

Observations and Modeling of the Surface Aerosol Radiative Forcing during UAE²

K. M. MARKOWICZ,* P. J. FLATAU,⁺ J. REMISZEWSKA,[#] M. WITEK,* E. A. REID,[@] J. S. REID,[@]
A. BUCHOLTZ,[@] AND B. HOLBEN[&]

**Institute of Geophysics, University of Warsaw, Warsaw, Poland*

⁺Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California

[#]Institute of Geophysics, Polish Academy of Sciences, Warsaw, Poland

[@]Marine Meteorology Division, Naval Research Laboratory, Monterey, California

[&]Biospheric Sciences Branch, NASA Goddard Space Flight Center, Greenbelt, Maryland

(Manuscript received 28 June 2007, in final form 27 December 2007)

ABSTRACT

Aerosol radiative forcing in the Persian Gulf region is derived from data collected during the United Arab Emirates (UAE) Unified Aerosol Experiment (UAE²). This campaign took place in August and September of 2004. The land–sea–breeze circulation modulates the diurnal variability of the aerosol properties and aerosol radiative forcing at the surface. Larger aerosol radiative forcing is observed during the land breeze in comparison to the sea breeze. The aerosol optical properties change as the onshore wind brings slightly cleaner air. The mean diurnal value of the surface aerosol forcing during the UAE² campaign is about -20 W m^{-2} , which corresponds to large aerosol optical thickness (0.45 at 500 nm). The aerosol forcing efficiency [i.e., broadband shortwave forcing per unit optical depth at 550 nm, $\text{W m}^{-2} (\tau_{500})^{-1}$] is $-53 \text{ W m}^{-2} (\tau_{500})^{-1}$ and the average single scattering albedo is 0.93 at 550 nm.

1. Introduction

The United Arab Emirates (UAE) Unified Aerosol Experiment (UAE²) was conducted in August and September 2004, and included two radiation and aerosol surface supersites and 15 Aerosol Robotic Network (AERONET) sites with sun photometers deployed in the coastal and desert regions of the UAE. One of the two ground observational stations, the Naval Research Laboratory (NRL) Mobile Atmospheric Aerosol and Radiation Characterization Observatory (MAARCO) was located on the Persian Gulf coast approximately 50 km northeast of Abu Dhabi, United Arab Emirates, and away from the primary city plume. This location allowed for studies of the aerosol properties over land in the proximity of both the desert and the Persian Gulf. The strong sea–land–breeze circulation is a regular phenomenon in the coastal zone of the Persian Gulf in the absence of strong large-scale flow (Zhu and Atkinson 2004). During the sea breeze strong winds often

lead to local mineral dust production. Low humidity and middle-troposphere subsidence are responsible for the mostly clear-sky conditions; thus, it was a good site to study the direct aerosol radiative forcing.

The purpose of this study is to extend previous knowledge about the aerosol optical properties in the Middle East region. We investigate the influence of aerosol on the surface radiation budget. The results are based on the data collected during 6 weeks of measurements performed at the MAARCO site, including surface and columnar observations. We discuss aerosol forcing estimated by independent methods, including surface observations of the solar fluxes and radiative transfer model calculations.

The role of atmospheric aerosols in modifying the radiation budget of the earth–atmosphere climate system is being increasingly recognized (Hansen et al. 1997; Haywood et al. 1999; Ramanathan et al. 2001a). However, there are still large uncertainties of the aerosol radiative forcing on the regional scale (Houghton et al. 2001) resulting from a lack of sufficient knowledge of the aerosols' optical, physical, and chemical properties, and large spatial and temporal variability. Much of the recent work has been devoted to reducing the aerosol forcing uncertainties by improving global circulation

Corresponding author address: K. M. Markowicz, Institute of Geophysics, University of Warsaw, Pasteura 7, Warsaw 02093, Poland.

E-mail: kmark@uninet.com.pl

models (Chin et al. 2002; Takemura et al. 2002) and transport models (Collins et al. 2001). Also, establishment of the observational networks, such as the AERONET (Holben et al. 2001), the European Aerosol Research Lidar Network (EARLINET), and the Micropulse Lidar Network (MPLNET), dedicated to monitoring the aerosol properties and their vertical distribution (Wielicki et al. 1996), resulted in fast progress in this field. Another important component of the aerosol forcing research are the observational campaigns involved—smoke, clouds and radiation—Brazil (SCAR-B; Kaufman et al. 1998), Tropospheric Aerosol Radiative Forcing Observational Experiment (TARFOX; Hobbs 1999; Redemann et al. 2000), First Aerosol Characterization Experiment (ACE1; Bates et al. 1998), Second ACE (ACE2; Raes et al. 2000), INDOEX (Ramanathan et al. 2001b), Mediterranean Intensive Oxidant Study (MINOS; Lelieveld et al. 2003), and Saharan Dust Experiment (SHADE; Tanré et al. 2003).

Based on the Total Ozone Mapping Spectrometer aerosol index (Goudie and Middleton 2001), the region of the Persian Gulf has recently been identified as the world's third largest dust source (Léon and Legrand 2003). Maximum dust activity occurs during the pre-monsoonal (spring) and monsoonal (summer) periods (Ackerman and Cox 1989). During the summer active dust areas also are located along the Persian Gulf from Kuwait to the UAE. Large anthropogenic emissions (Langner et al. 1992) and the interaction of pollution with mineral dust can lead to complex aerosol particles. Pollution is emitted mostly by local refineries, factories, and fossil fuel combustion, in addition to being transported from the Indian subcontinent. The aerosol optical properties in the Persian Gulf region are not well characterized. The ground and airborne measurements made in the spring and summer of 1991 (Draxler et al. 1994; Hobbs and Radke 1992; Nakajima et al. 1996) were focused on the Kuwaiti oil fires of 1991.

Recently, one site in the Persian Gulf region (Bahrain) was characterized in terms of the background aerosol properties (Smirnov et al. 2002) using a scanning radiometer. The results indicate significant diurnal and annual variability of the columnar aerosol optical properties and aerosol size distribution. The annual cycle mostly is due to seasonal dust production and, therefore, aerosol optical thickness (AOT) is negatively correlated with the Ångström exponent. The maximum of AOT is observed during the summer and reaches 0.45 at 500 nm. Based on the Meteorological Satellite (Meteosat), observations in both the UV-visible and the infrared spectrum (Deepshikha et al. 2005) derive the dust-absorbing efficiency over the north Indian Ocean.

Deepshikha et al. (2005) noticed that dust originating from the desert areas of the Arabian Peninsula is less absorbent than dust interacting with the anthropogenic aerosols; this provided strong evidence for occasional mixing of dust particles with black carbon.

To the best of our knowledge there is no extensive body of literature related to the aerosol influence on the radiation budget in the Persian Gulf. However, there were several intensive field studies related to the aerosol radiative forcing over the Indian Ocean and Arabian Sea. One paper, based on 5 yr of satellite data analysis (Ramachandran 2005), finds a fairly large annual reduction of the solar radiation (-22 W m^{-2}) at the earth's surface over the Arabian Sea. The annual mean radiative forcing at the top of the atmosphere (TOA) was found to be -9 W m^{-2} , and large atmospheric absorption of about 13 W m^{-2} indicates the significant influence of soot emitted from fossil fuel and biomass burning. These results are influenced by the strong emission from the Indian subcontinent and cannot be directly extended to the Persian Gulf region. Satheesh et al. (2006), based on AERONET and Moderate Resolution Imaging Spectroradiometer (MODIS) data, found that the radiative forcing at the TOA over Saudi Arabia is in the range from $+2$ to $+4 \text{ W m}^{-2}$ (annual mean). While TOA forcing is nearly zero because of the large surface albedo, the surface aerosol forcing is negative and varies in the range from -30 to -70 W m^{-2} . However, the calculated aerosol radiative effect by Derimian et al. (2006) over Negev Desert in Israel shows cooling both at the top of the atmosphere and at the surface during the entire year.

Model studies (e.g., The Florida State University Limited Area Model) of mesoscale dynamics sensitivity to the radiative transfer parameterization were performed (Mohalfi et al. 1998) with the two-stream radiative transfer approximation with assumed single scattering dust properties. The results indicated improvements in the modeling of the diurnal variability of the air temperature when dust was included in the simulations. The dust layer weakens the sea breeze, but does not have any significant effect on the land breeze. However, the assumed single scattering albedo ($\text{SSA} = 0.8$) in this paper was probably too low.

2. Description of the observational site

During UAE² the aerosol optical, physical, and chemical properties as well as meteorological parameters were collected at the NRL MAARCO. MAARCO is an air-conditioned shipping container that has been modified to function as a portable labo-

TABLE 1. List of instrumentation and derived quantities.

