APPLICATION OF LIDAR AND SUN PHOTOMETER MEASUREMENTS TO AEROSOL RADIATIVE FORCING CALCULATIONS

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Abstract

Solar and infrared instantaneous aerosol forcing are calculated exploiting sun photometer and lidar data in an EARLINET – AERONET site. It is shown that, for the studied cases, the actual vertical profile has an important role mainly for the top-of-the-atmosphere infrared forcing that can give a not negligible effect on the total forcing.

1. Introduction

Aerosols play an important role in the energy budget of the Earth-Atmosphere system, but the quantitative assessment of this budget is still uncertain because of the complex dynamics and short lifetime of aerosols. Many progresses have been accomplished through different global observation systems, both ground and satellite based. Two important examples are the sun photometer network AERONET and the MODIS spectrometer on board of AQUA and TERRA satellites. From such passive instruments, a global coverage of the optical properties of aerosols can be obtained. Thus, different calculations of aerosol radiative forcing, i.e. the change in the radiative budget of Earth and atmosphere, has been performed. However, the data obtained by passive instruments have the important limitation that vertical properties of aerosol cannot be taken into account. To avoid this limitation it is interesting to exploit, when possible, information on vertical distribution of aerosol. Apart from cases like instruments on board of airplanes or tethered balloons, the only practical way to do this in the long run is the use of a lidar. The establishment of different lidar network like EARLINET (Bösenberg et al., 2003) makes available systematic measurements of aerosol vertical profiles and there are actually many lidar stations that are also equipped with automatic sun photometers.

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Thus, it is important to investigate the impact of aerosol vertical distribution on both the solar and the infrared radiative forcing. In this paper we use contemporary sun photometer and lidar measurements to study the variation of the solar and the infrared aerosol forcing along the day in two different aerosol load situations. We will discuss a case of Saharan dust transport, in which aerosol properties change noticeably during the day, and a second case in which aerosol properties are stationary along the day.

2. Experimental Apparatus and Data

The measurements site is located at the Department of Physics, University of Lecce, Italy (40.33º N, 18.10º E). The environment is semi-rural and it is far from big anthropogenic aerosol source. The main characteristic of the site is that it is located in the narrow Salentum peninsula. The width of the peninsula at the measurement site is less than 40 km and the site is located approximatively in the center.

The experimental data used for the presented radiative forcing calculation are :

i. Backscatter coefficients obtained by a lidar operating at 351 nm (XeF excimer laser)
ii. Aerosol data at 4 wavelengths (440, 670, 870, and 1020 nm): optical thickness (AOT), size distribution, complex refractive index obtained by a CIMEL sun photometer.
iii. Meteorological radiosoundings from the station of Brindisi, located 40 km away from the measurement site, every 6 hours.
iv. Surface albedo values retrieved by the MODIS instruments on board of the Terra satellite.

The lidar system has been described in De Tomasi and Perrone (2003). Although it is a Raman lidar, such possibility is available only at nighttimes. Since in this paper we are interested to the diurnal forcing, only elastic backscattering measurements will be used. The used profiles are typically 50 min averaged to reduce the influence of solar background. From the elastic return profiles, an extinction profile has been obtained using the Klett-Fernald method with a constant extinction to backscatter ratio (lidar ratio) chosen to reproduce the observed sun photometer AOT at 380 nm. The lidar system is part of the European network EARLINET.

The CIMEL photometer is an automated system and it is part of the global network AERONET (Holben et al., 1998). The data used in this paper are cloud-screened AERONET products (Level 1.5) and they are downloadable from the site http://aeronet.gsfc.nasa.gov/.

Radiosoundings give atmospheric pressure, temperature and relative humidity at 5, 11, 17, 23 UTC, all the days. These data can be downloaded (http://raob.fsl.noaa.gov) and they are important for calculation of molecular scattering, water vapor absorption and aerosol IR emissivity.

Finally, the solar surface albedo used in this study is based on MODIS satellite sensor data (the filled land surface albedo product from MOD43B3) in the 0.3-0.7 µm (for the visible) and 0.7 – 5 µm (for the near-infrared) spectral ranges.

