Characterization of the vertical structure of Saharan dust export to the Mediterranean basin

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Abstract. We present the results of our investigations into the vertical structure of several North African dust plumes exported to the Mediterranean in 1997. Two backscatter lidar systems were operated in the western and eastern parts of the Mediterranean basin during dust events identified using Meteosat visible images. Dust transport soundings have shown that dust particles are trapped and transported inside well-defined layers in the free troposphere. In general, the dust transport appeared to be multilayered, with several distinct layers at different altitudes between 1.5 and 5 km. The analysis of Meteosat IR images, the Total Ozone Mapping Spectrometer aerosol index, and back-trajectories clearly shows that these layers have different origins in Africa. Finally, in addition to the free troposphere transport, the presence of dust particles inside the planetary boundary layer has been assessed and quantified for two particular events with aerosol optical thickness of 0.3-0.4, using simultaneous lidar and Sun photometer measurements. In one case only, significant dust load (dust optical thickness of ~0.1) occurred in the boundary layer.

1. Introduction

The scientific community has made great efforts to document and understand the interactions of mineral aerosols (dust) with the environment. We currently know that these are numerous and concern key phenomena such as Earth's radiative budget [Legrand et al., 1992; Tegen and Fung, 1995]. Studies have covered a large domain from the particles' chemical and optical properties to the dynamical processes responsible for their uptake and transport. However, there is still a lack of observations, which makes the quantification of these impacts uncertain.

Regarding the radiative impact of dust particles, an important lacking observation is the vertical structure of dust clouds. Indeed, to a small extent in the short wave domain and to a greater extent in the long wave domain, radiative forcing calculations need to consider the vertical repartition of the particles [Ackerman and Chung, 1992; Chazette et al., 1998]. The latter is transport dependent. Indeed, during the transport the vertical structure may be modified by mechanisms such as convective erosion of the dust layer and gravitational settling of particles. Prospero and Carlson [1972] have documented such changes in the vertical structure for the African dust transport over the Atlantic. The latter is known to occur inside a well-defined structure called the Saharan air layer. This structure is created by the wind shear that occurs when the emerging warm dust-loaded Saharan air mass pushed by trade winds is undercut by the cool northerly coastal winds.

The Mediterranean transport it is very different from the Atlantic one. Indeed, the average atmospheric circulation over the Mediterranean does not explain the northward transport of dust particles because winds mostly come from the west [La Fontaine et al., 1990]. Different authors have reported on the processes that govern such a transport. The compilation of their work has led to three main scenarios [Moulia et al., 1998]. They all have drawn the conclusion that Mediterranean dust outbreaks were related to the cyclone activity inside and around the basin [Bergametti et al., 1989; Alpert and Ziv, 1989]. Indeed, their associated frontal zones have been shown to be able to mobilize and transport great amounts of dust over large distances [Prodi and Fea, 1979; Reiff et al., 1986; Bergametti et al., 1989; Dulac et al., 1992a; Franzén et al., 1994]. Thus the transport of dust to the Mediterranean basin will be driven by more complex wind fields than is the tropical Atlantic transport, and its vertical structure may reflect that complexity. Actual knowledge is based on three-dimensional (3-D) air mass trajectories of dust transport over the Mediterranean. These have indicated synoptic upward movement of the dry and turbid African air masses over the basin [e.g. Bergametti et al., 1989; Dulac et al., 1992a], which suggests that dust particles are primarily transported over the boundary layer. Moreover, Dulac et al.[1996] have estimated an average altitude of the dust layer of 2300 m from dust winds derived from Meteosat during a summer dust transport event in the western Mediterranean region. However, unlike the vertical structure of the tropical Atlantic transport of dust, to our knowledge, no corresponding direct documentation is available for the Mediterranean.

We present the results of the vertical sounding of several Saharan dust outbreaks to the Mediterranean basin using two backscatter lidar systems operated at two different sites and complemented by coincident local and regional scale remote
sensing observations from a Sun photometer, Meteosat, and, when available, the Total Ozone Mapping Spectrometer (TOMS) UV-absorbing aerosol index [Herman et al., 1997].

The presentation of the work is organized as follows. In section 2, information is given on the technical characteristics of the instruments that have been used. The procedures used for retrieving the particles' optical properties are outlined, and finally, we discuss and quantify the associated uncertainties. Section 3 is devoted to the data analysis. It is composed of three subsections, devoted to the analysis and interpretation of the results obtained with each of our remote sensing devices. Section 4 is a discussion of the synergy between the lidar and the Sun photometer that can be used to separate the dust particles' contribution to the total optical thickness from that of the measurement site background particles. Finally, in section 5 we present our conclusions.