Instruments	Quantity	Accuracy
Pyranometer CM21	Total shortwave flux	$\pm 9 \text{ W m}^{-2}$ and $\pm 10 \text{ W m}^{-2}$ for noncosine response ^a
Shaded pyranometer CM21	Diffuse shortwave flux	$\pm 9 \text{ W m}^{-2a}$
Pyrheliometer CH1	Direct shortwave flux	$\pm 3 \text{ W m}^{-2a}$
Cimel (AERONET)	Aerosol optical thickness	0.01–0.02 (Schmid et al. 1999)
	Single scattering albedo	0.03 (Dubovik and King 2000)
	Asymmetry parameter	3%–5% (Dubovik et al. 2006)
	Precipitable water	10% (Schmid et al. 2001)
	Scattering coefficient	15% (Anderson et al. 1996)
Nephelometer TSI	Absorption coefficient	8% (Remiszewska et al. 2006)
Aethalometer AE-30	Range-corrected signal	2% ^b
Ceilmeter CT25K	Aerosol extinction coefficient	15%–20% (Schmid et al. 2006; Welton and Campbell 2002)
MPL	Precipitable water	5% ^c
Radiosondes		

^a Technical information provided by Kipp & Zonen (2001, 2004).

^b Technical information provided by the Vaisala.

^c Estimate based on Vaisala technical information about accuracy of the relative humidity, temperature, and pressure measurement for the RS92 radiosonde.

ratory. The observational station was located in the northeastern part of the UAE at 24.700°N, 54.659°E, about 10 m from the seashore. The site is close to the local port with occasional small ships sailing back and forth to the Persian Gulf. There was some building construction in the surrounding area. The absence of high buildings made MAARCO a good location to investigate radiative fluxes. The coastline in the proximity of our site was directed from the southwest to the northeast.

3. Description of the instrumentation

The total and diffuse broadband radiative (280–2800 nm) fluxes were obtained using the CM22 Kipp & Zonen pyranometers (Table 1). The direct flux was measured by the CH1 Kipp & Zonen pyrliometer mounted on the two-axis sun tracker. All radiometers were calibrated before and after the campaign at the Kipp & Zonen calibration facility. The calibration factor for each instrument was determined by an indoor side-by-side comparison with a reference radiometer of a similar type under a stable laboratory calibration lamp. This follows the indoor radiometer calibration procedure guidelines of ISO 9847, appendix 3 (Kipp & Zonen 2004). The reference pyranometer and pyrliometer used in the comparison were calibrated at the World Radiation Centre (WRC) at Davos, Switzerland. The calibration procedure for each type of radiometer is more fully explained in the instruction manuals for the pyranometer (Kipp & Zonen 2004) and the pyrliometer (Kipp & Zonen 2001). The postcampaign calibration factors were found to be within 1.7% of the precampaign values. In addition, the zero offset correc-

tion was performed for pyranometers and pyrliometers.

The total aerosol optical depth was measured with a Cimel instrument (Holben et al. 1998). At the MAARCO site two Cimel sunphotometers (303 and 328) were installed. These instruments measure direct and diffuse solar radiation at eight spectral bands (340, 380, 440, 500, 675, 870, 936, and 1020 nm). AERONET software processed the data from these instruments. In addition, one of the instruments had a polarizer to measure polarization of the diffuse radiance at 870 nm. The AOT and the retrieved parameters, such as the SSA, the asymmetry parameter (Bohren and Huffman 1983), and the total water vapor content (Halthore et al. 1997; Bruegge et al. 1992), are analyzed herein.

The Vaisala CT25K laser ceilometer was operating during the experiment, but its sensitivity limits retrievals to the lower troposphere (in this study only limited CT25K data are presented).

The vertical profiles of the aerosol extinction coefficient at 532 nm were also measured by a micropulse lidar (MPL; Welton and Campbell 2000). The vertical resolution of this instrument is 75 m. The aerosol extinction coefficient was obtained from the calibrated lidar signal and the Cimel observations of the AOT (in this study only limited MPL data are presented).

The aerosol absorption coefficient at the surface was obtained from the AE-30 aethalometer produced by the Magee Scientific Company (Allen et al. 1999; Hansen et al. 1996). The measurements were performed at seven wavelengths (370, 470, 520, 590, 660, 880, and 950 nm) and were corrected for the scattering error (Remiszewska et al. 2007), the deposit spot size, the AE-30 flow rate, and the manufacturer's calibra-

tion. Measurements of the aerosol scattering and the hemispheric backscattering coefficients were made with an integrating nephelometer (Model 3563, TSI, Inc.; Anderson et al. 1996) at 450, 550, and 700 nm. During the experiment two nephelometers were simultaneously operated—one at near-ambient conditions and the other one at a constant relative humidity of about 35%.

Radio soundings from the Abu Dhabi International Airport (about 30 km from MAARCO) were used, and several soundings were performed at the MAARCO site. A weather station (ET106) from Campbell Scientific provided local weather conditions, including air temperature, relative humidity, wind speed, and wind direction. In addition, the cloud cover was estimated using the Yankee Whole Sky Camera.

To minimize instruments uncertainties we performed several calibrations during and after UAE². In the case of pyranometers and pyrhelometer we performed the zero (thermal) offset calibration. This offset is mostly caused by the disturbance of the thermal equilibrium within the instruments. Longwave emission of the pyranometer glass domes is the major source of zero offset for this instrument under stable temperature conditions. Zero offset depends on the difference between sensor temperature and effective sky temperature and may be different during the day. This effect is usually small (Bush et al. 2000) and it is not taken into account in this study. We used ventilated pyranometers for which the zero offset is smaller in comparison to unventilated pyranometers. Drummond and Roche (1965) show that ventilation decreases the zero offset by using the air to maintain a more uniform temperature over the surface of the instrument. In addition, we performed intercomparisons of these instruments to check their calibrations and response differences.

Daily average downward total flux was calculated using the direct pyrhelometer and shadowed pyranometer techniques. In this method the shadow disk effect leads to some uncertainties. The direct flux is measured by the pyrhelometer (CH1) with the field of view (FOV) of 5.7°, while for the diffuse flux the shaded disk blocks 6.4° of the sun's aureole; this correction is usually less than 1%. The mean bias of the total fluxes for both methods is 4.1 W m⁻² and the rms difference is 8.9 W m⁻². The total flux obtained directly from the nonshaded pyranometer is slightly larger than that determined from the direct and the diffuse fluxes.

The total water content from Cimel observations was verified with the radio soundings. The correlation coefficient between both observations is 0.94, but the Cimel measurements give significantly larger values (7%). We recalibrated the precipitable water from Ci-

mel resulting from this comparison (Halthore et al. 1997).

During the UAE² campaign we performed several MPL (Welton et al. 2002) and CT25K ceilometer overlap calibrations. The overlap corrections were determined with instruments pointing horizontally.

4. Meteorological overview

August and September in the UAE is characterized by a shallow low, located in the southern part of the Arabian Peninsula (Fig. 1a). This low pressure system is a part of the large pressure system dominating the Indian subcontinent (Zhu and Atkinson 2004). The typical mesoscale pressure gradient around the UAE is small [<0.5 hPa (100 km)⁻¹]. There is large contrast between the land and sea temperature, which is responsible for the local circulation development in the coastal regions of the UAE. Sea-breeze flow is localized close to the shore and produces circulation with a wind speed of about 8–10 m s⁻¹ during the day. Climatologically, the flow at 1000 hPa shows stable southwest wind in the southeast part of the Arabian Peninsula, which is a part of the summer monsoon circulation. This flow is stronger in August than in September and sometimes reaches the section of gulf bordered by UAE. During this period the low-level flow carries mineral aerosol from the UAE desert close to the MAARCO site.

Above 700 hPa the circulation over the UAE is anticyclonic with northerly wind over the region (Fig. 1b). The 500-hPa flow is determined by the large anticyclone system, which is located in the eastern part of the Sahara and the northwest part of the Arabian Peninsula (Figs. 1c,d). These upper-level northeast winds bring dust particles from Iran and Afghanistan, where dust storm activity is significant during the summer season.

Thus, during the summer, the UAE is a convergence region for pollution from surrounding countries, such as Kuwait, Saudi Arabia, and India, but there is also strong dust storm activity in the region itself. During the UAE² campaign we observed several small dust storms close to the MAARCO site, which reduced visibility to several hundred meters.

The Arabian Peninsula in summer is under the influence of the subsiding branch of the Hadley circulation. This is responsible for the mostly clear-sky and extreme dry conditions throughout the atmospheric column. The main inversion level is close to 5.5 km, and this altitude is most often limiting the aerosol layer. With the addition of moisture altocumulus clouds develop at around the 700-hPa level, but this is fairly uncommon.

