Solar and thermal radiative effects during the 2011 extreme desert dust episode over Portugal

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HIGHLIGHTS

- Aerosol thermal capability to offset part of the solar radiative effect.
- Lidar and sun photometer data to initialize a radiative model.
- Aerosol could enhance the response between absorption and re-emission processes.

ARTICLE INFO

Article history:
Received 17 May 2016
Received in revised form 17 October 2016
Accepted 20 October 2016
Available online 20 October 2016

Keywords:
Dust
Aerosol vertical profiles
EARLINET
AERONET
Forcing

ABSTRACT

This paper analyses the influence of the extreme Saharan desert dust (DD) event on shortwave (SW) and longwave (LW) radiation at the EARLINET/AERONET Évora station (Southern Portugal) from 4 up to 7 April 2011. There was also some cloud occurrence in the period. In this context, it is essential to quantify the effect of cloud presence on aerosol radiative forcing. A radiative transfer model was initialized with aerosol optical properties, cloud vertical properties and meteorological atmospheric vertical profiles. The intercomparison between the instantaneous TOA shortwave and longwave fluxes derived using CERES and those calculated using SBDART, which was fed with aerosol extinction coefficients derived from the CALIPSO and lidar-PAOLI observations, varying OPAC dataset parameters, was reasonably acceptable within the standard deviations. The dust aerosol type that yields the best fit was found to be the mineral accumulation mode. Therefore, SBDART model constrained with the CERES observations can be used to reliably determine aerosol radiative forcing and heating rates. Aerosol radiative forcings and heating rates were derived in the SW (ARFSw, AHRSw) and LW (ARFLw, AHIRLw) spectral ranges, considering a cloud–aerosol free reference atmosphere. We found that AOD at 440 nm increased by a factor of 5 on 6 April with respect to the lower dust load on 4 April. It was responsible by a strong cooling radiative effect pointed out by the ARFSw value (−99 W/m² for a solar zenith angle of 60°) offset by a warming radiative effect according to ARFLw value (+21.9 W/m²) at the surface. Overall, about 24% and 12% of the dust solar radiative cooling effect is compensated by its longwave warming effect at the surface and at the top of the atmosphere, respectively. Hence, larger aerosol loads could enhance the response between the absorption and re-emission processes increasing the ARFlw with respect to those associated with moderate and low aerosol loads. The unprecedented results derived from this work complement the findings in other regions on the modifications of radiative energy budget by the dust aerosols, which could have relevant influences on the regional climate and will be topics for future investigations.

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1. Introduction

Desert dust (DD) plays an important role in many atmospheric processes modifying the Earth’s energy balance on shortwave (SW) and longwave (LW) spectral ranges through direct effects (scattering and absorption) and indirect effects (cloud formation and
layering is a critical point in order to reliably retrieve radiative effects in the entire atmospheric column. In fact, most of studies in the Iberian Peninsula have focused on the evaluation of DD radiative effects taking into account aerosol properties integrated in the atmospheric column. However, nowadays a large uncertainty still remains in the evaluation of total radiative effects, mainly caused by uncertainties on the estimation of dust aerosol properties (IPCC, 2013). Such uncertainties on the dust optical and microphysical properties are due to the high variability in space and time of the particle’s size, chemical composition, shape and vertical distribution affecting their light-absorbing and scattering properties. One of reasons for this variability is that dust optical properties at origin, as issued by the sources, are influenced by the potential changes during the path of the air masses as well as by the local aerosol properties at the reception site (e.g., Hand et al., 2010; Bauer et al., 2011; Valenzuela et al., 2014). Close to the source regions, mostly pure dust is found, but after long-range transport the aging of dust and mixing with other aerosol types modify the optical properties of DD (e.g., Bauer et al., 2011).