2. Observation means

In the framework of the European Mediterranean Dust Experiment (MEDUSE) we have been making from April 1996 to the end of June 1997 a daily local and regional scale survey of African dust exports to the Mediterranean basin. MEDUSE has been set up between several European laboratories in order to develop and implement a prototype system for forecasting and monitoring Saharan dust outbreaks over the Mediterranean [Soderman and Dulac, 1998] and to build a database of regional and local scale observations, including both optical and chemical measurements.

Four different and complementary remote sensing devices have been used in order to get a comprehensive characterization of North African dust exports to the Mediterranean basin: visible and infrared Meteosat imagers, TOMS ultraviolet radiometer, ground-based backscattering lidars, and Sun photometer. On one hand, spaceborne radiometers have given access to synoptic information such as the spatial distribution of the dust cloud and its displacement. For this we have used Meteosat visible images—derived dust optical thickness distribution over seawater [Moulin et al., 1997b, c] and both Meteosat infrared images [Legrand et al., 1994] and the TOMS aerosol index [Herman et al., 1997] over land. On the other hand, backscatter lidar systems have given access to the vertical structure of the dust transport so that finally, it was characterized in all three dimensions.

We have been in charge of optical measurements performed in Thessaloniki (22.967°E, 40.52°N) and Observatoire de Haute Provence (OHP, 5°E, 44°N) with the Laboratory of Atmospheric Physics (LAP, Greece) and the Service d'Études (SA, France), respectively. Both sites were equipped with backscatter lidar systems, but in Thessaloniki a Sun photometer was also operated. Thus this site offered the possibility of using both instruments in synergy to get a more precise characterization of the dust cloud optical properties. Indeed, Thessaloniki is a rather polluted coastal city (the second biggest industrial and urban city in Greece). Thus its aerosol background (i.e., when no dust transport occurs) is characterized by a large optical thickness and composed of small particles from the gas-to-particle conversion (carbonates and sulphates aerosols) [Levon et al., 1998]. Lidar measurements were performed under so-called “dust alerts,” issued by the Mediterranean Research Centre, Erice (Italy), when according to the MEDUSE dust transport model [Nickovic and Dobricic, 1996] and Meteosat daily observation [Dulac et al., 1997], dust-loaded air masses were likely to be advected above the measurement sites.

2.1. Local Scale Observations

2.1.1. Lidar measurements. The main drawback of ground-based and spaceborne passive instruments is that they can only provide vertically integrated information. Active instruments such as the lidar enable a time-resolved investigation of the atmosphere from which the position of the backscattering layers can be retrieved.

For both lidar stations, measurements started in April 1996 and ended in summer 1997. The number of days of observation during dust alerts for Thessaloniki was 24 in 1996 and 12 in 1997. For OHP it was 3 in 1996 and 6 in 1997. This rather low number of measurements is primarily due to a decreased dust transport activity inside the Mediterranean basin in 1996 and 1997, in relation to the North Atlantic Oscillation [Moulin et al., 1997a]. Moreover, dust transports to the Mediterranean being associated with meteorological fronts, dust layers are often found over water clouds, which makes lidar soundings inefficient in characterizing the dust particle properties. Thus, in low-cloud conditions, no lidar soundings were performed.

The backscatter lidar systems used in MEDUSE are based on a Neodymium: Yttrium/aluminum/garnet (Nd:Yag) laser source emitting at 532 nm. The emission is coaxial with the receiver system in order to have a complete overlap between the laser source and the telescope field of view at a range of several hundred meters: the first 600 m are not sounded for Thessaloniki, and the first 1.5 km are not sounded for OHP. The analogic acquisition mode was used with an associated vertical resolution of 15 m. The signal was then filtered out to enhance the signal-to-noise ratio using a binomial filter [Darcehand Lamer, 1983] which degraded the vertical resolution to ~70 m. Aerosol extinction profiles were retrieved using Klein's [1981] algorithm. This procedure requires the knowledge of the aerosol backscattering phase function as well as an upper boundary condition on the extinction coefficient. Chazette et al. [1995] have discussed the uncertainties associated with the retrieval of the extinction coefficient. They identified three main causes: (1) the determination of the boundary condition, (2) the a priori fixing of the aerosol backscattering phase function, and (3) the detection noise. We have set the boundary condition in the free troposphere where the aerosol's contribution is thought to be negligible and used the low-troposphere background aerosol extinction used in the computer code LOWTRAN 6 as a reference value [Kneizys et al., 1983]. The aerosol backscattering phase function was calculated using Mie's theory for the dust aerosol model used in Meteosat analyses [Moulin et al., 1997b]. Its value is 0.035 sr⁻¹. Finally, in our case the detection is performed with a signal-to-noise ratio >20, so the detection noise can be neglected. The resultant uncertainty is 20% for the extinction coefficient. The optical thickness was calculated using the extinction coefficient profiles with an associated relative standard deviation of 15%.