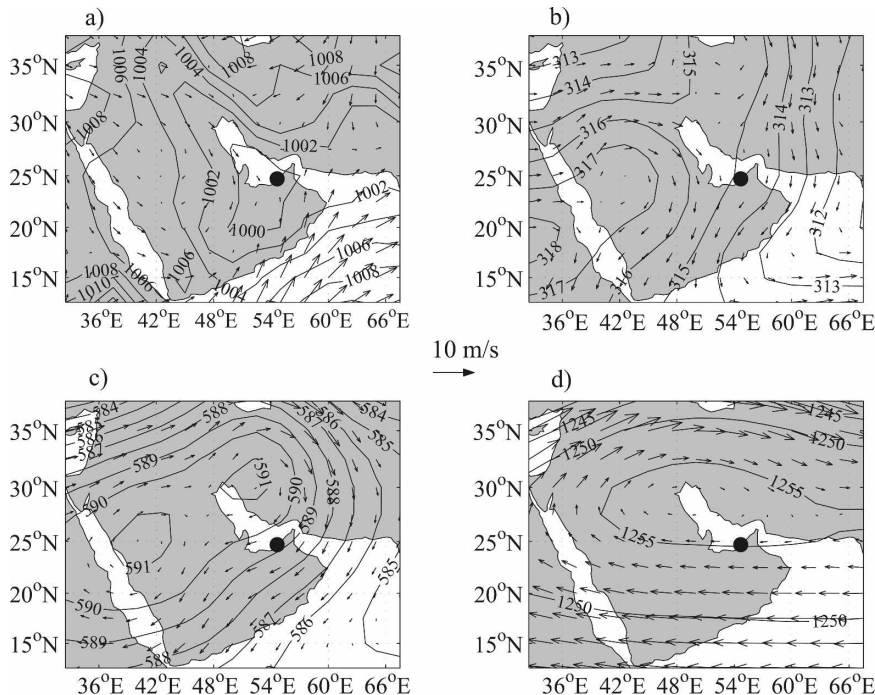


FIG. 1. Mean wind (m s^{-1}) and geopotential height (m) in August 2004 from the NCEP–NCAR reanalysis: (a) winds at 1000 hPa and the mean sea level pressure, and winds and geopotential heights at (b) 700, (c) 500, and (d) 300 hPa. Solid circle marks MAARCO's position.

5. Methodologies of the direct aerosol forcing estimations

Aerosol forcing is the perturbation of the earth–atmosphere system radiative fluxes caused by the aerosols. Direct aerosol forcing is defined as the difference between net (down minus up) radiative flux for a clear-sky atmosphere with aerosols and net clear-sky radiative flux without aerosol. The solar aerosol direct forcing at the surface A_s is defined as

$$A_s = (F_a^\downarrow - F_a^\uparrow) - (F_c^\downarrow - F_c^\uparrow), \quad (1)$$

where $F_c^\uparrow$ and $F_a^\uparrow$ are upward aerosol-free and aerosol-modified solar flux, and $F_c^\downarrow$ and $F_a^\downarrow$ are the same quantities but for downward solar flux only. We use terminology “mean” diurnal aerosol forcing to indicate the 24-h-averaged aerosol forcing.

There are several approaches for estimating aerosol radiative forcing. One is based on observations of the radiative fluxes and aerosol optical properties only. However, the observed solar flux is strongly affected by the water vapor direct absorption (Kay and Box 2000) and by the aerosol growth (Markowicz et al. 2003; Im et al. 2001). Reduction of the solar radiation by water vapor absorption is, to the first approximation, proportional to the total water vapor content. Because the

water vapor absorption is the most important source of aerosol forcing uncertainties we take this into account by deriving the aerosol forcing efficiency $F_{\text{eff}} = \Delta A_s / \Delta \text{AOT}$ (slope of the mean net diurnal flux versus the mean aerosol optical thickness at 500 nm) for days with similar water vapor content. To eliminate diurnal changes of the solar flux at the surface during several weeks of measurements we normalized the mean diurnal flux at the surface by the mean solar flux at the top of the atmosphere averaged over the time that our campaign lasted. Another way to deal with the water vapor issue is to calculate the solar flux from the radiative transfer model for the case of atmosphere without aerosols. This method requires information about the total water vapor and total ozone, as well as temperature, pressure, and humidity profiles, which we obtained from the radio soundings at Abu Dhabi International Airport (about 30 km from MAARCO). Also, we used some of the radiosondes at the MAARCO site to recalibrate the total water vapor retrieved from the Cimel sun photometer. Because of model and observational uncertainties we use slope method to calculate the aerosol forcing efficiency (Satheesh and Ramanathan 2000). In this method the mean diurnal aerosol forcing (calculated from the difference between observed and modeled aerosol-free fluxes) is plotted as a function of

TABLE 2. Errors in the surface aerosol forcing resulting from different surface composition.

Surface composition	100% land, 0% water	75% land, 25% water	50% land, 50% water (base calculations)	25% land, 75% water	0% land, 100% water
Errors in the aerosol forcing with respect to base calculations	−10.9%	−5.4%	0.0%	5.3%	10.6%

the mean diurnal aerosol optical thickness. The slope of the linear fit defines the aerosol forcing efficiency, which can be used to obtain the mean aerosol forcing for individual days. This method assumes that the aerosol optical properties are constant and that the assumption about linearity is valid (this assumption is valid for a certain range of optical thicknesses). The advantage of this method is that aerosol forcing estimated from the slope method is not influenced by observational and model bias for certain range of AOTs. However, if this bias changes with the aerosol optical thickness, then the slope method can be sensitive to observational and model offset. Information about the total water vapor and total ozone, as well as temperature, pressure, and humidity profiles, was obtained from the radio soundings at Abu Dhabi International Airport (again, about 30 km from MAARCO). Also, we used some of the radiosondes at the MAARCO site to correct the total water vapor retrieved from the Cimel sun photometer.

Even though measurements were performed over the land we calculate the aerosol forcing over the water surface. This allows us to compare our results with results from other field campaigns. To this end we computed a correction to the net solar flux resulting from the change of the surface albedo from land to sea. This transfer requires information about the albedo close to the measurement station. Because the measuring station was located just on the border of water and land we assume that the surface albedo is a combination of 50% desert and 50% water. We use observed albedo based on the MODIS product MOD43B3 (Moody et al. 2008), which includes information about spectral and broadband “white sky” and “black sky” albedo at nine wavelengths (0.47, 0.55, 0.67, 0.86, 1.24, 2.1, 0.3–0.7, 0.3–5.0, and 0.7–5.0 μm). This dataset contains spatially complete land surface albedo at 1-min resolution on an equal-angle grid. The database includes only albedo at the mean solar zenith angle close to noontime, and we use the streamer model (Key and Schweiger 1998) to compute the variability of the albedo resulting from solar elevation. The model of the water albedo assumes Fresnel reflectance, with the correction resulting from whitecaps and the mean concentration of chlorophyll at the seashore.

The MODTRAN version 4.1 (Berk et al. 1998) ra-

diative transfer code with the discrete ordinate radiative transfer (DISORT; Stamnes et al. 1998) solver was used for the radiative transfer calculations. The DISORT model includes multiple scattering effects. To obtain uncertainties of the aerosol radiative forcing δA we assumed that modeled $\delta F_{\text{modeled}}$ and measured $\delta F_{\text{measured}}$ uncertainties are uncorrelated and can be estimated from the expression

$$\delta A = \sqrt{(\delta F_{\text{modeled}})^2 + (\delta F_{\text{measured}})^2}. \quad (2)$$

MODTRAN accuracy is about 2% for the aerosol-free solar flux [cf. the Intercomparison of Radiation Codes in Climate Models (ICRCCM) of Fouquart et al. (1991) and the Halthore et al. (2005) study].

Errors associated with the surface albedo assumption

Our measuring station was located just on the border between water and land. Therefore, the assumption that the albedo is a mix of 50% land and 50% water seems to be quite reasonable. However, to estimate the maximum error associated with this assumption we performed simulations and calculated aerosol forcing assuming that the albedo composition is different with respect to the base case of 50%–50%. Table 2 presents the results of such an analysis. For example, the 75% land and 25% sea surface assumption will lead to about 5% uncertainties in the calculated aerosol forcing over water (we calculate all forcing over water).

6. Results

a. Columnar aerosol optical properties

The UAE² was conducted during the summer months when the dust activity is the strongest and the mean AOT was large—0.49 at 500 nm. The mean value of AOT for days without any clouds, which we use for the radiative forcing calculations, was 0.45 at 500 nm. In general, the AOTs were in the range of 0.12–1.2, with a standard deviation of 0.15. These high values of the AOT have a significant influence on the local energy balance.

The AOT variability during UAE² can be explained

by the following three mechanisms: (a) large-scale transport of natural and anthropogenic pollution, (b) local production of mineral dust, and (c) transport related to the land–sea circulation. The northern, low-tropospheric flow brings pollution from the Kuwaiti region. The large-scale flow, from the northeast, carries mineral dust transported in the middle troposphere from Afghanistan, Iran, and Pakistan. Southern flow originates as a clean air mass from the southern Indian Ocean but becomes polluted over the Arabian Peninsula.

Figure 2 shows that the AOT is not correlated with the Abu Dhabi International Airport and MAARCO observations (radiosondes launched at 0000 and 1200 UTC) of wind direction at 1 km AGL. The AOT measurements were matched with the radiosonde observations within the 2-h time frame. This finding holds for different heights (not shown) and can be explained by the horizontal homogenous distribution of the AOT in the Persian Gulf region. Note that the mean wind speed during UAE² measured by the radiosondes launched from the Abu Dhabi International Airport is only 4.6 m s^{-1} at 1 km and 5.2 m s^{-1} at the 2.5 km (Fig. 1), and is responsible for the weak average pollution advection and ventilation.