The Mediterranean basin is a place of special interest for the study of the atmospheric aerosol due to its strategic geographic location surrounded by continents with different surface characteristics and where local winds, complex coastlines and orography have strong influence on the atmospheric flow. North African dust is injected into the atmosphere through resuspension processes in source regions, and transported at different altitudes (up to 7 km) to different regions of the world (e.g., Tesche et al., 2009, Papayannis et al. (2008) reported more than 130 daily observation of the horizontal and vertical extent of Saharan dust intrusions over Europe in the frame of the European Aerosol Research Lidar Network (EARLINET). They found that Saharan dust source regions play a key role in dust transport to Europe for the height layer between 3 and 5 km. Particularly, the Iberian Peninsula is frequently affected by African dust intrusions with large aerosol loads that modulate the aerosol climatology in different areas of this region (e.g., Toledano et al., 2007; Cachorro et al., 2008; Guerrero-Rascado et al., 2008, 2009; Wagner et al., 2009; Pereira et al., 2011; Peißler et al. (2011); Valenzuela et al., 2012a; Obregon et al., 2015). Moreover, in many studies the vertical aerosol distribution is assumed to be homogeneous in the entire atmospheric column causing erroneous estimates of the radiative effects both at the surface and at the top of the atmosphere. In fact, most of studies in the Iberian Peninsula performed the evaluation of the DD radiative effects taking into account aerosol properties integrated in the atmospheric column (Santos et al., 2008; Antón et al., 2012; Valenzuela et al., 2012b; Román et al., 2013; Obregón et al., 2015). However, the aerosol layering is a critical point in order to reliably retrieve radiative effects at the surface, top of the atmosphere and within the atmosphere. Gómez-Amo et al. (2014) used a radiative transfer model to simulate SW and LW irradiances under different vertical profiles of the aerosol absorption during a DD event over Rome, Italy on 20 June 2007. Their calculations showed that different LW/all-wave was observed in the boundary layer depending on the boundary layer aerosol absorption. Many authors have provided information about DD radiative effects in SW interval during last years (e.g., Meloni et al., 2005; Derimian et al., 2006; Tafuro et al., 2007; Bergamo et al., 2008; Saha et al., 2008; di Sarra et al., 2008; di Biagio et al., 2010; Sicard et al., 2012; Román et al., 2013; Esteve et al., 2014). However, few studies look into the LW radiative effects associated with DD intrusions over the Mediterranean basin (e.g., di Sarra et al., 2011; Perrone et al., 2012; Sicard et al., 2014; Gómez-Amo et al., 2014; Meloni et al., 2015; Romano et al., 2016). The scarcity of papers related to the DD effects on LW interval in Mediterranean regions against the numerous works about the effects in the SW spectral range is mainly due to the fact that the aerosol LW effect is often small and is within the uncertainties associated with the LW measurements and the model estimates. Most aerosol types, as anthropogenic particles (fine aerosols), have a negligible radiative effect in thermal spectral range compared with the radiative effects in solar interval (Sicard et al., 2014). However aerosols like dust may enhance the greenhouse effect by trapping the outgoing terrestrial radiation through absorption and scattering processes and lead to a positive radiative effect in the LW (Hansell et al., 2010). Some authors point out that, on the daily time scale, the LW effect may compensate for a large fraction (up to 50%) of the substantial decrease of SW irradiance recorded at the surface during DD episodes. Sicard et al. (2014) reported that the scattering process associated with DD particles may notably contribute to the LW radiative effect at surface up to 18%.

Finally, the presence of clouds can noticeably enhance the radiative impact of aerosols, especially when the aerosol layer is above the cloud layer (Podgorny and Ramanathan, 2001). This is one of the most challenging problems due to lack of continuous measurements of aerosols and clouds (IPCC 2013). In this framework, this work aims to determine the aerosol radiative forcing and heating rate profiles in the SW and LW spectral ranges, taking into account sun-photometer and lidar measurements during an extreme DD event over Évora, Portugal that has been given as input to a radiative transfer model (SBDART). In addition, in our simulations we always deemed a cloud layer whose optical properties were derived from CALIPSO platform and ECMWF meteorological model. In order to validate the reflected solar radiation measured at the top of the atmosphere (TOA) from the Clouds and Earth’s Radiant Energy System (CERES) (Weilicki et al. 1996) Single Satellite Footprint (SSF), vertical distributions of aerosol extinction coefficients derived from the CALIPSO lidar observations together with OPAC dataset parameters were used to constrain the dust aerosol type employed in the radiation model. The main purpose of this study is the assessment of the thermal capability to offset part of the solar radiative effect. The analyzed dust episode took place from 4 up to 7 April 2011 and it has been identified as one of the strongest DD events recorded over the Iberian Peninsula in the last years. This atypical high-aerosol load case was described in detail by Peißler et al. (2011) using active and passive remote sensing techniques. Therefore, this paper completes the analysis about dust radiative effects, contributing to understand the influence of DD on SW and LW irradiances at the surface, within the atmosphere as well as its top.

2. Experimental site, instrumentation and measurements

2.1. Experimental site description

The instruments used in this study are installed at the Institute of Earth Sciences (ICT) in Évora (38.57°N, 7.91°W and 290 m a.s.l.). Évora, located in the Southern Portugal, is the capital of the Alentejo district, a rural region of Portugal. Although this region covers about one third of the area of Portugal it is only inhabited by less than 10% of its population. Évora is a non-industrialized city situated about 100 km from the Western coast of Portugal and the nearest urban and industrial area is that of Lisbon. The main aerosol sources are traffic and, during the winter season, domestic fuel burning injecting anthropogenic aerosols into the atmosphere. Due to its proximity to the African continent, our study area is often affected by DD intrusions (e.g., Elias et al., 2006; Pereira et al., 2008; Wagner et al., 2009; Peißler et al., 2011).

2.2. AERONET data

The aerosol optical depth (AOD ($\lambda$)), angstrom exponent ($\alpha$), single scattering albedo ($w(\lambda)$) and asymmetry parameter ($g(\lambda)$) were taken from the sun photometer inversion products. Measurements of total columnar aerosol properties were obtained...
using a CIMEL CE-318 sun photometer included in the AERONET network (Holben et al., 1998). This sun-photometer makes direct sun measurements with a 1.2° full field of view at 340, 380, 440, 500, 675, 870, 1020 and 1640 nm. The full-width at half-maximum functions of the interference filters are 2 nm at 340 nm, 4 nm at 380 nm and 10 nm at all other wavelengths. The sky radiance measurements (almucantar configuration) are carried out at 440, 670, 870 and 1020 nm. This instrument is fully described by Holben et al. (1998). The direct sun measurements are used to compute the AOD. In this work, AERONET AOD level 2 data (cloud screened and quality assured) are used. The uncertainty in the retrieval of AOD under cloud free conditions is ±0.01 for wavelengths larger than 440 nm and ±0.02 for shorter wavelengths (Eck et al., 1999). Sky radiance measurements together with solar direct irradiance measurements are used to retrieve aerosol optical properties like single scattering albedo, \( \omega(\lambda) \), using the AERONET inversion algorithm developed by Dubovik and King (2000) as improved by Dubovik et al. (2006). The uncertainty in the retrieval of \( \omega(\lambda) \) is ±0.03 for high aerosol load (AOD (440 nm) > 0.4) and solar zenith angle > 50°. For measurements with low aerosol load (AOD (440 nm) < 0.2), the retrieval accuracy of \( \omega(\lambda) \) drops down to 0.02–0.07 (Dubovik et al., 2006). There were no \( \omega(\lambda) \) level 2 retrievals at Évora on 4 and 5 April 2011 due to strong limitations imposed by the AERONET inversion algorithm (AOD (440 nm) > 0.4 and solar zenith angle > 50°) and reduced sampling of almucantar sky radiance measurements as well as the presence of clouds during measurements. Thus, the AERONET \( \omega(\lambda) \) level 1.5 (cloud screened data with pre and post calibrations applied) for AOD >0.2 and solar zenith angle > 50° were used here for these two days. Other studies also employed successfully level 1.5 AERONET \( \omega(\lambda) \) data (Iyamani et al., 2012; Obregón et al., 2015). AERONET \( \omega(\lambda) \) level 2 were retrieved on 6 and 7 April 2011.