2.1.2. Sun photometer measurements. Thessaloniki site also operated a Sun photometer to measure the vertically integrated aerosol optical thickness. The automated Sun photometer developed by CIMEL® Electronique has been used to measure the atmospheric attenuation of the solar light [Holben et al., 1999]. Measurements were performed at three different
wavelengths ranging from visible to near infrared (440, 670, and 870 nm). Provided that there is proper calibration and that well-known corrections for the contribution of absorption and scattering by air molecules are made to the total attenuation, the Sun photometer allows direct measurement of the total atmospheric aerosol optical thickness at the considered wavelengths $\lambda$ (AOT$_\lambda$). In such retrieval the major source of uncertainty is calibration error, which is directly proportional to the associated uncertainty of the AOT$_\lambda$. Indeed, the relative error $\varepsilon$ on the AOT$_\lambda$ derived from the Beer-Lambert law [Lenoble, 1993] is simply given by:

$$\varepsilon = \frac{\Delta \text{AOT}_\lambda}{m \text{ AOT}_\lambda}$$  \hspace{1cm} (1)

where $\Delta \text{AOT}_\lambda$ is the relative uncertainty of the calibration coefficient and $m$ is the air mass.

In order to minimize $\varepsilon$, we have used two different calibration procedures. In a first step the well-known Langley-Bouguer [Lenoble, 1993] method was used to select the best days for calibrating. These days obeyed four criteria: (1) The optical thickness is fairly constant; all points with an AOT$_\lambda$ differing from the daily mean by more than one standard deviation were discarded. (2) The number of remaining measurements after cloud decontamination during the day is high enough (>80). (3) The dynamic on the air mass range is large enough (from <2 to >5). (4) The optical thickness is weak (<0.3). The three first criteria ensure that the extrapolation of the measurements to air mass 0, requested for calibration, will be as correct as possible. The fourth criterion ensures a minimum sensitivity of the calibration coefficients to variations of the optical thickness. Eight days were selected. The calibration coefficients were then retrieved using the procedure proposed by Herman et al. [1981], which minimizes the associated uncertainties by giving stronger statistical weights to measurements performed at small air masses. This is an improvement of the Langley-Bouguer method, which weights all measurements equally. Finally, all these coefficients were normalized to the same day with a seasonal model of the solar irradiation so that we could take the mean value as the final calibration coefficient. The resulting relative uncertainty of the calibration ($\Delta \text{AOT}_\lambda/\text{AOT}_\lambda$), which is simply calculated by dividing the standard deviation of the calibration coefficients by their mean value, is $\sim$3.9% at 443 nm and $<2.5\%$ at both 670 and 870 nm. In Figure 1 we have plotted the associated relative error of the retrieval of the optical thickness at 443 and 670 nm as a function of the air mass for both the smallest and the biggest optical thicknesses found between May 8 and June 2, 1997, which are two days of dust transport. These represent extreme error functions between which the error fluctuates during these days. Thus we can see that at the uppermost the relative error of the optical thickness is 11% for small air masses. However, for most measurements performed on the May 8 and June 2, 1997, air masses would be >2 so that, as can be seen in Figure 1, the relative error would not exceed 6% at 443 nm or 4% at 670 nm.

Concerning all our Sun photometer measurements, we have rejected here all measurements for air masses <2 or leading to AOT <0.1. Thus the maximum relative error of the AOT$_{443}$ is at uppermost ~20%, but in most cases it will be <10%.

From the optical thickness at $\lambda_1 = 443$ nm and $\lambda_2 = 670$ nm one can derive the Ångström exponent $\alpha$ which describes the spectral dependence of the AOT$_\lambda$ in the visible domain [Ångström, 1964]:

$$\alpha = \ln \left( \frac{\text{AOT}_{\lambda_2}}{\text{AOT}_{\lambda_1}} \right) / \ln \left( \frac{\lambda_2}{\lambda_1} \right)$$  \hspace{1cm} (2)

Uncertainty of the optical thickness will evidently result in uncertainty of $\alpha$. The absolute error $\Delta \alpha$ of $\alpha$ can be expressed on the first order as

$$\Delta \alpha = \sqrt{\frac{\varepsilon_{\text{AOT}_{\lambda_1}}^2}{\lambda_1^2} + \frac{\varepsilon_{\text{AOT}_{\lambda_2}}^2}{\lambda_2^2}} \ln \left( \frac{\lambda_2}{\lambda_1} \right)$$  \hspace{1cm} (3)