The Ångström exponent α is defined by the logarithmic fit of the aerosol optical thickness τ using seven wavelengths λ ($\ln \tau = -\alpha \ln \lambda + \text{const}$). This quantity shows more significant variability with the wind direction. The larger values were observed during the on-shore winds from the western and northern sectors. This is caused by the advection of the anthropogenic aerosols. The Ångström coefficient decreases during the time the offshore winds (east and south wind direction) transport mineral dust. This result is consistent with the Ångström exponent pattern observed at the surface (MAARCO), based on the nephelometer and aethalometer (Remiszewska et al. 2007). The single scattering albedo obtained from the AERONET retrievals is not significantly correlated with the wind direction (diamonds in Fig. 2).

For the onshore and offshore winds the mean single scattering albedo is almost the same (0.93). However, surface observations (Remiszewska et al. 2007) indicate a small decrease of the single scattering albedo for the offshore wind. The mean surface single scattering albedo during the land breeze is 0.91 and during the sea breeze is 0.93 at 550 nm.

Figure 3 shows, based on Cimel measurements, that there is a significant correlation between the AOT at 1020 nm and the total water vapor content. This correlation increases with the Ångström exponent. For the Ångström coefficient exceeding 0.5 the AOT is strongly

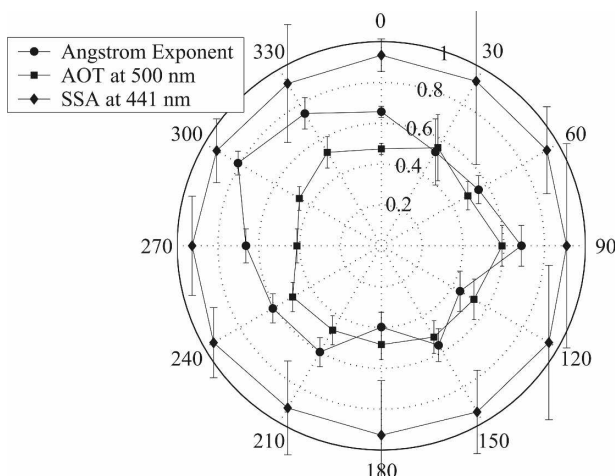


FIG. 2. The aerosol optical thickness at 500 nm (squares), the Ångström exponent (circles), and the single scattering albedo (diamonds) as a function of the wind direction at 1 km. The wind direction is based on the radiosondes launched from the Abu Dhabi International Airport at 1200 UTC.

correlated with the total water vapor content and the slope of the AOT–total water vapor linear dependence is about 0.1. For a smaller Ångström exponent (<0.5) the AOT–water vapor relationship is poor, apparently because these are large and nonhygroscopic particles.

b. Vertical distribution of aerosol extinction coefficient

The ceilometer and MPL [we used Klett's (1985) modified algorithm as described in Voss et al. (2001)] are used here to illustrate the vertical distribution of aerosols. The aerosol layer with relatively large extinction coefficient values extended from surface up to 600 m. The significant reduction of the aerosol extinction coefficient, that is, less aerosol loading, is observed above this altitude. However, some aerosols are still observed up to 5 and 6 km. This altitude is associated with a strong and steady subsidence layer. Figure 4 shows the midnight profiles of the aerosol extinction coefficient, potential temperature, and relative humidity. Weak winds and high relative humidity result in the accumulation of the pollution at the surface. Some aerosols are observed at about 5–5.5 km, and this layer is correlated with the temperature inversion and the sharp decrease of relative humidity. For an illustration of typical conditions the mean diurnal variability of the aerosol extinction coefficient obtained from the ceilometer is presented in Fig. 5. The useable vertical range of the ceilometer aerosol retrieval is only around 1.5 km. A strong surface return is observed at all times, and observations show that maximum extinction is observed in the first 500 m. However, a significant reduc-

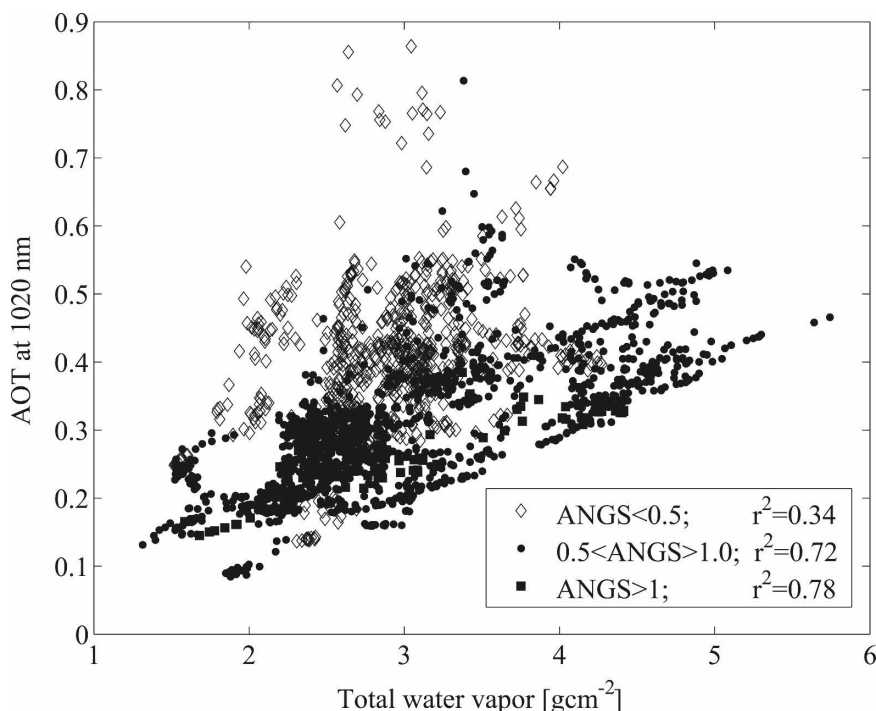


FIG. 3. Correlation between the total water vapor and the aerosol optical thickness at 1020 nm: observations with Ångström exponent <0.5 (open diamonds), between 0.5 – 1 (solid circles), and >1 (solid squares).

tion of the extinction coefficient is observed between 0700 and 1000 UTC (between 1100 and 1400 local time). The diurnal variability surface extinction is strongly correlated with the land–sea breeze. The large

est extinction coefficient is observed in the morning (0300–0600 UTC) when the wind speed and vertical mixing are minimal. After the sea-breeze onset, both the surface aerosol layer horizontal ventilation and vertical mixing (intense around local noon) lead to a more uniform distribution of the aerosols. This daily structure is discussed in more detail in Remiszewska et al. (2007).

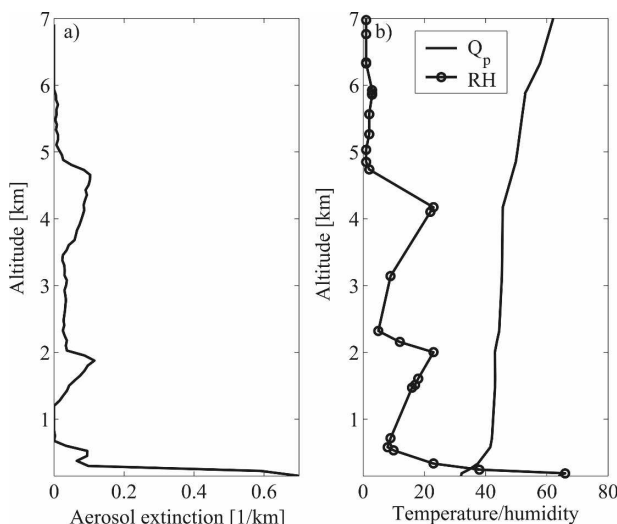


FIG. 4. Profile of (a) the extinction coefficient at 532 nm, and (b) the potential temperature (solid line) and the relative humidity (open circles and solid line) observed by the lidar and the radiosonde at 0000 UTC 7 Sep 2004.

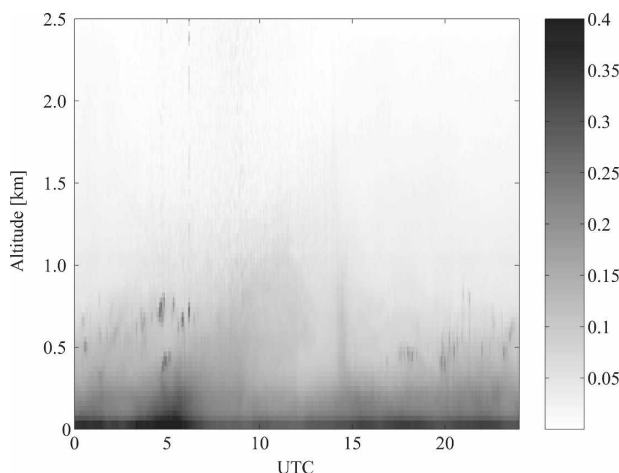


FIG. 5. Mean diurnal variability of the extinction coefficient [1 (km)^{-1}] at 905 nm as a function of the altitude based on the ceilometer observations.