2.3. Lidar system

Vertical profiles of atmospheric aerosol data were derived from the lidar system PAOLI (Portable Aerosol and Cloud Lidar), a multiwavelength Raman lidar belonging to the PollyXT family (Althausen et al., 2009) with high temporal and spatial resolution. PAOLI is operated at ICT (formerly CGE) since September 2009. It is included in EARLINET (European Aerosol Research Lidar Network) (Pappalardo et al., 2014) and SPALINET (Spanish and Portuguese Aerosol Lidar Network) (Sicard et al., 2011). The PAOLI system provides profiles of particle extinction coefficients at 355 and 532 nm, particle backscatter coefficients at 355, 532 and 1064 nm and depolarization ratios at 532 nm. Due to the gap in PAOLI data in the particle backscatter coefficients at 532 nm on 5 April, only vertical particle backscatter coefficients at 355 nm were used. The ratio between the extinction and the backscatter is the height-independent lidar ratio (LR). We chose the LR so that the integral of the vertical extinction coefficient reproduced the AOD provided by AERONET (Landulfo et al., 2003; Guerrero-Rascado et al., 2011). In this study, only daytime lidar data from the quality-assured EARLINET database were considered. We assumed only one aerosol vertical profile per day representative of the aerosol vertical distribution in those 24 h, obtained from the average of the profiles available for that day. The lidar system and its optical analysis procedure during this dust event are detailed described in Preißler et al. (2011).

2.4. Satellite information: CALIPSO and CERES data

The Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) onboard the NASA’s satellite CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations) provides vertical resolved aerosol and cloud observations (Winker et al., 2009). The CALIPSO lidar is designed to acquire vertical profiles of elastic backscatter at two wavelengths (532 and 1064 nm) from a near nadir-viewing geometry during both day and night phases of the Sun-synchronous orbit, which has a 13:30 LT equatorial crossing time (Winker et al., 2009). In this study, we use the Version 3 (3.01 and 3.02) of the Level 2 Vertical Feature Mask (VFM) and Aerosol Profile Products (APro) files, available from June 2006 to February 2013, both derived from the NASA’s Earth Observing System Data and Information System (http://reverb.echo.nasa.gov/). The aerosol profile products are generated at a uniform horizontal resolution of 5 km (http://www-calipso.larc.nasa.gov/products/CALIPSO_DPC_Rev3x6.pdf), while the vertical resolution varies 20 from 60 to 180 m depending on the altitude range and the parameter.

CERES sensors currently operating on the EOS (Earth Observing System) Terra and Aqua platforms have provided data used in this paper. We have analyzed the SW and LW radiative flux at the TOA available in the Single Scanner Footprint (SSF) Level 2 products of the CERES sensors. TOA fluxes, meteorological data, and cloud property retrievals are incorporated into the CERES processing stream to evaluate radiative fluxes at the surface (Kratz et al., 2010). The SW and LW data were retrieved with the PAOLI and AERONET \( \omega(\lambda) \) level 2. The uncertainty in the TOA fluxes, meteorological data, and cloud property retrievals are incorporated into the CERES processing stream to evaluate radiative fluxes at the surface (Kratz et al., 2010).