The Ångström exponent value depends mostly on the aerosol size distribution. Small or even negative values are found for the large particles such as sea salts or mineral particles, whereas values between 1 and 2 are found for submicron size particles such as sulphates or carbonaceous aerosols [Liousse et al., 1995;
Plate 1. (a and b) Meteosat-derived dust optical thickness at 550 nm for May 8 and 9, 1997, respectively (pink circles show the measurement site location); (c and d) Sun photometer-derived optical thickness at 532 nm and Angström exponent for the same dates; (e and f) extinction coefficient profiles at 532 nm derived from lidar measurements for the same dates (A.M. profiles are performed between 0930 and 1100 UT whereas P.M. profiles are performed between 1700 and 2000 UT).
the obscuring of the aerosol layer by clouds. The procedure used to discriminate between them. However, the amount of aerosol detected in a cloudy region will be reduced because of aerosol and clouds is significantly different, so that it can be detected. Secondly, the spectral signature of UV-absorbing radiation for the detection of absorbing aerosols above land is very small, allowing the aerosol signal to be convenient in two main respects: First of all the UV contribution to the aerosol index [Tremaine et al., 1997]. The use of UV data from the TOMS UV-absorbing aerosol index model [Herman et al., 1997] allows the aerosol load over Thessaloniki to be assessed on a synoptic scale from measurement of AOT550. The relative error of AOT550, εr, can be expressed as a function of the relative errors of AOTx and AOTy as follows:

\[
\epsilon_{r} = \left(1 + \frac{\ln(\lambda)}{\ln(\lambda_x)}\right) \epsilon_{\lambda} - \left(1 + \frac{\ln(\lambda)}{\ln(\lambda_y)}\right) \epsilon_{\lambda_y}
\]

(4)

The value of εr does not exceed 8% for either May 8 or June 2, 1997.

2.2. Synoptic Scale Observations (Meteosat)

Meteosat visible images are used to monitor dust plumes and to assess the intensity of dust outbreaks in terms of optical thickness at 550 nm on a synoptic scale. The procedure described by Moulin et al. [1997b] is based on the use of the radiative transfer code SS [Taut et al., 1990]. It uses a dust type aerosol model, made of the size distribution of the background desert aerosol model of Shuttle [1984] and with a refractive index of 1.50 - 0.01i [Moulin et al., 1997b]. The uncertainty of the retrieved optical thickness has been assessed over the Atlantic by comparing Sun photometer measurements to Meteosat retrieval [Moulin et al., 1997c]. It was shown to be ~25%.

The presence of dust particles over the African continent can also be inferred from Meteosat infrared images. Indeed, because of its altitude the airborne dust radiates at a cooler temperature than the surface, inducing a characteristic temperature brightness contrast [Legrand et al., 1994]. Here we use this property to highlight the likely dust source regions with a procedure similar to that described earlier by Legrand et al. [1994]. Unfortunately, this method also highlights clouds. However, since their cooling effect is usually much stronger than that of dust particles, we are able to make a crude distinction between both dust particles and clouds by applying a simple threshold on the pixel contrast intensities.

2.3. Exogenous Information

In order to somehow back up and complement Meteosat infrared information we have been using the TOMS UV-absorbing aerosol index [Herman et al., 1997]. The use of UV radiation for the detection of absorbing aerosols above land is convenient in two main respects: First of all the UV contribution of the surface is very small, allowing the aerosol signal to be detected. Secondly, the spectral signature of UV-absorbing aerosol and clouds is significantly different, so that it can be used to discriminate between them. However, the amount of aerosol detected in a cloudy region will be reduced because of the obscuring of the aerosol layer by clouds. The procedure associated with the retrieval of the aerosol index makes use of the spectral contrast of the measured signal between 340 and 380. The aerosol index charts we have been using were obtained with TOMS on Earth Probe. For these charts there are 10 classes of index ranging from 0.2 to 5.2 with a step of 0.5.

We will also present results of rawinsonde measurements and back trajectory calculations. OHP rawinsonde measurements were performed by Météo-France at noon at the Meteorological Center of Nimes, which is at ~150 km west of OHP. Thessaloniki rawinsonde measurements were obtained from the routine noon measurements performed by the Hellenic National Meteorological Service at the airport of Thessaloniki, which lies ~10 km southeast of the Lidar measuring site.