3. Radiative Forcing Model

In this paper, we examine the instantaneous aerosol radiative forcing in the solar (0.35 – 4 µm) and in the infrared (4 – 80 µm) spectral ranges to study the effect of different aerosol types on the radiative energy balance of the Earth-Atmosphere-System.

The aerosol radiative forcing has been calculated as the change of the net radiative fluxes at the Top of Atmosphere (ToA) and at the surface (sfc), when the aerosol are present with respect to the case without aerosols.

The radiative transfer model used in this study is based on a two-stream method (Meador and Weaver, 1980). This necessitates repeated applications (ca 150 times) to properly approximate the spectral variability of atmospheric properties of particles (via 8 solar and 12 infrared spectral sub-bands) and of major trace-gases (Ozone, CO_2, CO, N_2O, and CH_4 - through a number of exponential terms in each of the sub-bands). The two stream method provides the radiative fluxes at the boundaries of homogeneous plane-parallel layers, which were chosen to approximate a vertically inhomogeneous atmosphere. With a focus on tropospheric aerosols, half of the selected 20 layers are below 5 km altitude.

To determine the aerosol radiative forcing, the model needs informations on the vertical profiles of the main meteorological parameters, and the concentration vertical profiles of the main atmospheric gases to take in account the Rayleigh scattering and ozone absorption. Besides these informations the local vertical distribution and also the local microphysical and optical properties of aerosols are required. Finally, regional data on the solar surface albedo is needed for the calculation of the radiative forcing.

Radiosoundings are used to retrieve the vertical profiles up to 23 km of atmospheric density, pressure, temperature and water vapor content, at the different hours of the day. Above the 23 km altitude we use the standard, mid-latitudes (30º – 60º N), vertical profiles of density, pressure, temperature, and water vapour provided by the US Air Force Geophysics Laboratory (AFGL) for summer months, as July is. The concentration vertical profiles of ozone, N2O, CO and CH4 provided by AFGL, are also used to take in account the Rayleigh scattering and ozone absorption.

The infrared ground emissivity has been set to 96%.

Aerosol volume size distributions and real and imaginary refractive indices are used to characterize aerosol properties. Then, MIE calculations (assuming a spherical particle shape) are applied to translate the data on size, concentration and refractive indices into optical depths (a measure for amount), single scattering albedos (a measure for absorption) and asymmetry-factors (capturing the scattering behaviour).

The sun photometer data are not vertically resolved, so they represent a column average; thus, the only vertical resolved property of aerosol is their extinction coefficient: all the other parameters assumed constant, it is proportional to the total concentration of particles. Since SSA and asymmetry factor are intensive parameters, the only aerosol property varying vertically is the optical thickness of the different atmospheric layers.

4. Case Study 1: Saharan Dust Transport event

4.1 Sunphotometer and Lidar Measurements

We will present in this section the radiative forcing calculation in two days, 18 and 21 July 2005 in which a significative layering of the aerosol load is present. These two cases are discussed in detail in Tafuro et al. (2006), but the main features will be repeated here for completeness.

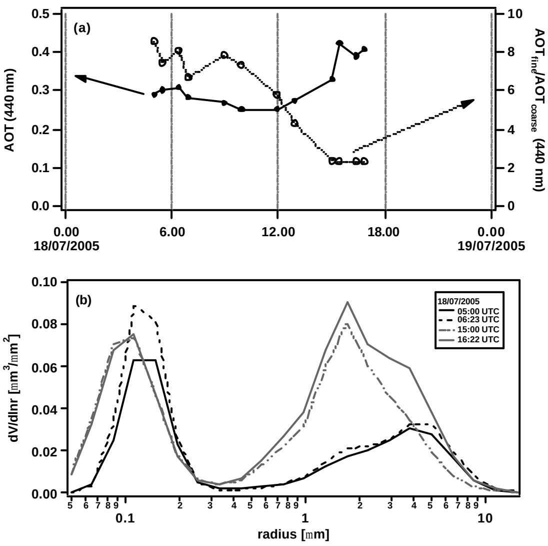

Fig. 1. (a) AOT and AOT_{fine}/AOT_{coarse} at 440 nm evolution during the 18^{th} July, 2005; (b) volume size distributions retrieved by sun photometer measurements of 18^{th} July, 2005.