3. Radiative transfer procedure

3.1. Radiative transfer code

The Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) model (Ricchiazzi et al., 1998) was set up to estimate SW (0.31–4 μm) and LW (4–40 μm) irradiances from the surface until top of the atmosphere (TOA). The model runs with sixteen streams using DISORT algorithm for SW irradiances, while LW irradiances were estimated using k-correlated method. Atmospheric parameters from ECMWF (European Centre for Medium-Range Weather Forecasts) meteorological model (ERA-Interim) were used from 1 up to 10 km altitude. As the DD event was accompanied of some clouds, a cloud vertical layer has been deemed in our simulations. It can be considered as an optically thin stratus continental low-level cloud. In this sense, specific cloud liquid water content in each atmospheric layer was derived from ECMWF meteorological model. Hence, according with this data and CALIPSO information, a cloud optical depth at 532 nm of 0.3 at 2 km of altitude was used as input in SBDART model. In this sense, we have assumed some other cloud properties based on the available bibliography (Subramanyam and Kumar 2013). Above 10 km of altitude, vertical profiles of pressure, temperature, water vapor and oxygen were extended with corresponding mid-latitudes standard atmosphere data provided by the Air Force Geophysics Laboratory (AFGL) for summer months (Anderson et al., 1986). Vertical profiles of oxygen, ozone and well-mixed trace gases were established by the AFGL standard atmosphere for mid-latitudes in summer months. The vertical extinction coefficients at 440, 675, 870 and 1020 nm were then obtained from the one at 355 nm (provided by the lidar system) using the Ångström relationship and the AERONET Ångström parameter \( \alpha \). Inputs feeding the model include: vertical profile of meteorological parameters (temperature, water vapor, pressure and ozone) and those properties related with the aerosol extinction as AOD, \( \omega \) and g retrieved at four wavelengths (440, 675, 870 and 1020 nm) provided by AERONET and the vertical extinction coefficient retrieved with the PAOLI system. The aerosol optical properties in these wavelengths were used to interpolate and extrapolate into the spectral divisions of the SBDART model (Xia et al., 2007). \( \omega \) and g parameters were kept constant in the whole atmospheric column. One uncertainty in the radiative effect calculations is the lack of knowledge about the vertical variability of aerosol optical properties, especially \( \omega \) and g parameters. Guan et al. (2010) found that the difference in radiative
forcing for different $\omega$ profiles cause small differences at the surface but differences up to 10% at TOA if absorbing aerosol is assumed to be present in the elevated layer. These uncertainties become obvious when radiative transfer calculations assume a constant column-mean $\omega$ for the multilayer aerosol vertical distribution, especially when aerosols originate from different source regions for each layer (Wang et al., 2010). Based on the PAOLI system measurements no cases of multiple aerosol layers were found in our study.

The AOD, $\omega$ and g for mineral accumulation mode at wavelengths larger than 4 $\mu$m were derived from the mineral dust aerosol type by the Optical Properties of Aerosols and Clouds (OPAC) dataset (Hess et al., 1998).

Input data include the total ozone column derived from the satellite Ozone Monitoring Instrument (OMI) and the surface spectral albedo provided by the AERONET algorithm, based on dynamic spectral and spatial model estimation at four wave-lengths: 440, 675, 870 and 1020 nm. For land surface covers, the Lie-Ross model was adopted (Lucht and Roujean, 2000), considering the bidirectional reflectance distributions from MODIS (Moody et al., 2005). Mean values of surface spectral albedo for SW of 0.05 $\pm$ 0.04 at 440 nm, 0.16 $\pm$ 0.03 at 675 nm, 0.31 $\pm$ 0.04 at 870 nm and 0.32 $\pm$ 0.04 at 1020 nm for the analyzed DD days were used in this work. A thermal surface emissivity value of 0.97 was assumed for LW above 4 $\mu$m.

3.2. Aerosol radiative forcing

Aerosol radiative forcing (ARF) at the surface and TOA is defined as the instantaneous increase or decrease of the net radiation flux at the surface (TOA) that is due to an instantaneous change in the aerosol atmospheric content. The atmospheric ARF is defined as the radiative forcing at TOA minus the radiative forcing at the surface.

In this study, we have chosen the atmosphere without aerosol particles and without cloud as the reference case. Thus, the instantaneous values of ARF can be derived from the following expression (Meloni et al., 2005):

$$\text{ARF} \approx \left( F_1^\downarrow - F_1^\uparrow \right) - \left( F_0^\downarrow - F_0^\uparrow \right)$$

where $F_1^\downarrow$ and $F_0^\downarrow$ denote the global irradiances with/without aerosol, respectively. The arrows indicate the direction of the global irradiances (downward and upward). Aerosol radiative forcing was assessed in the SW (ARFSw) and LW (ARFLw) spectral ranges.

Most of the literature on SW and LW dust radiative effects report values averaged over the daylight hours or over the 24 h. We estimated daily average ARFSw by assuming that atmospheric optical properties remained unchanged during 24 h and only SZA changed. Regarding the aerosol vertical profile, we also assume that it does not change with time. Similarly, a cloud vertical profile was deemed constant over 24 h. Although this is a restriction in our study, we are mainly interested in assessing how much the aerosol mineral particles cool/heat the atmosphere. The output parameters of the code are the downward and the upward global SW and LW irradiance at the surface and at the TOA. In order to take into account the sun-elevation changes, simulations were addressed in eight different solar zenith angles (SZAs) varying from 10° to 80°. Vertical profiles of aerosol radiative forcing values were derived assuming that all the other model inputs do not vary during the day. The daily values of ARFSw are calculated by integrating the instantaneous values of ARFSw over 24 h. A similar assumption was made for the ARFLw.

The atmospheric heating rate was computed in the SW (AHRSw) and LW (AHRLw) spectral ranges for each layer, based on finite difference estimates of the irradiance divergence at each pair of levels (Liou, 2002);

$$\frac{dT}{dt} = \frac{g}{C_{pd}} \frac{\Delta F_{\text{Atmospheric}}}{\Delta p}$$

where $T$ is the temperature (K), $t$ is the time (s), $g$ is the gravitational acceleration (9.8 m/s²), $C_{pd}$ is the specific heat of dry air at constant pressure, $F$ is the net all-wave flux (W/m²), and $p$ is the pressure (Pa). In this study, we calculated the aerosol heating rate for the whole atmospheric column which is the difference in heating rates between an aerosol laden and an aerosol free atmosphere.