Up to 5 day back trajectories were calculated during dust transport events using the TM2Z model [Ramonet et al., 1996]. We computed trajectories arriving at the station in the four following different model layers: (1) 0.4 to 1.2 km, (2) 1.2 to 2.6 km, (3) 2.6 to 4.7 km, and (4) 4.7 to 7.3 km. For this calculation we have used the 0000 and 1200 UT predicted wind fields from the European Centre for Medium-Range Weather Forecasts. The methodology of trajectory calculation is based on emission of a unit tracer mass into the grid box where the back trajectory is desired and then calculation of the dispersion of the tracer with inversion of time and wind. Finally, we have been using direct meteorological observations reported in the European Meteorological Bulletin (EMB) [Deutschen Wetterdienstes, 1996, 1997].

3. Data Analysis

3.1. Sun Photometer Observations

Figure 2 is a two-dimensional histogram of the relative occurrence of a particular optical thickness at 870 nm compared to a particular Angström coefficient (two independent parameters). It was constructed using the cloud-decontaminated Sun photometer measurements performed in Thessaloniki from April 1996 to mid June 1997.

Two main classes corresponding to different conditions of aerosol load over Thessaloniki are apparent. The first one, which is the most frequent, corresponds to rather high values of the Angström exponent (1.25 ≤ a ≤ 2). These high values are due to Thessaloniki's usual aerosol load, which is made of small
particles present in polluted air masses (carbonaceous particles, and anthropogenic sulphates and nitrates) and/or in marine air masses (biogenic sulphates). The mean total atmospheric optical thickness at 532 nm is ~0.3 for both years, and the mean visible Angström exponent appears to be ~1.6. This pattern indicates a relatively high pollution aerosol load in Thessaloniki. The second class is characterized by a lower Angström exponent (0.3<α<0.9) and rather large optical thickness. Thus it is more characteristic of bigger particles such as mineral aerosols. A value ranging between 0.3 and 0.9 for dust Angström exponent would, however, be rather big compared to previous observations. Indeed, since dust particles are usually big (from a few tenths of a micron to 10 μm), their optical thickness is less spectrally dependent. This leads to values of the Angström exponent close to zero. Dulac et al. [1997] recently reported an Angström exponent of about a few hundredths for a dust event observed above Finokalia (Crete) in good agreement with measurements during dust transport in Cape Verde Islands [Chiapello et al., 1999] and Sahel [Holben et al., 1991]. It is most likely that the specificity of our site plays an important role. Indeed, because it is often highly polluted with a background AOT532 of 0.3, pollution and dust particles will contribute to the total aerosol optical thickness. The resulting equivalent Angström exponent measured will range somewhere in between the typical pollutionlike particle’s Angström exponent and the typical dustlike particle’s Angström exponent.

In Table 1 we have summed up the dates for which the Angström exponent decreased significantly (α<0.9). The presence of dust over Thessaloniki was confirmed by Meteosat observations for all these dates. TOMS observations (1997 only) indicate that dust is effectively present in the area of Thessaloniki. Thessaloniki events 5 and 8 (Ths-5 and Ths-8) are associated with an aerosol index ranging between 0.7 and 1.2 in the area of Thessaloniki whereas for all the others (Ths-6, Ths-7, and Ths-9) the aerosol index ranges between 0.2 and 0.7.

Plates 1a and 1b show Meteosat-derived optical thickness at 550 nm for May 8 and 9, 1997, respectively, together with the corresponding Sun photometer-derived optical thickness at about the same wavelength and the visible Angström exponent (Plates 1c and 1d). As revealed by Meteosat there is evidence of a dust plume along a SSW-NNE axis over the eastern Mediterranean, carrying dust to Thessaloniki on May 8. The presence of large dust particles above the city is confirmed by the Sun photometer measurements, as the Angström exponent is low (α ≈ 0.4) and the AOT532 is large (of the order of 0.4). On May 9, the plume has moved eastward, and though some residual dust particles may still remain in Thessaloniki’s atmosphere, the aerosol load comes back to average conditions. Indeed, the Angström exponent increases (α ≈ 1.2), indicating that smaller particles are controlling the aerosol load.

3.2. Lidar Observations

The evolution, from start to end, of the above described event has also been characterized by lidar soundings. In Plates 1e and 1f we have plotted the lidar-retrieved vertical profiles of the aerosol extinction coefficient at 532 nm for May 8 and 9. For both days the lower minimum of extinction coefficient corresponds to the top of the planetary boundary layer (PBL). Indeed, as can be seen in figure 3a, which represents the vertical profiles of relative humidity and temperature measured on May 8 at noon, there is a clear coincidence of the characteristic temperature inversion and the lower minimum of extinction. Both are situated ~0.8 km. At this stage, no conclusion can be drawn on the dust content of the PBL, but a discussion will be made in section 4.2.