The 18 July case is an intrusion of Saharan dust that results in an high aerosol layer which is superposed on a preexisting lower altitude layer; it is interesting because it shows the effect of the change in aerosol properties on radiative forcing. The backtrajectories analysis from AERONET project (at 00:00 UTC and 12:00 UTC of 19 July) shows that dust has been transported to the lidar site above 2 km altitude. From Fig. 1a we can see that sun photometer measurements confirm the change of aerosol properties along the day. Fig. 1a shows the aerosol optical thickness (AOT) at 440 nm that increases after 12 UTC. Figure 1a shows also the evolution of the ratio between the AOT due to fine mode (radius smaller than 0.6 µm) particles (AOT_{fine}) and the AOT due to coarse mode (radius larger than 0.6 µm) particles (AOT_{coarse}). The AOT_{fine}/ AOT_{coarse} ratio indicates that the contribution of fine mode particles decreases on the afternoon of July 18, related to the arrival of Sahara dust particles, in accordance to backtrajetories, satellite images, and lidar measurements. These two features indicate that the increasing aerosol load changes its population size distribution increasing its coarse mode population. This is visible explicitly in Fig. 1b where we report the volume size distributions as calculated by AERONET inversion techniques in some selected times to evidence the variations. In particular, one can observe from Fig. 1b that the fine mode particles dominate in the morning of July, 18, while, a significant increase of the coarse mode particles is observed from the post-meridian volume size distributions.

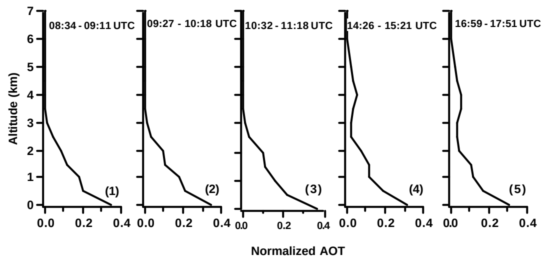

Fig. 2. Aerosol vertical profiles retrieved by lidar measurements along the 18^{th} July, at different hours.

In Fig. 2 we report the fraction of the total AOT, obtained by lidar measurements, corresponding to the layers in which the atmosphere is decomposed in the model calculations. We can see that, for measurements after 12:00 UTC a new layer of aerosols is present above 2.5 km altitude. Since the structure of the vertical distribution of aerosol does not change very much at lower altitude, we can deduce that most of the changes in the detected properties of the aerosols are due to the arrival of Saharan dust.

4.2 Aerosol radiative forcing

Aerosol radiative forcing, both in the solar region (0.35-4 µm) and the infrared region (4-80 µm) has been calculated at different times along the day, when sunphotometer data corresponding to sky almucantar measurements are available. We use for these calculations two values of the surface albedo in the solar region (Tab. 1).

Tab. 1. Solar surface albedo values chosen for Lecce

	Solar Surface Albedo	
	VIS (0.3 – 0.7 µm)	VIS (0.7 – 5 µm)
MODIS	0.07	0.16
Zhou et al., 2003	0.18	0.36

The first is the albedo retrieved from MODIS measurements corresponding to a region of about 1°x1° centered on the lidar site; the other one is an approximation for a barren-sparse vegetated soil (Zhou et al., 2003). The two values are different because of the peninsular nature of the measurement site, so that the measured surface albedo is an average between ocean and land values. The TOA and SFC solar radiative forcing are shown in Fig. 3; the calculation at the two different surface albedos appears shifted of a constant value of about 10 W/m2, otherwise they share a similar behavior. Both the SFC and the TOA are in most of cases negative and in absolute value the solar forcing reach a minimum that is situated before local noon (11 UTC) and after local noon respectively. In stationary conditions, it is expected that the negative radiative forcing follows a diurnal cycle with a local minimum at minimum zenithal angle, a maximum at intermediate angles zenithal angle, and a minimum at high zenithal angle (Meloni et al., 2005); in this case this behavior is masked by the variations in AOT and the effect of the arrival of Saharan dust is visible in the absolute increase of the forcing in the later hours as compared to the earlier hours of the day.