3.3. Comparison of output SBDART model against CERES solar and thermal fluxes

As an attempt to validate the solar and thermal results modeled with SBDART code at TOA, we compared the upward TOA flux, calculated in this work, with those provide from CERES sensors aboard Terra and Aqua over Evora region. In order to optimize the aerosol single scattering albedo and asymmetry parameter used over this area, we compared the CERES TOA solar and thermal fluxes with SBDART model simulations along the CALIPSO orbit following the recommendations given by Huang et al. (2009). Unfortunately, for this DD event only two CALIPSO overpasses were coincident with CERES products on 5 and 7 April (Fig. 1a and b). We have tested this comparison using the 4 four different dust aerosol types from OPAC dataset for these two days (varying the aerosol single scattering albedo and asymmetry parameter). The dust aerosol type that fits best both data sets was found to be the mineral accumulation mode (Table 1). On the other hand, $\omega$ and g daily averaged values from AERONET measurements on 5 April are plotted (symbols) versus wavelengths in Fig. 2a and b, respectively. Solid lines in these Figures represent $\omega$ and g values for mineral accumulation mode type by OPAC dataset, respectively. AERONET $\omega$ and g values retrieved at 440, 675, 870 and 1020 nm agree well with the corresponding OPAC values within their uncertainty limits. As for 4 and 6 April, no CALIPSO data were available; hence, we used ground-based extinction coefficients derived from lidar-PAOLI and single scattering albedo and asymmetry parameter retrieved from AERONET data set (Table 2), in order to compare modeled flux using SBDART against flux retrieved from CERES for these two days. The intercomparison between the instantaneous TOA at shortwave and longwave fluxes obtained using CERES and those calculated using SBDART are shown in Fig. 3a and b. The comparison shows that the agreements between calculated and measured upward TOA shortwave and longwave fluxes are reasonably acceptable within the standard deviations. The apparent mismatch of about 4 W/m² (SBDART-CALIPSO) and 6 W/m² (SBDART-PAOLI) between the calculated and measured values (CERES) represents approximately 3% of the upward shortwave flux at the TOA. The averaged difference between the longwave flux model simulations and CERES measurements in the worst of cases (SBDART-PAOLI) is 8 W/m². Therefore it is clear from these comparisons that the SBDART model constrained with the CERES observations can be used to reliably determine cloud-aerosol radiative forcing and heating rates with the input of aerosol vertical distributions. Besides these comparisons, we can use both extinction coefficients from CALIPSO measurements or from ground-based lidar-PAOLI measurements indistinctly. As CALIPSO only provides profiles for two days, ground-based lidar-PAOLI measurements were used for our calculations during the whole the DD event.
4. Results and discussion

4.1. Evolution of the aerosol optical properties during the dust event

In this study the radiative effects of an intense DD episode affecting Évora site during four days (4–7 April 2011) was evaluated. In Fig. 4c–d, four satellite images obtained from the Moderate Resolution Imaging Spectroradiometer (MODIS) aboard NASA’s Terra satellite, illustrate the position, evolution and extension of the dust plume. The intensity of the event can be quickly anticipated. Fig. 4a corresponds to the image on 4 April, when the plume still did not appear over the Iberian Peninsula. Fig. 4b shows an image on 5 April, where the dust plume contaminated by clouds can be observed reaching the southwestern coast of Portugal. In Fig. 4c and d, which correspond to 6 and 7 April, respectively, the dust outbreak over the Iberian Peninsula and hence over Évora station placed in the central south of Portugal, can be clearly appreciated.

The air mass pathways from its origin until Évora monitoring station were analyzed for each day during the DD episode using back trajectories. Five back trajectories from 4 up to 7 April 2011 are shown in Fig. 5a–d. Back trajectories were calculated at 12:00 UTC using the Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT model available online at http://ready.arl.noaa.gov/HYSPLIT.php) at three levels 500, 1500 and 3000 m a.g.l. arriving over Évora station.

Fig. 2. a) Single scattering albedo and b) asymmetry parameter values on daily average from AERONET sun/sky photometer measurements retrieved on April 5, 2011 (symbols). Solid lines represent single scattering albedo and asymmetry parameter values from OPAC dataset for mineral particles.

Table 1

<table>
<thead>
<tr>
<th>Wavelength (μm)</th>
<th>Single scattering albedo</th>
<th>Asymmetry parameter</th>
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<tr>
<td>0.45</td>
<td>0.81</td>
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<td>0.65</td>
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</table>

The CALIPSO vertical profiles of 532 nm extinction coefficient intensity over Évora on (a) 5 April and (b) 7 April.

Fig. 1. The CALIPSO vertical profiles of 532 nm extinction coefficient intensity over Évora on (a) 5 April and (b) 7 April.
Table 2
Daily average of the aerosol optical properties and standard deviations used in the simulations from 4 up to 7 April 2011.