On May 8 two other maxima of extinction can be detected in the morning (A.M. profiles are performed between 0930 and 1100 UT). A first one, of weak intensity, is situated between 1 and 2 km, and a second one clearly appears at 4 km. This vertical structure evolves during the day, in such a way that three maxima appear in the evening (P.M. profiles are performed between 1700 and 2000 UT): a broad one between 0.8 and 2.5 km, one at 3 km, and finally, one at 4.7 km. It is most likely that

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Total Aerosol Optical Thickness at 532 nm</th>
<th>Visible Angström Exponent</th>
<th>R</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ths-1</td>
<td>April 29, 1996</td>
<td>~0.35</td>
<td>~0.85</td>
<td>~0.8 ± 0.46</td>
</tr>
<tr>
<td>Ths-2</td>
<td>May 23, 1996</td>
<td>~0.2</td>
<td>~0.8</td>
<td>~0.8 ± 0.46</td>
</tr>
<tr>
<td>Ths-3</td>
<td>June 28, 1996</td>
<td>~0.35</td>
<td>~0.85</td>
<td>~0.8 ± 0.46</td>
</tr>
<tr>
<td>Ths-4</td>
<td>July 8, 1996</td>
<td>~0.35</td>
<td>~0.8</td>
<td>~0.8 ± 0.46</td>
</tr>
<tr>
<td>Ths-5</td>
<td>May 8, 1997</td>
<td>~0.40</td>
<td>~0.40</td>
<td>~2.4 ± 1.0</td>
</tr>
<tr>
<td>Ths-6</td>
<td>May 15, 1997</td>
<td>~0.30</td>
<td>~0.90</td>
<td>~0.68 ± 0.4</td>
</tr>
<tr>
<td>Ths-7</td>
<td>May 21, 1997</td>
<td>~0.35</td>
<td>~0.50</td>
<td>~1.8 ± 0.8</td>
</tr>
<tr>
<td>Ths-8</td>
<td>May 22, 1997</td>
<td>~0.30</td>
<td>~0.75</td>
<td>~1 ± 0.55</td>
</tr>
<tr>
<td>Ths-9</td>
<td>June 2, 1997</td>
<td>~0.30</td>
<td>~0.80</td>
<td>~0.8 ± 0.46</td>
</tr>
</tbody>
</table>

R is the ratio between the dust optical thickness and the background aerosol optical thickness (see section 4.1).

*These events have been sounded by the lidar system (events Ths-5 and Ths-9 in Tables 2-6).
Figure 3. Profiles derived from the rawinsonde measurements of May 8, 1997, performed at noon in Thessaloniki: (a) relative humidity, thermodynamic temperature, and potential temperature and (b) wind intensity and wind direction.

Further insights into the properties of the observed layers are obtained by looking at the rawinsonde measurements. Figure 3a clearly shows that the presumed dust layers are associated with relative humidity minima (<30%). On the other hand, the calculated potential temperature profile gives evidence that in the observed layers the potential temperature gradient is weak and positive, indicating that the atmosphere is stratified. The low relative humidity and associated weak potential temperature gradients are characteristics that are reminiscent of the so-called Saharan air layer observed during tropical Atlantic transport by Prospero and Carlson [1972]. Finally, Figure 3b, which gives the vertical profiles of wind direction and intensity, shows that winds have a large southerly component. This ensemble of facts seems to indicate that the extinction maxima observed up to 5 km above the PBL correspond to the layers in which African dust particles are trapped and transported.

Several other dust events were sounded during MEDUSE at both stations, and similar results have been obtained. In Figure 4 we have plotted the lidar-derived extinction coefficient profiles for May 15, 1997 at OHP. Two main aerosol layers are present both in the morning (between 0600 and 0800 UT) and afternoon.
At this stage we have seen two different examples of the possible evolution of the vertical structure of the dust transport. In the first case (Th-5) the structure passed from two to three layers in several hours. In the second case (OHP-1) the vertical structure remained unchanged for several days. We will now present a last example for which the vertical structure evolved from two layers to one (OHP-2). Plate 2 shows the temporal evolution of the lidar-retrieved extinction coefficient for May 17, 1997. (OHP-2) above OHP. One can clearly see the rapid transition of the vertical structure of the dust cloud from two layers to one at ~0830 UT.