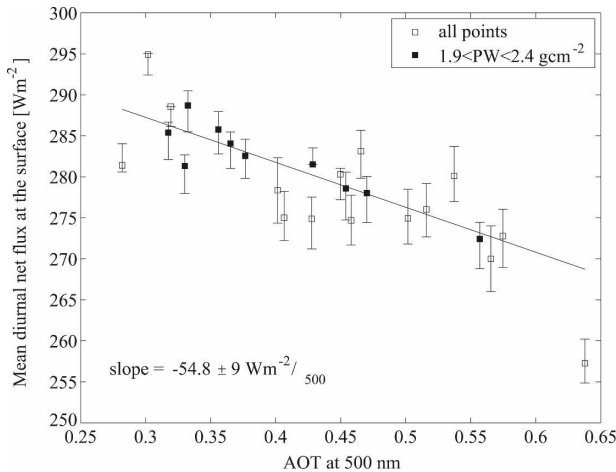


FIG. 6. Mean diurnal net solar flux at the surface corrected for mean incoming solar radiation at the top of the atmosphere as a function of the aerosol optical thickness at 500 nm (open squares). Days with the average total water vapor in the range of 1.9–2.4 g cm^{-2} (solid squares) and the linear fit to these points (solid line). The vertical lines below the squares are proportional to surface single scattering albedo at 550 nm and the vertical lines above square indicate total water vapor (in arbitrary units).

c. Aerosol forcing observations

In this study we used the 24-h mean radiative fluxes, which are based on the continuous observation of solar fluxes for days without any cloud. We choose these days based on the whole-sky camera pictures. Figure 6 shows the relationship between the mean diurnal net solar flux at the surface and the AOT at 500 nm. These fluxes were corrected for mean diurnal flux at the top of the atmosphere, which takes into account changes of the solar radiation as time progresses (change in sun declination). The range of mean diurnal incoming solar radiation at the top of the atmosphere changes between 380 and 445 W m^{-2} .

The solid squares represent cases with the total water vapor in the range from 1.9 to 2.4 g cm^{-2} , which corresponds to the 25th and 75th percentiles of the water vapor distribution as a function of water vapor content. The linear fit (slope) for these points defines the aerosol forcing efficiency. Days with extreme total water vapor content were eliminated to minimize the water vapor absorption influence on the solar radiation variability. The vertical lines above the squares on Fig. 6 are proportional to the water vapor content and the vertical lines below the squares to surface single scattering albedo. The lengths of the vertical lines are scaled to total variability of these quantities, and zero length corresponds to either the minimal value of the water vapor content or the single scattering albedo. These two parameters can explain the departure of sev-

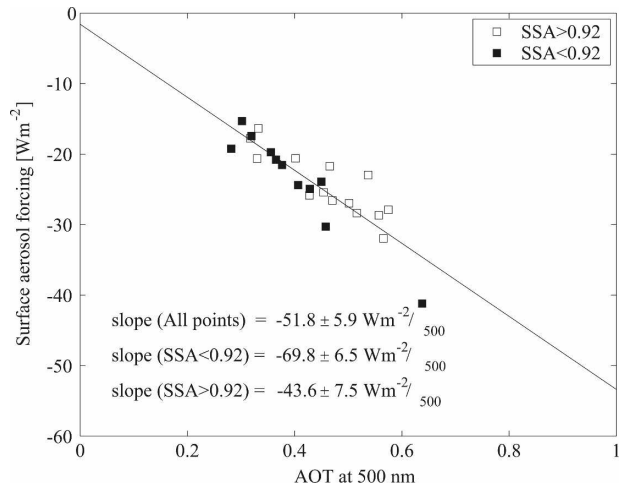


FIG. 7. The mean diurnal aerosol radiative forcing at the surface as a function of aerosol optical thickness at 500 nm: days with mean single scattering albedo at the surface larger than 0.92 (open squares) and smaller than 0.92 (solid squares). The solid line represents the linear fit to all data points.

eral points from the linear fit given by the solid line in Fig. 6. For example, a few points with small single scattering albedo and/or large total water vapor content are below the solid line because of strong absorption of solar radiation. Some data points deviate from the linear trend resulting from other causes, such as, for example, particle nonsphericity.

The slope of linear fit is $-53.7 \text{ W m}^{-2} (\tau_{500})^{-1}$ where τ_{500} is the aerosol optical depth at 500 nm. The estimated uncertainty of the aerosol forcing efficiency is large [$9.0 \text{ W m}^{-2} (\tau_{500})^{-1}$] resulting from water vapor variability. Therefore, we present the second method, which uses the modeled solar radiation fluxes for the atmosphere without aerosols but with the same vertical structure as that observed (including water vapor content). The MODTRAN version 4.1 (Berk et al. 1998) radiative transfer code with a DISORT (Stamnes et al. 1998) solver was used for the radiative transfer calculations. The DISORT model includes multiple scattering effects.

Figure 7 shows the aerosol forcing at the surface based on this observational modeling approach. The open squares represent days with mean single scattering albedo that is larger than 0.92 and solid squares represent days with mean single scattering albedo smaller than 0.92. The slope of the linear fit to all data points is slightly smaller compared to the previous method, and it is $-52.9 \pm 5.4 \text{ W m}^{-2} (\tau_{500})^{-1}$. The slope for days with single scattering albedo lower than 0.92 (at 550 nm) is $-68.8 \pm 6.2 \text{ W m}^{-2} (\tau_{500})^{-1}$ and for days with single scattering albedo larger than 0.92 (at 550 nm) is only $-45.6 \pm 6.9 \text{ W m}^{-2} (\tau_{500})^{-1}$. The increase of the aerosol forcing efficiency for aerosol with a larger

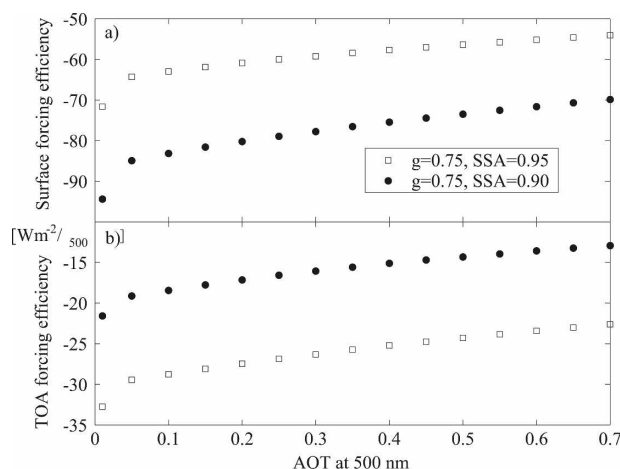


FIG. 8. Modeled mean diurnal aerosol forcing efficiency [$\text{W m}^{-2} (\tau_{500})^{-1}$] at the (a) surface and (b) TOA. An aerosol model with single scattering albedo of 0.95 and asymmetry parameter of 0.75 (dotted circles) and single scattering albedo 0.9 and the same asymmetry parameter (squares).

ratio of the absorption to extinction coefficient is expected. The mean aerosol forcing efficiency during the UAE² campaign is somewhat lower in comparison to other experiments (Haywood et al. 2003; Markowicz et al. 2003; Satheesh and Ramanathan 2000).

To better understand experimental results radiative transfer simulations were performed with vertically uniform aerosol optical properties. Figure 8 presents results of these computations for two different aerosol models with the same asymmetry parameter (0.75). The first model describes aerosol with SSA equal 0.95 (dotted circles) and the second is with SSA equal 0.9 (open squares). The aerosol radiative forcing efficiency at the surface (Fig. 8a) and the TOA (Fig. 8b) decreases significantly with the AOT. Thus, the aerosol radiative forcing is a nonlinear function of the AOT. Therefore, in the region where AOT is small the aerosol forcing efficiency is larger even when the aerosol optical properties are the same. In addition, the solar energy (diurnal mean) at the top of the atmosphere influences both the aerosol radiative forcing and aerosol radiative efficiency. Note that the increase of the AOT from 0.2 to 0.5 in our aerosol model leads to a decrease of the aerosol forcing efficiency by about $5\text{--}10 \text{ W m}^{-2} (\tau_{500})^{-1}$ at the surface and about $3 \text{ W m}^{-2} (\tau_{500})^{-1}$ at the TOA. These values depend on the aerosol optical model and the surface albedo.

d. A simple model of aerosol direct radiative forcing nonlinearity

In this section we discuss the nonlinear relationship between the aerosol radiative forcing and AOT. Al-

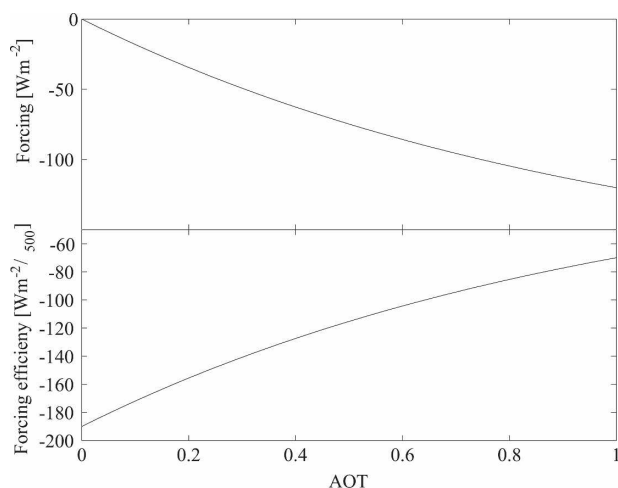


FIG. 9. (a) Instantaneous surface aerosol forcing and (b) aerosol forcing efficiency as a function of the aerosol optical thickness. These quantities are based on a conceptual model, which assumes a homogeneous aerosol layer with a single scattering albedo of 0.9, a backscatter ratio of 0.1, and a completely absorbing surface.

though a precise determination of the radiative transfer requires solving the radiation transfer equation, satisfactory results can be obtained by means of a simple model.