<table>
<thead>
<tr>
<th>Wavelength (nm)</th>
<th>AOD ± (SD)</th>
<th>ω ± (SD)</th>
<th>g ± (SD)</th>
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<tr>
<td>04/04/11 440</td>
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<td>0.72 ± 0.02</td>
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<td>670</td>
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<tr>
<td>870</td>
<td>0.06 ± 0.02</td>
<td>0.90 ± 0.05</td>
<td>0.69 ± 0.02</td>
</tr>
<tr>
<td>1020</td>
<td>0.05 ± 0.02</td>
<td>0.90 ± 0.05</td>
<td>0.70 ± 0.01</td>
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<td>0.88 ± 0.01</td>
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<tr>
<td>07/04/11 440</td>
<td>0.48 ± 0.02</td>
<td>0.91 ± 0.01</td>
<td>0.76 ± 0.01</td>
</tr>
<tr>
<td>670</td>
<td>0.44 ± 0.02</td>
<td>0.97 ± 0.01</td>
<td>0.74 ± 0.01</td>
</tr>
<tr>
<td>870</td>
<td>0.42 ± 0.02</td>
<td>0.98 ± 0.01</td>
<td>0.73 ± 0.01</td>
</tr>
<tr>
<td>1020</td>
<td>0.39 ± 0.02</td>
<td>0.98 ± 0.01</td>
<td>0.74 ± 0.01</td>
</tr>
</tbody>
</table>

(Fig. 4a), the DD plume still did not reach the Portuguese coast, which is corroborated by the back trajectories that show the three levels of air masses arriving from the Atlantic Ocean (Fig. 5a). The next day (5 April) all the three atmospheric levels (500, 1500 and 3000 m a.g.l.) show that the air mass origin was from North Africa continent (Fig. 5b). This situation carry on with very similar pathways for back trajectories on 6 and 7 April indicating a possible mineral dust transport arriving to the Iberian Peninsula and to the measurement site, Évora (Fig. 5c and d).

Despite the previous information, it is necessary to report the intensity of this event with quantitative values. Fig. 6a shows the evolution of the AOD (grey dots) at 440 nm and the Angstrom intensity of this event with quantitative values. Fig. 6a shows the aerosol extinction profiles retrieved at 355 nm from 4 to 7 April between 20 and 21 UTC. The thickness of the aerosol layers was quite variable ranging between −1.5 km (on 4 April) close to the surface and −5 km (on 6 April). On 4 April most of the aerosol load is confined to altitudes below 2.5 km a.g.l., typical of background conditions. Overall, the extinction coefficient values ranged between 0.05 km⁻¹ and 0.2 km⁻¹ during the entire dust episode.

4.2. Radiative effects

4.2.1. Aerosol radiative forcing

ARFs, ARFLw, AHRsw and AHRlw profiles were calculated from the surface to the TOA for every day from 4 up to 7 April. Fig. 6a−d shows instantaneous ARFs evolution with the solar zenith angle (SZA) from the surface up to 7 km altitude. The presence of scattering and absorbing particles as it is well shown by the vertical extinction profiles trapped a large amount of energy in the atmosphere. As we can see in Fig. 8, the vertical aerosol distribution caused a reduction of solar radiation at the surface. In fact, ARFs increased its magnitude (absolute value) for decreasing altitude, reaching the highest values at the surface. ARFs negative values were found for the entire dust episode causing a cooling effect, with the largest value (absolute value) on 6 April (−99 W/m² at the surface for SZA = 60°) (Fig. 8c). Larger aerosol load and single scattering albedo on 6 April than on 4 April decreased the downward SW irradiance reaching the surface. Perrone et al. (2012) found for DD particles lower instantaneous ARFs values ranging from −24.2 to −26.6 W/m² at the surface for SZA between 53.8° and 59.8°. This large difference regarding the highest absolute ARFs value in our study for SZA = 60° may be due to lower aerosol load over Lecce with an AOD (550) value of 0.2. Another important reason may be the cloud layer at 2 km altitude considered in our simulations, which caused together with the aerosol layer a larger shortwave irradiance reduction reaching the surface. On the other hand, di Sarra et al. (2011) during a very intense Saharan dust event, which was detected on 25–26 March 2010 at Lampedusa retrieved AOD (500) values of up to 1.9. This aerosol load caused an ARFsw of −209 ± 10 W/m² for SZA = 35°. In our analysis, the instantaneous ARFs value for the closest SZA of 30° was −89 W/m². The reasons for the difference were mainly the higher aerosol load (AOD values) and single scattering albedo values retrieved in our analysis with respect to those values associated to mineral dust found for Lampedusa. On the other hand, ARFLw showed a variability from +3.4 W/m² for AOD (440) = 0.12 to +21.9 W/m² for AOD (440) = 0.63 at the surface according with Tables 2 and 3. Larger aerosol loads could enhance the feedback between the absorption and re-emission processes increasing the ARFLw more than those associated with moderate and low aerosol load. Instantaneous ARFlw span the 3.2–22.1 W/m² range at the surface, offsetting 14–24% of the ARFs (Fig. 9). The ARFlw results of this work are slightly larger than the ones obtained in other studies related to the Mediterranean basin. For instance, our findings are larger to the ones retrieved by Sicard et al. (2014) ranging between +2.8 and +10.2 W/m² within the 21.1–88.8° SZA range during 11 dust outbreaks observed in Barcelona (Spain) over the period 2007–2012. They also found that the ARFlw represented from 11% to 26% (with opposite sign) of the corresponding ARFs component. Likely the difference is due to the cloud layer presence, which caused larger longwave downward irradiance reaching the surface and larger outgoing irradiance at the TOA than in the clear sky situation. On the other hand, the aerosol layers is a critical point to take into account with respect to the radiative effects in the atmosphere. Gómez-Amo et al. (2014) using different profiles of the
aerosol absorption reported that different shortwave/longwave contributions were retrieved in the boundary layer depending on the aerosol absorption in this layer. Some authors have considered aerosol vertical profiles during dust episodes retrieved from lidar system in order to evaluate SW and LW aerosol radiative forcing at the surface and TOA (Perrone et al., 2012; Meloni et al., 2015; Sicard et al., 2014). They highlighted that vertical distribution has on average a large impact on aerosol heating rates in SW and LW domains. Table 3 summarizes the daily ARFs at the surface and TOA during the four days of intense dust over Évora. As can be appreciated from this table and Fig. 8, changes in the vertical extinction coefficient during these four days produced changes by about ±22 W/m² in daily mean ARFSw at the surface between 4 and 6 April days. According to this table, ARFLw reached the largest value on 6 April (+21.9 W/m²). Rather large (absolute value) ARFSw and ARFLw values were also reached on 7 April (−68.5 and +21.1 W/m², respectively). In fact, 6 and 7 April represent the days most affected by the African dust episode, according to the increase of the contribution of coarse mineral particles as AOD, z and V(r) reveals (Fig. 6a and b). The minor amount of dust loading was mainly responsible for the smaller values of the ARFSw and ARFLw, respectively, retrieved on 4 and 5 April. Overall, on daily average, ARFLw dust effect was large and contributed (~24%) to offset the daily ARFSw effect at the surface during the extreme dust event.
the TOA, aerosols triggered a cooling effect in the SW interval with ARF$_{SW}$ values ranging between $-19.2$ and $-35.1$ W/m$^2$. In the LW spectral range aerosols caused a heating effect with values ranging between $+1.1$ and $+4.3$ W/m$^2$. Perrone et al. (2012) have observed using a radiative transfer model that the ARF$_{LW}$ increases with the increase of the AOD and the coarse/ fine mode ratio in accordance with our findings. Meloni et al. (2015) retrieved daily average ARF$_{SW}$ of $-55.5$ and $-39.0$ W/m$^2$ at the surface and TOA, respectively, during a DD event. These last results agree well with the values obtained in our study. Overall, ARF$_{SW}$ associated to DD particles presented important radiative effects and should be taken into account in climatic studies.