Table 2 is a list of direct observations of the vertical structure of several African dust transports that occurred over the Mediterranean between January 1 and June 30, 1997. It should be remembered here that the referred soundings do not take into account the planetary boundary layer, which may also contain dust particles. Another point concerning Thessaloniki measurements is that because of a low signal-to-noise ratio above 5 km high it is not possible to highlight eventual scattering structures above that height. Thus the reported layers are not exclusive of possible higher layers.

The lidar-retrieved optical thickness is also given in Table 2. It was calculated from the top of the boundary layer (indicated by the temperature inversion) to the top of the highest dust layer indicated. It ranges between 0.1 and 0.25.

As is reported in Table 2, all the sounded events for both lidar stations have shown that particles are trapped and transported inside well-defined layers. Rawinsonde measurements have shown that these layers are relatively drier than the neighboring atmosphere. Most of the soundings have revealed several layers. A striking feature is the rather constant height (~2 km and ~4 km) at which the layers appear. Such a high-altitude transport may be linked to synoptic uplift that occurs when the warm Saharan air mass encounters the cooler Mediterranean air mass. Dulac et al. [1992a] have documented a case of such a synoptic uplift of the Saharan air mass in the western Mediterranean. They reported that it was uplifted with an average vertical velocity of 1 to 1.5 cm s⁻¹ up to 5 km high after 2 or 3 days of transport in the low troposphere. However, if such a mechanism can explain the presence of dust layers at high altitudes, it cannot explain observed complex vertical structures composed of several distinct dust layers.
Table 2. List of the Lidar-Sounded Dust Transports. DOT, dust optical thickness.

<table>
<thead>
<tr>
<th>Date (1997)</th>
<th>DOT at 532 nm</th>
<th>Number of Dust Layer(s)</th>
<th>Altitude of Layer(s) (km)</th>
<th>Thickness of Layer(s) (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>May 14 and 15</td>
<td>0.1</td>
<td>2</td>
<td>2.4</td>
<td>1.3, 0.8</td>
</tr>
<tr>
<td>May 17</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Early morning</td>
<td>0.1</td>
<td>2</td>
<td>2.5, 4</td>
<td>1.3, 0.6</td>
</tr>
<tr>
<td>Late morning</td>
<td></td>
<td>1</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>May 8</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Late morning</td>
<td>0.1</td>
<td>2</td>
<td>1.5, 4</td>
<td>1.0, 5</td>
</tr>
<tr>
<td>Late afternoon</td>
<td>0.2</td>
<td>3</td>
<td>1.5, 3, 4, 7</td>
<td>1.5, 0.8, 0.8</td>
</tr>
<tr>
<td>June 2</td>
<td>0.15</td>
<td>2</td>
<td>2.5, 4</td>
<td>0.8, 1.2</td>
</tr>
</tbody>
</table>

*These events have been sounded by the lidar system.

3.3. Dust-Loaded Air Mass Origin

For each of the lidar-sounded events we have been investigating the likely origins of the dust particles arriving above the measurement site at 0000 and 1200 UT. For that purpose we have been combining back trajectory calculations, Meteosat infrared observation, TOMS aerosol index [Herman et al., 1997], and meteorological surface charts from the European Meteorological Bulletin (EMB). Using these different sources of information, we have tried to highlight the likely dust origins.

Up to 5 day back trajectory were performed for the four levels mentioned in section 2.3. Their respective daily positions were projected onto the corresponding noon Meteosat infrared image to which we had previously applied the procedure described in section 2.2 and onto the corresponding TOMS aerosol index charts.

This superimposition of trajectories and turbid regions in North Africa suggests that in general, dust particles in the different layers do not originate from the same area because of different wind directions or different wind speed. The identified areas for the sounded events are reported in Table 3. It should be mentioned here that these areas are only given as an indication of the origins of the dust particles and that it is not intended here to precisely identify the dust sources since such a precise identification could be difficult to perform using Meteosat infrared observation or TOMS.

From Tables 2 and 3 it seems that there exists a more or less straight link between the number of different identified regions and the number of dust layers. Further insights into this link can be obtained by looking at two particular events for which we have been able to observe a significant change in the vertical structure (OHP-2 and Ths-5). In sections 3.3.1-3.3.2 we discuss in parallel changes in the number of identified regions and the evolution of the vertical structure of these two events.

3.3.1. Case study of May 17, 1997 (OHP-2). In Table 4 we have indicated when and where the day-by-day positions (1200 UT positions) of the 0000 and 1200 UT back trajectories have been coincident with zones identified as being dust loaded using Meteosat infrared observation. We have noted the position in which region local ground observations reported in the 0000 and 1200 UT surface charts of the EMB have reported dust uptake.

