson B, McFarlane S A, Gustafson W I Jr, Fast J D and Easter R 2010 The spatial distribution of mineral dust and its short-wave radiative forcing over North Africa: Modeling sensitivities to dust emissions and aerosol size treatments; *Atmos. Chem. Phys.* **10** 8821–8838.

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