In this model an analytical equation explains why the increase of mean AOT leads to a reduction of the aerosol radiative forcing efficiency. To this end, we will consider a one-layer aerosol model (Seinfeld and Pandis 1998; Charlson et al. 1991) and assume the following: that the solar radiation interacts with aerosol particles and the sun is directly overhead, that the molecular scattering is assumed to be negligible, and that the surface albedo is equal to zero. The total incident radiation transmitted downward through the atmosphere layer is

$$t = e^{-\tau} + (1 - e^{-\tau})\omega(1 - b), \quad (3)$$

where b is the hemispheric backscatter ratio and ω is the single scattering albedo. The first term of this equation shows the direct part of radiation transmitted by the aerosol layer and the second term is the diffuse part. If we assume that I_0 is the solar flux at the top of the atmosphere, the aerosol radiative forcing at the surface [see Eq. (1)] can be written as

$$A_s = -I_0(1 - e^{-\tau})[1 - \omega(1 - b)]. \quad (4)$$

The aerosol forcing efficiency F_{eff} is defined as

$$F_{\text{eff}} = -I_0 e^{-\tau} [1 - \omega(1 - b)]. \quad (5)$$

Figure 9 shows that the increase of the AOT leads to a reduction of the aerosol forcing efficiency. Notice that for small optical depths the aerosol forcing is a linear

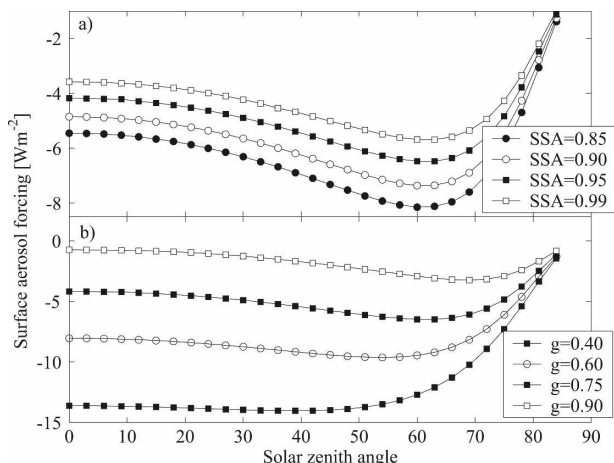


FIG. 10. Modeled surface aerosol forcing at 500 nm as a function of the solar zenith angle for (a) constant asymmetry parameter $g = 0.7$ and the values of the single scattering albedo listed, and (b) constant scattering albedo $SSA = 0.90$ and the values of the asymmetry parameter listed.

function of the optical depth and the aerosol forcing efficiency is independent on the optical depth. Thus, this nonlinear relationship is caused by the fact that in the first approximation aerosol forcing is proportional to the aerosol transmittance, which is a nonlinear function of the AOT. However, our observation (see Figs. 6 and 7) does not follow it because of the following two reasons: a large uncertainty in the aerosol radiative forcing estimation and a small range of AOT variability. In our study the AOT varies between 0.3 and 0.6 only. Nevertheless, the slope of the linear fit of aerosol radiative forcing versus AOT will be slightly different in case of a different range of AOT variability.

e. Diurnal cycle of the aerosol optical properties and the aerosol forcing land–sea modulations

Until now we discussed only the mean 24-h direct aerosol forcing during the UAE² campaign. However, diurnal variability of the aerosol forcing is important in regards to the local energy balance. In the coastal region this local energy balance determines the land–sea-breeze circulation and a modification of this balance can influence the breeze strength similar to monsoon circulation (Ramachandran 2005). At the MAARCO site we observed significant temporal variability of surface aerosol optical properties (Remiszewska et al. 2007) resulting from land–sea-breeze circulation. Therefore, in this section we discuss the diurnal variability of aerosol radiation forcing caused by aerosol optical properties change.

The diurnal cycle of the aerosol forcing depends on a solar zenith angle and aerosol optical properties, such

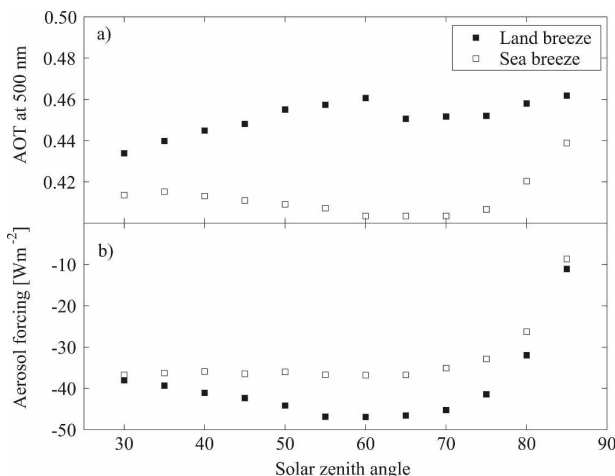


FIG. 11. (a) The aerosol optical thickness at 500 nm and (b) the aerosol radiative forcing at the surface as a function of the solar zenith angle. Afternoon sea breeze (open squares) and before noon land breeze (solid squares).

as the AOT, scattering phase function (or asymmetry parameter), and SSA. The surface aerosol forcing efficiency is associated with an upscatter fraction of solar energy; forcing becomes larger (more negative) for small aerosols particles (small asymmetry parameter). Let us consider the following two cases: when the sun is overhead and when the sun is close to the horizon. For very small particles the upscatter flux for both zenith angles is similar because of the phase function symmetry.

In general, the upscatter flux at sunset and sunrise is larger in comparison to that at local noon. However, the solar flux decreases for larger solar zenith angle. The combination of these two effects results in an aerosol forcing efficiency minimum for solar zenith angles from 50° to 70°. Figure 10a presents surface aerosol as a function of the solar zenith angle, a fixed asymmetry parameter ($g = 0.7$), and several SSAs. The magnitude of the aerosol forcing slightly decreases for absorbing aerosols because of the reduction of the scattering effect. Figure 10b shows this behavior for several values of the asymmetry parameter. For larger particles the characteristic minimum (larger asymmetry parameter) is shifted toward a larger solar zenith angle. The aerosol forcing for small particles is almost constant for the solar zenith angles less than 50°, and forcing is large in comparison to that caused by the larger particles.

Figure 11a shows diurnal variability of the AOT at 500 nm as a function of the solar zenith angle during the UAE² campaign. Solid squares indicate observations before noon (during the land breeze) and open squares are the AOT after noon (during the sea breeze). The difference of the AOT between the land and sea breeze

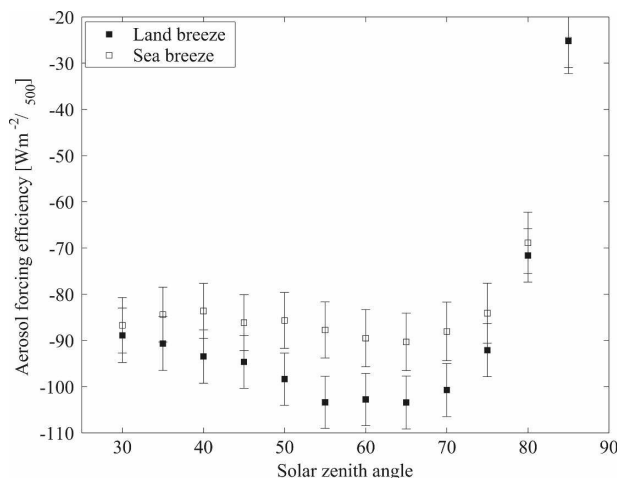


FIG. 12. Aerosol radiative forcing efficiency for land breezes (solid squares) and sea breezes (open squares) as a function of the solar zenith angle. Vertical lines show uncertainties of the aerosol efficiency.

is small but consistent with the surface observations discussed by Remiszewska et al. (2007). The onshore winds bring cleaner air, which results in decreasing values of the surface absorption and scattering coefficient; this is also observed for the columnar aerosol optical properties.

Figure 11b illustrates that the diurnal forcing is larger during the land breeze when the AOT is also larger. During the afternoon the forcing is almost constant up to the solar zenith angle of about 70° . Before noon the aerosol forcing reaches minimum for solar zenith angles in the range of 55° – 65° . Comparing Figs. 11a,b one concludes that the aerosol forcing can be explained by the increase of the single scattering albedo and the reduction of the asymmetry parameter during the sea breeze. This progression of the aerosol optical properties is consistent (Remiszewska et al. 2007) with the surface observation findings.

The diurnal cycle of the aerosol forcing efficiency (Fig. 12) has a similar daily progression to that of the aerosol forcing illustrated in Fig. 11. The 20 W m^{-2} differences for the solar zenith angle of 55° between the land and sea breeze is again caused by a variation of the aerosol optical properties. However, the daily mean forcing efficiency difference between the land and sea breezes are small and equal -8.8 W m^{-2} ; the efficiency is -4 W m^{-2} when an alternative method based on the “slope” linear fit is used. Notice that the noontime reduction of the solar radiation at the earth’s surface is relatively small and reaches -90 W m^{-2} per unit of the AOT.