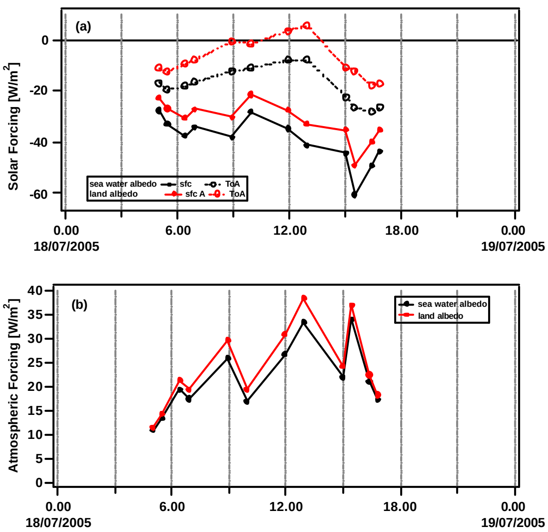

Fig. 3. Solar forcing on 18 July 2005. Black traces refer to a surface albedo retrieved from MODIS observations and red traces to fixed values for land. (a) Solar forcing at the top of the atmosphere (open symbols) and at the surface (full symbols). (b) Atmospheric forcing

The two forcings are negative and the surface forcings is larger in absolute value; these means that the atmospheric forcing, which represents the variation of radiative energy stocked in the atmosphere, is positive (Fig. 3b). Thus, absorbing aerosols will produce a heating of the atmosphere. The atmospheric forcing has an irregular behavior that is due to the fact that the surface forcing is anticorrelated with the optical thickness, but the TOA forcing has a smoother variation. However, we can detect a trend toward increasing value of the atmospheric forcing.

The infrared forcing has also an important role for the total climate aerosol impact. It has been shown that it can be a relevant fraction of the total solar forcing (Markowicz et al., 2003). The emissivity properties of the ground and the gas absorption are in this case very important.

IR forcing is positive both at the surface and at the top of the atmosphere (Fig. 4). At the surface, this is mainly due to the increase of the downward IR flux due to the IR emission of the aerosols, while at the TOA the positive forcing is due to the reduction of the upward flux. The IR radiative forcings at the top of the atmosphere are very small in comparison with the corresponding solar values, however they can be significative in some cases.

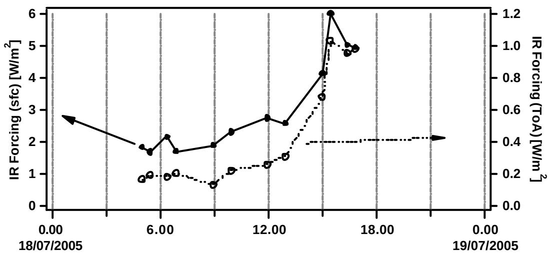

Fig. 4. Infrared top of the atmosphere (open symbols) and surface forcing on 18 July 2005.

The diurnal variation of the IR aerosol forcing shows a huge increase with the corresponding arrival of the Saharan dust particles. This is probably due to the arrival of larger particle that contribute more to the scattering in the IR region of the spectrum. Though the IR forcing is much less than the solar forcing, it could actually be important in the global energy budget because the solar forcing calculations actually are quite sensitive to the value assumed for the surface albedo. In fact, if the solar forcing is calculated with the barren-sparse vegetated albedo, its order of magnitude at the TOA is the same than the IR forcing, and it will be important for the calculation of the absolute value and the sign of the total forcing (solar + IR).

4.3 Sensitivity to aerosol profiles

The sensitivity of the forcing calculations to the aerosol load profile obtained by lidar has been tested using, for the different sun photometer measurements two fixed linear profiles (number 2 and 4 in Fig. 2), corresponding to the absence and presence of Saharan dust between 3 and 5 km. The percentage variations are reported relatively to the profile 2 calculations.