### 4.2.2. Aerosol heating rate

Figs. 10 and 11 show AHR$_{SW}$ and AHR$_{LW}$ vertical profiles for different values of SZA. The daily forcing values trend between 4 and 7 April were translated to the similar trend in daily AHR$_{SW}$ values at the surface (Table 4). The infrared heating between the lower atmosphere and 3 km altitude on 4 April was attributable to the absorption of local thermal radiation emitted from below, the cloud layer at around 2 km and the above aerosol layers. Two AHR$_{SW}$ maximum values were found at around 4 km and 2 km (1.3 K/day) altitudes at SZA = $10^\circ$, while the AHR$_{LW}$ profile reached $-0.7$ K/day between 1 and 3 km altitude on 5 April. For this day the peak of AHR$_{SW}$ profile occurred at a higher altitude than the peak of the AHR$_{LW}$ profile. In the LW interval, the dust and the cloud layer caused a cooling in the dust layer and a heating in the lowest atmospheric layers. The AHR$_{LW}$ contributions were about 50% of the AHR$_{SW}$ between 2 and 4 km altitude for SZA = $10^\circ$. On the other hand, the largest (absolute value) AHR$_{LW}$ ($-0.82$ K/day) occurred at
4 km altitude on 6 April. The $\Delta H_{RLw}$ next to the surface reached positive value (0.3 K/day) caused by the downward irradiance from the cloud layer. The LW heating/cooling was dependent on the scattering and absorption of aerosols and of the cloud layer. The heating below 1 km was mainly due to the absorption of downward LW irradiance from dust layers and cloud layer. The $\Delta H_{RLw}$ profile was affected by the particle extinction profile, with maxima of cooling just above the thickest aerosol layer on 6 April. For this day $\Delta H_{RLw}$ profile showed two maximum values around 1 km (1.53 K/day) and 5 km (0.72 K/day) altitudes. The peak around 5 km was caused by the interaction of solar radiation with the higher dust layer. Comparable values were found in the Mediterranean basin: Barcelona, Spain (Sicard et al., 2014), Roma, Italy (Gómez-Amo et al., 2014), Lecce, Italy (Perrone et al., 2012) and at the island of Lampedusa, Italy (Meloni et al., 2015). Hence, we must highlight that the shape of the vertical aerosol distribution had significant influence on the radiative effects at the surface. These results confirm that dust layers can influence the radiative effects in the LW spectral range.

**Fig. 5.** Back trajectories at 500, 1500 and 3000 m a.g.l. arriving over Évora at 12:00 UTC on a) 4 April, b) 5 April, c) 6 April and d) 7 April.
5. Conclusions

The extreme Saharan dust event, which advected large dust load from North African continent over Portugal on April 4, 2011 and following days was analyzed. There was also some cloud occurrence in the period. AERONET products, CALIPSO profiles, lidar-PAOLI measurements, back trajectory analysis and MODIS images and CERES data, cloud vertical profiles, supported the reasoning of this advection. In this study, aerosol optical and microphysical properties derived from sun-photometer, vertical particle extinction coefficient from the PAOLI system and those aerosols optical properties provide in the OPAC model were used, as well as atmospheric meteorological profiles to initialize a radiative transfer model. In order to optimize the aerosol optical properties, we compared the CERES TOA solar and thermal fluxes against SBDART model simulations along the CALIPSO orbit. The dust aerosol type that fits best both data sets was found to be the mineral accumulation mode. The intercomparison between the instantaneous TOA at shortwave and longwave fluxes obtained using CERES and those calculated using SBDART (CALIPSO profiles or lidar-PAOLI) was reasonably acceptable within the standard deviations. Therefore it is clear from these comparisons that the SBDART model
constrained with the CERES observations can be used to reliably determine aerosol radiative forcing and heating rates with the input of aerosol vertical distributions. SW and LW aerosol radiative forcings and heating rates were calculated with this model considering a cloud-aerosol free atmosphere as reference.