We would like to emphasize that aerosol optical properties presented in this section are based on the

observation of solar radiation and columnar aerosol optical properties performed during daytime. Thus, the variability of these parameters does not reflect the entire pattern of the local sea–land–breeze circulation. A more complete discussion of the diurnal optical properties at the surface is given in Remiszewska et al. (2007).

7. Conclusions

In this paper we discuss the reduction of solar radiation at the surface observed during the UAE² campaign. Good agreement of the aerosol radiative forcing is obtained by three independent methods. The results indicate a relatively small value of the aerosol forcing efficiency at the surface (-53 W m^{-2}), although the single scattering albedo at 550 nm is not very high—0.92 on the basis of surface observations—and 0.93—based on the AERONET retrieval. This apparent inconsistency with other field projects (Haywood et al. 2003; Markowicz et al. 2003; Satheesh and Ramanathan 2000) can be described by large observed AOT, which was 0.45 at 500 nm. We show that the aerosol forcing efficiency decreases with the increasing AOT and, therefore, estimated aerosol forcing efficiency is small in comparison to other campaigns.

We found that aerosol during the UAE² campaign leads to a significant reduction of the incoming solar radiation at the surface (about 9%). Obtained values of the aerosol direct radiative forcing show that aerosol significantly influences local radiative budget partially because we observed mostly clear-sky conditions (85%).

The land–sea circulation influences the aerosol optical thickness and the aerosol forcing diurnal variability. During the sea breeze we observed a slightly smaller aerosol optical thickness and larger Ångström exponent, which was caused by reduction of the amount of mineral dust particles. Also, surface observations indicate the reduction of the absorption coefficient (Remiszewska et al. 2007) during the sea breeze. Changes to the columnar and surface optical properties lead to modification of the aerosol radiative forcing. We found an increase of the aerosol forcing efficiency during the land breeze. On the basis of model calculations and observed diurnal variability of the aerosol forcing efficiency we show that its change is due to a reduction of the single scattering albedo and an increase of the asymmetry parameter during the land breeze.

Figure 13 presents a conceptual model of the aerosol radiative forcing and its relationship with local and large-scale circulation during the UAE² experiment. All major circulations are illustrated, but not all of

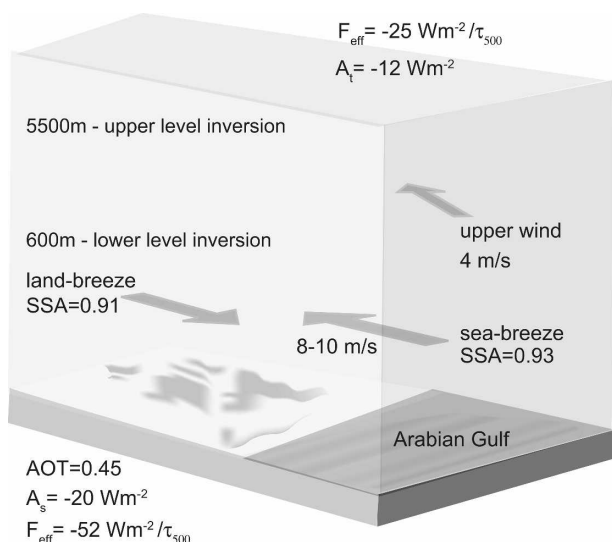


FIG. 13. A conceptual model of aerosol radiative forcing and its relationship with local and large-scale circulations.

them appear at the same time in nature. The sea breeze brings relatively cleaner air to the region (larger single scattering albedo) in comparison to the land breeze. Conceptually, the atmosphere can be divided into three layers. First, 600 m is polluted and there is a sharp decrease of dust and soot concentrations above this layer. There is an elevated large-scale subsidence region at around 5000 m. The mean flow and advection in the midtroposphere is relatively weak in comparison to strong thermal circulations. The differences between optical properties during the land and sea breeze are relatively small but observable.

Acknowledgments. We (KM and JR) would like to acknowledge support from the ONR Global. NRL's participation in this effort was provided by ONR Code 32 and NASA ESE Radiation Sciences Program. We thank for the hospitality of the UAE Department of Water Resource Studies and the Foundation for Polish Science for KM support. The NCEP–NCAR monthly reanalysis data were obtained from the NOAA/CDC Web site.

REFERENCES

- Ackerman, S. A., and S. K. Cox, 1989: Surface weather observations of atmospheric dust over the southwest summer monsoon region. *Meteor. Atmos. Phys.*, **41**, 19–34.
- Allen, G. A., J. Lawrence, and P. Koutrakis, 1999: Field validation of a semi-continuous method for aerosol black carbon (aethalometer) and temporal patterns of summertime hourly black carbon measurements in southwestern PA. *Atmos. Environ.*, **33**, 817–823.
- Anderson, T. L., and Coauthors, 1996: Performance characteristics of a high-sensitivity, three-wavelength, total scatter/backscatter nephelometer. *J. Atmos. Oceanic Technol.*, **13**, 967–986.
- Bates, T. S., and Coauthors, 1998: Processes controlling the distribution of aerosol particles in the lower marine boundary layer during the First Aerosol Characterization Experiment (ACE 1). *J. Geophys. Res.*, **103**, 16 369–16 383.
- Berk, A., L. S. Bernstein, G. P. Anderson, P. K. Acharya, D. C. Robertson, J. H. Chetwynd, and S. M. Adler-Golden, 1998: MODTRAN cloud and multiple scattering upgrades with application to AVIRIS. *Remote Sens. Environ.*, **65**, 367–375.
- Bohren, C. F., and D. Huffman, 1983: *Absorption and Scattering of Light by Small Particles*. Wiley, 544 pp.
- Bruegge, C. T., J. E. Conel, R. O. Green, J. S. Margolis, R. G. Holm, and G. Toon, 1992: Water vapor column abundance retrieval during FIFE. *J. Geophys. Res.*, **97**, 18 759–18 768.
- Bush, B. C., F. P. J. Valero, A. S. Simpson, and L. Bignone, 2000: Characterization of thermal effects in pyranometers: A data correction algorithm for improved measurement of surface insolation. *J. Atmos. Oceanic Technol.*, **17**, 165–175.
- Charlson, R. J., J. Langner, H. Rodhe, C. B. Leovy, and S. G. Warren, 1991: Perturbation of the northern hemisphere radiative balance by backscattering from anthropogenic sulfate aerosols. *Tellus*, **43A–B**, 152–163.
- Chin, M., and Coauthors, 2002: Tropospheric aerosol optical thickness from the GOCART model and comparisons with satellite and sun photometer measurements. *J. Atmos. Sci.*, **59**, 461–483.
- Collins, W. D., P. J. Rasch, B. E. Eaton, B. V. Khattatov, J. F. Lamarque, and C. S. Zender, 2001: Simulating aerosols using a chemical transport model with assimilation of satellite aerosol retrievals: Methodology for INDOEX. *J. Geophys. Res.*, **106**, 7313–7336.
- Deepshikha, S., S. K. Satheesh, and J. Srinivasan, 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., and Coauthors, 2006: Dust and pollution aerosols over the Negev desert, Israel: Properties, transport, and radiative effect. *J. Geophys. Res.*, **111**, D05205, doi:10.1029/2005JD006549.
- Draxler, R. R., J. T. McQueen, and B. J. B. Stunder, 1994: An evaluation of air pollutant exposures due to the 1991 Kuwait oil fires using a Lagrangian model. *Atmos. Environ.*, **28**, 2197–2210.
- Drummond, A., and J. Roche, 1965: Corrections to be applied to measurements made with Eppley (and other) spectral radiometers when used with Schott colored glass filters. *J. Appl. Meteor.*, **4**, 741–744.
- Dubovik, O., and M. D. King, 2000: A flexible inversion algorithm for retrieval of aerosol optical properties from Sun and sky radiance measurements. *J. Geophys. Res.*, **105**, 20 673–20 696.
- , and Coauthors, 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.
- Fouquart, Y., B. Bonnel, and V. Ramaswamy, 1991: Intercomparing shortwave radiation codes for climate studies. *J. Geophys. Res.*, **96**, 8955–8968.
- Goudie, A. S., and N. J. Middleton, 2001: Saharan dust storms: Nature and consequences. *Earth-Sci. Rev.*, **56**, 179–204.
- Halothore, R. N., T. F. Eck, B. N. Holben, and B. L. Markham,