The solar forcing is not very sensitive to the used profile: the variation of the forcing ranges between -1.2 and 0.5 % for the SFC and between -6.6 % and 2.2 % for the TOA. Accordingly to the results of Meloni et al. (2005), this low sensitivity is expected because the observed SSA shows that the particles are not very absorbing.

Also the SFC IR forcing is not very sensitive to the profile: relative variations are almost constant and ranges between -4.5 % to -5.1 %. The larger sensitivity is obtained with TOA IR forcing that shows a minimum relative variation of 61 % and a maximum of 110 %. This important variation on the positive forcing show that aerosols distributed on a longer column decrease the IR upward flux, as it can be expected because of the decreasing in temperature. Thus, a precise calculation of the total solar forcing at the TOA requires the actual profiles of aerosols, in particular for the case in which the land albedo is used because the TOA solar forcing can have the same order of magnitude of TOA IR forcing.

5. Case Study 2: Long Range transport event

5.1. Aeronet and Lidar Measurements

On 21 July 2005, the Saharan dust transport event of 18 July is terminated, as t can be seen from backtrajectories and satellite images.

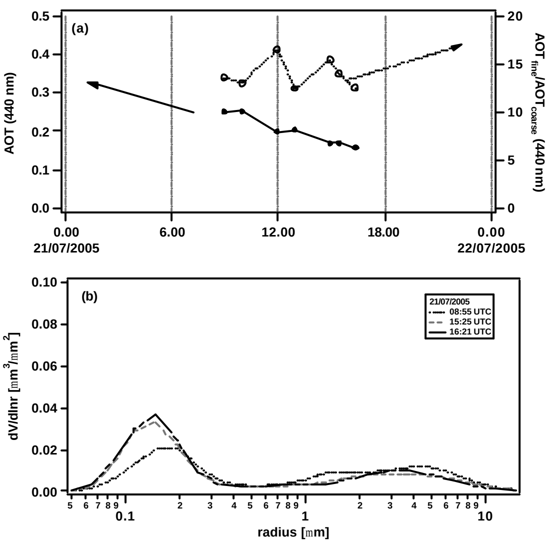

Fig. 5. a) AOT and AOT_{fine}/AOT_{coarse} at 440 nm evolution during the 21st July, 2005; (b) volume size distributions retrieved by sun photometer measurements of 21st July, 2005.

In fact, north-west Europe was the source region of the 950 and 850 hPa air masses that have reached Lecce on July 21, while the Atlantic Ocean was the source region of the 700 hPa air mass.

Sun photometer measurements reported in Fig. 5 show that the the fine fraction is, on average, larger than that observed on 18 July before the arrival of the dust. The Angstrom coefficient is on average 1.8, which shows that the aerosol load is composed mainly of fine mode particles. The AOT at 440 nm slightly decreases and it has an average value of 0.2, lower than that observed on July 18. The single scattering albedo has a maximum around local noon and ranges between 0.88 and 0.94 at 440 nm, indicating slightly absorbing particles. The lidar profiles (Fig. 6) show that aerosol are concentrated in the lower 3 km, with a stable layer between 2 and 3 km.

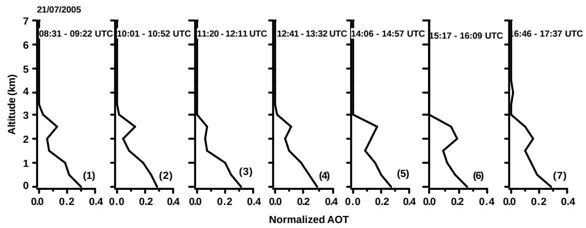

Fig. 6. Aerosol vertical profiles retrieved by lidar measurements along the 21st July, at different hours.