Table 3
Shortwave and longwave daily aerosol radiative forcing at the surface and TOA from 4 up to 7 April 2011.

<table>
<thead>
<tr>
<th>Date</th>
<th>ARF_{sw} (W/m²) surface</th>
<th>ARF_{sw} (W/m²) TOA</th>
<th>ARF_{lw} (W/m²) surface</th>
<th>ARF_{lw} (W/m²) TOA</th>
</tr>
</thead>
<tbody>
<tr>
<td>04/04/11</td>
<td>-63.9</td>
<td>-19.2</td>
<td>+3.4</td>
<td>+1.1</td>
</tr>
<tr>
<td>05/04/11</td>
<td>-56.9</td>
<td>-25.9</td>
<td>+19.1</td>
<td>+3.9</td>
</tr>
<tr>
<td>06/04/11</td>
<td>-85.7</td>
<td>-35.1</td>
<td>+21.9</td>
<td>+4.3</td>
</tr>
<tr>
<td>07/04/11</td>
<td>-68.5</td>
<td>-29.8</td>
<td>+21.2</td>
<td>+4.1</td>
</tr>
</tbody>
</table>

Fig. 9. a-d Modeled vertical profiles of the aerosol radiative forcing for longwave interval from 4 up to 7 April 2011.

Fig. 10. a-d Modeled vertical profiles of the aerosol heating rate for shortwave interval from 4 up to 7 April 2011.

Instantaneous ARF_{sw}, ARF_{lw}, AHR_{sw} and AHR_{lw} profiles underwent significant changes during the four days of the DD event due to the high variability in dust aerosol load and its optical properties, as well as in the aerosol vertical distribution. We found that AOD at 440 nm increased by about a factor of 5 on 6 April, with respect to
the low dust load on 4 April. It was responsible for the ARF$_{SW}$ and ARF$_{LW}$ increase up to $-99$ W/m$^2$ for SZA = 60° and $+21.9$ W/m$^2$, respectively, at the surface. ARF$_{LW}$ showed a variability from $+3.4$ W/m$^2$ for AOD (440) = 0.12 to $+21.9$ W/m$^2$ for AOD (440) = 0.63 at the surface. Hence, larger aerosol loads could enhance responses between the absorption and re-emission processes, increasing the ARF$_{LW}$ more than those associated with moderate and low aerosol load. Instantaneous ARF$_{LW}$ spanned the 3.2–22.1 W/m$^2$ range at the surface offsetting 14–24% of the ARF$_{SW}$. At the TOA, aerosols had always a cooling effect in the SW interval with ARF$_{SW}$ values ranging between $-19.2$ and $-35.1$ W/m$^2$. In the LW spectral range aerosols caused, at the TOA, a heating effect with values ranging between $+1.1$ and $+4.3$ W/m$^2$. ARF$_{SW}$ showed two maximum values around 4 km and 2 km (1.3 K/day) altitudes at SZA = 10°, on 5 April. For that same day ARF$_{SW}$ reached $-0.7$ K/day between 1 and 3 km altitude, in a way that the ARF$_{SW}$ contribution was about 50% of the ARF$_{SW}$ for SZA = 10°. The LW heating/cooling was dependent on the scattering and absorption of aerosols and of the cloud layer. The heating below 1 km was mainly due to the absorption of downward LW irradiance from dust layers and cloud layer. Overall, ARF$_{LW}$ profiles associated to DD particles presented important radiative effects, which were variable with altitude. The unprecedented results derived from this work complement the findings in other regions on the modifications of radiative energy budget by the dust aerosols, which could have relevant influences on the regional climate and will be subject for future investigations.

Acknowledgements

Antonio Valenzuela thanks Universidad de Granada for the award of a postdoctoral grant ("Plan Propio. Programa 8. Con-vocatoria 2014"). The work is co-funded by the European Union through the European Regional Development Fund, included in the COMPETE 2020 (Operational Program Competitiveness and Internationalization) through the ICT project (UID/GEO/04683/2013) with the reference POCI-01-0145-FEDER-007690. This work was also partially funded by the University of Granada through the contract “Plan Propio. Programa 9. Convocatoria 2013”. This work was partially supported by the Andalusia Regional Government through project P12-RNM-2409, by the Spanish Ministry of Economy and Competitiveness through project CGL2013-45410-R and by the European Union’s Horizon 2020 research and innovation programme through project ACTRIS-2 (grant agreement No 654109). The authors thankfully acknowledge the FEDER program for the instrumentation used in this work. The authors also acknowledge Samuel Bárias for maintaining instrumentation used in this work. The authors acknowledge EARLINET for providing aerosol lidar profiles available from the EARLINET webpage. Thanks are due to AERONET/PHOTONS and RIMA networks for the scientific and technical support. We also thank the lidar team of the Évora Geophysics Centre. CIMEL calibration was performed at the EARLINET-EUROPE GOA calibration center, supported by ACTRIS under agreement no. 262254 granted by European Union FP7/2007–2013 and by European Union’s Horizon 2020 research and innovation programme (grant agreement No 654109). The provision of the HYSPLIT model is due to the NOAA Air Resources Laboratory (ARL).

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