- 1997: Sunphotometric measurements of atmospheric water vapor column abundance in the 940-nm band. *J. Geophys. Res.*, **102**, 4343–4352.
- , and Coauthors, 2005: Intercomparison of shortwave radiative transfer codes and measurements. *J. Geophys. Res.*, **110**, D11206, doi:10.1029/2004JD005293.
- Hansen, A. D. A., H. Rosen, and T. Novakov, 1996: The aethalometer—An instrument for the real-time measurement of optical-absorption by aerosol-particles. *Sci. Total Environ.*, **36**, 191–196.
- Hansen, J., M. Sato, and R. Ruedy, 1997: Radiative forcing and climate response. *J. Geophys. Res.*, **102**, 6831–6864.
- Haywood, J., and Coauthors, 2003: Radiative properties and direct radiative effect of Saharan dust measured by the C-130 aircraft during SHADE: 1. Solar spectrum. *J. Geophys. Res.*, **108**, 8577, doi:10.1029/2002JD002687.
- Haywood, J. M., V. Ramaswamy, and B. J. Soden, 1999: Tropospheric aerosol climate forcing in clear-sky satellite observations over the oceans. *Science*, **283**, 1299–1303.
- Hobbs, P. V., 1999: An overview of the University of Washington airborne measurements and results from the Tropospheric Aerosol Radiative Forcing Observational Experiment (TARFOX). *J. Geophys. Res.*, **104**, 2233–2238.
- , and L. F. Radke, 1992: Airborne studies of the smoke from the Kuwait oil fires. *Science*, **256**, 987–991.
- Holben, B. N., and Coauthors, 1998: AERONET—A federated instrument network and data archive for aerosol characterization: An overview. *Remote Sens. Environ.*, **66**, 1–16.
- , and Coauthors, 2001: An emerging ground-based aerosol climatology: Aerosol optical depth from AERONET. *J. Geophys. Res.*, **106**, 12 067–12 097.
- Houghton, J. T., Y. Ding, D. J. Griggs, M. Noguer, P. J. van der Linden, X. Dai, K. Maskell, and C. A. Johnson, 2001: *Climate Change 2001: The Scientific Basis*. Cambridge University Press, 892 pp.
- Im, J., V. K. Saxena, and B. N. Wenny, 2001: An assessment of hygroscopic growth factors for aerosols in the surface boundary layer for computing direct radiative forcing. *J. Geophys. Res.*, **106**, 20 213–20 224.
- Kaufman, Y. J., and Coauthors, 1998: Smoke, Clouds, and Radiation-Brazil (SCAR-B) experiment. *J. Geophys. Res.*, **103**, 31 783–31 808.
- Kay, M. J., and M. Box, 2000: Radiative effects of absorbing aerosols and the impact of water vapor. *J. Geophys. Res.*, **105**, 12 221–12 234.
- Key, J., and A. J. Schweiger, 1998: Tools for atmospheric radiative transfer: Streamer and FluxNet. *Comput. Geosci.*, **24**, 443–451.
- Kipp & Zonen, 2001: CH1 normal incidence pyrliometer instruction manual. 42 pp. [Available online at http://www.kippzonen.com/download/kipp_manual_ch1_1880.pdf.]
- , 2004: CMP series pyranometer and CMA series albedometer instruction manual. 36 pp. [Available online at http://www.kippzonen.com/download/kipp_manual_cmp22_1737.pdf.]
- Klett, J. D., 1985: Lidar inversion with variable backscatter/extinction ratios. *Appl. Opt.*, **24**, 1638–1643.
- Langner, J., H. Rodhe, P. Crutzen, and P. Zimmerman, 1992: Anthropogenic influence on the distribution of tropospheric sulfate aerosol. *Nature*, **359**, 712–715.
- Lelieveld, J., and Coauthors, 2003: Global air pollution crossroads over the Mediterranean. *Science*, **298**, 794–799.
- Léon, J.-F., and M. Legrand, 2003: Mineral dust sources in the surroundings of the north Indian Ocean. *Geophys. Res. Lett.*, **30**, 1309, doi:10.1029/2002GL016690.
- Markowicz, K. M., and Coauthors, 2003: Influence of relative humidity on aerosol radiative forcing: An ACE-Asia experiment perspective. *J. Geophys. Res.*, **108**, 8662, doi:10.1029/2002JD003066.
- Mohalfi, S., H. S. Bedi, T. N. Krishnamurti, and S. D. Cocker, 1998: Impact of shortwave radiative effects of dust aerosols on the summer season heat low over Saudi Arabia. *Mon. Wea. Rev.*, **126**, 3153–3168.
- Moody, E. G., M. D. King, C. B. Schaaf, and S. Platnick, 2008: MODIS-derived spatially complete surface albedo products: Spatial and temporal pixel distribution and zonal averages. *J. Appl. Meteor. Climatol.*, in press.
- Nakajima, T., T. Hayasaka, A. Higurashi, G. Hashida, N. Moharram-Nejad, Y. Najafi, and H. Valavi, 1996: Aerosol optical properties in the Iranian region obtained by ground-based solar measurements in the summer of 1991. *J. Appl. Meteor.*, **35**, 1265–1278.
- Raes, F., T. Bates, F. McGovern, and M. van Liedekerke, 2000: The 2nd Aerosol Characterization Experiment (ACE-2): General overview and main results. *Tellus*, **52B**, 111–125.
- Ramachandran, S., 2005: Premonsoon shortwave aerosol radiative forcings over the Arabian Sea and tropical Indian Ocean: Yearly and monthly mean variabilities. *J. Geophys. Res.*, **110**, D07207, doi:10.1029/2004JD005563.
- Ramanathan, V., P. J. Crutzen, J. T. Kiehl, and D. Rosenfeld, 2001a: Atmosphere—Aerosols, climate, and the hydrological cycle. *Science*, **294**, 2119–2124.
- , and Coauthors, 2001b: 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.
- Redemann, J., R. P. Turco, K. N. Liou, P. V. Hobbs, W. S. Hartley, R. W. Bergstrom, E. V. Browell, and P. B. Russell, 2000: Case studies of the vertical structure of the direct shortwave aerosol radiative forcing during TARFOX. *J. Geophys. Res.*, **105**, 9971–9979.
- Remiszewska, J., P. J. Flatau, K. M. Markowicz, E. A. Reid, J. S. Reid, and M. L. Witek, 2007: Modulation of the aerosol absorption and single-scattering albedo due to synoptic scale and sea breeze circulations: United Arab Emirates experiment perspective. *J. Geophys. Res.*, **112**, D05204, doi:10.1029/2006JD007139.
- Satheesh, S. K., and V. Ramanathan, 2000: Large differences in tropical aerosol forcing at the top of the atmosphere and Earth's surface. *Nature*, **405**, 60–63.
- , S. Deepshikha, and J. Srinivasan, 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.
- Schmid, B., and Coauthors, 1999: Comparison of aerosol optical depth from four solar radiometers during the fall 1997 ARM Intensive Observing Period. *Geophys. Res. Lett.*, **26**, 2725–2728.
- , and Coauthors, 2001: Comparison of columnar water vapor measurements during the fall 1997 ARM Intensive Observation Period: Solar transmittance methods. *Appl. Opt.*, **40**, 1886–1896.
- , and Coauthors, 2006: How well do state-of-the-art techniques measuring the vertical profile of tropospheric aerosol extinction compare? *J. Geophys. Res.*, **111**, D05S07, doi:10.1029/2005JD005837.
- Seinfeld, J. H., and S. N. Pandis, 1998: *Atmospheric Chemistry and*

- Physics: From Air Pollution to Climate Change*. John Wiley & Sons, 1326 pp.
- Smirnov, A., and Coauthors, 2002: Atmospheric aerosol optical properties in the Persian Gulf. *J. Atmos. Sci.*, **59**, 620–634.
- Stamnes, K., S. C. Tsay, W. Wiscombe, and K. Jayaweera, 1998: Numerically stable algorithm for discrete-ordinate-method radiative transfer in multiple scattering and emitting layered media. *Appl. Opt.*, **27**, 2502–2509.
- Takemura, T., T. Nakajima, O. Dubovik, B. N. Holben, and S. Kinne, 2002: Single-scattering albedo and radiative forcing of various aerosol species with a global three-dimensional model. *J. Climate*, **15**, 333–352.
- Tanré, D., and Coauthors, 2003: Measurement and modeling of the Saharan dust radiative impact: Overview of the Saharan Dust Experiment (SHADE). *J. Geophys. Res.*, **108**, 8574, doi:10.1029/2002JD003273.
- Voss, K. J., E. J. Welton, P. K. Quinn, J. Johnson, A. M. Thompson, and H. R. Gordon, 2001: Lidar measurements during Aerosols99. *J. Geophys. Res.*, **106**, 20 821–20 832.
- Welton, E. J., and J. R. Campbell, 2002: Micropulse lidar signals: Uncertainty analysis. *J. Atmos. Oceanic Technol.*, **19**, 2089–2094.
- , and Coauthors, 2000: Ground-based lidar measurements of aerosols during ACE-2: Instrument description, results, and comparisons with other ground-based and airborne measurements. *Tellus*, **52B**, 636–651.
- Wielicki, B. A., B. R. Barkstrom, E. F. Harrison, R. B. Lee III, G. L. Smith, and J. E. Cooper, 1996: Clouds and the Earth's Radiant Energy System (CERES): An Earth observing system experiment. *Bull. Amer. Meteor. Soc.*, **77**, 853–868.
- Zhu, M., and B. W. Atkinson, 2004: Observed and modeled climatology of the land-sea breeze circulation over the Persian Gulf. *Int. J. Climatol.*, **24**, 883–905.