5.2. Aerosol radiative forcing

The solar aerosol radiative forcing has been calculated following the same protocol of section 4. In the case of the MODIS retrieved surface albedo the forcing is always negative; at the surface, its absolute value decrease is correlated to the decrease of the AOT. The TOA forcing instead slightly increases along the day (Fig. 7). As a result, the atmospheric forcing decreases regularly from 25 W/m2 to 10 W/m2.

The IR surface forcing maintains between 1 and 2 W/m2 at the surface and between 0.1 and 0.2 w/m2 at the top of the atmosphere.

The sensitivity to the aerosol profile, also in this case, has been tested using two different profiles, corresponding to 3 in Fig. 6 and 4 in Fig.2. The main differences, also in this case, are that aerosols extend until 3 and 5 km, respectively. If we define, as before, the percentage variation relative to the 3 km profile, we have, for solar forcing, a small variation at the surface (-1.7 – 1%) and at the top of the atmosphere (-12 – 2%).

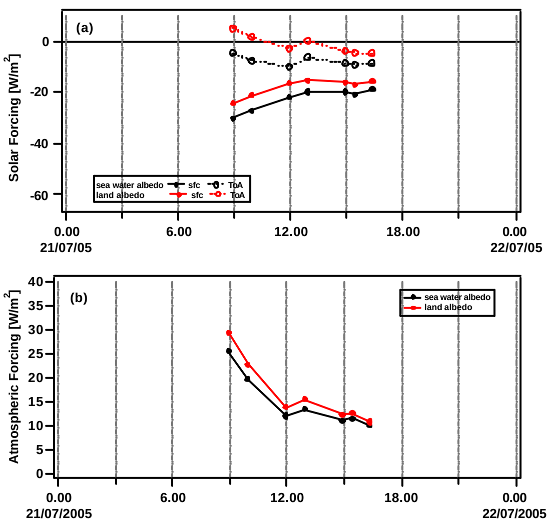

Fig. 7. . Solar forcing on 21 July 2005. Black traces refer to a surface albedo retrieved from MODIS observations and red traces to fixed values for land. (a) Solar forcing at the top of the atmosphere (open symbols) and at the surface (full symbols). (b) Atmospheric forcing

For infrared forcing we have a small negative variation at the surface (-4.8 - -6.5%) and, as in the previous case, a huge positive variation (60 -133%) at the top of the atmosphere.

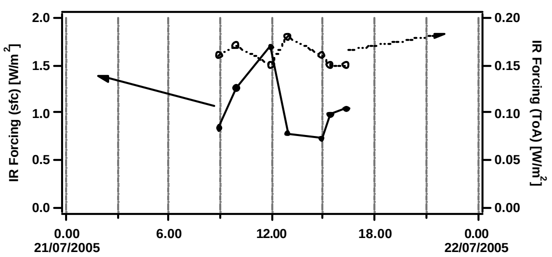

Fig. 8. Fig. 9. Infrared top of the atmosphere (open symbols) and surface forcing on 21 July 2005.

6. Conclusions

In this paper we have calculated the instantaneous solar and infrared aerosol radiative forcings in two days corresponding respectively to an intrusion of Saharan dust, which changes noticeably the aerosol optical parameters, and stationary conditions. In both cases, we get a surface solar negative forcing that is scarcely sensitive to the aerosol profile. The top-of-the-atmosphere forcing is in general larger (less negative) than surface forcing, giving a positive atmospheric forcing. The absolute value of forcing is very sensitive to surface albedo. In particular, we have used, for solar radiation, values obtained by MODIS measurements and a fixed value corresponding to land, because the measurement site is located in a narrow peninsula. In this last case, the TOA solar forcing can change sign, so the infrared forcing, even if in general small compared to the solar forcing, can have a non negligible effect. In particular, the arrival of Saharan dust doubles the TOA IR forcing. Independently of the nature of the aerosol particle, the TOA IR forcing is very sensitive to the aerosol profile obtained by lidar, mainly because of the changes in aerosol IR emission due to changes in temperature. As a consequence, the determination of this profile becomes important for the determination of the total (solar + IR) forcing.

Acknowledgements

This work has been partially supported by the European Commission through the project EARLINET-ASOS (2006-2010 contract n. 025991).

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