As it appears in Table 4 for all the period of study (from May 12 to May 16, 1997), at 0000 UT level 2 and 3 air masses originate...
from two well-differentiated zones which roughly correspond to east Morocco and south Tunisia while at 1200 UT they originate from east Morocco. This result is illustrated in Plate 3a, which shows the noon Meteosat infrared image for May 13, 1997. Dusty areas appear in hot colors (from yellow to red), whereas cloudy pixels have been set to white. The ending regions at 1200 UT on May 13 of the 0000 and 1200 GMT back trajectories performed in levels 2 and 3 are plotted as dashed lines. The corresponding TOMS aerosol index for the Morocco area ranges between 0.2 and 1.2, whereas it stands between 0.7 and 1.7 for Tunisia.

On May 13 and 14, the 1200 UT surface charts of EMB report the presence of two thermal lows (1005 mbar) centered on east Morocco and south Tunisia which are typical of that region for this time of the year [Jolliet et al., 1998]. During that period the meteorological conditions in these areas are unstable. The 0000 UT EMB charts reported direct observation of dust uptake on May 13 at one measurement station in south Tunisia. In Morocco, although no mention is made of the presence of dust, there is evidence of highly convective conditions which could be sufficient for dust uptake. Indeed, the region is highly cloudy, and the presence of vertically developed cumulus and cumulonimbus is reported.

From Table 4 it is apparent that levels 1 and 4 are also likely to transport dust to OHP since the air masses originate from a dust-loaded area. However, it is not possible to confirm the presence of dust in these layers above OHP because in the first case the layer is below the detection limit of 1.5 km of the lidar system and in the second case the presence of clouds does not allow the retrieval of the dust particles properties above 5 km.

It is striking that the evolution from two very different likely sources to one unique source for level 2 and 3 air masses between 0000 and 1200 UT is contemporary with the evolution of the lidar-derived vertical structure from two layers to one single layer. This seems to reinforce the hypothesis of a relation between the number of different likely sources for the different air masses in the different levels and the number of dust layers.

### Table 4. Results of the Comparison Between Likely Dust Sources and Day-by-Day Position of Air Mass Back-Trajectories for Event OHP-2

<table>
<thead>
<tr>
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<tbody>
<tr>
<td>Level 1</td>
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<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>no</td>
<td>no</td>
<td>yes (5.47°E, 36.48°N)</td>
<td>yes (8.45°E, 34.79°N)</td>
<td>yes (10.67°E, 33.70°N)</td>
</tr>
<tr>
<td>Level 2</td>
<td>0000</td>
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<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>no</td>
<td>no</td>
<td>yes (8.51°E, 35.16°N)</td>
<td>yes (10.16°E, 32.06°N)</td>
<td>yes (9.91°E, 30.5°N)</td>
</tr>
<tr>
<td>Level 3</td>
<td>0000</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>yes (0.46°W, 36.4°N)</td>
<td>yes (6.39°W, 33.63°N)</td>
<td>yes (6.42°W, 33.54°N)</td>
<td>yes (5.15°W, 35.43°N)</td>
<td>no</td>
</tr>
<tr>
<td>Level 4</td>
<td>0000</td>
<td>no</td>
<td>yes (8.07°W, 33.45°N)</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>no</td>
<td>no</td>
<td>yes (9.76°W, 31.97°N)</td>
<td>no</td>
<td>no</td>
</tr>
</tbody>
</table>

A "yes" means that the trajectory position coincides with a turbid area (position indicated in parentheses).

This means dust uptake is reported in the meteorological surface maps.

Table 5, for the entire period (from May 3 to May 7, 1997) at 0000 UT, level 2 air masses are from a dusty region situated roughly in north Libya (~16°E, 30°N), whereas level 3 air masses are from a region situated somewhere between south Morocco and north Algerian Sahara (around 7.19°W, 28.37°N and 4.55°W, 28.47°N). At 1200 UT the air mass source regions for levels 2 and 3 are the Grand Erg Occidental in north central Algeria (~2°E, 30°N) and the middle west of Tunisia (~5.30°E, 33.7°N) respectively. These results are illustrated in Plates 3b and 3c which are the same as Plate 3a but for May 4 and 7, 1997, respectively. Finally, from 0000 to 1200 UT level 2 and 3 air masses are of two different origins which are probably associated with the dust layers situated at 1.5 and 4 km, respectively, as can be seen in Plate 1c.

From May 5 to May 11, as revealed by the surface meteorological charts, a cold front having a S-N orientation has been progressing inside the Mediterranean basin. This cold front was associated with a low centered on Europe. The resulting increased winds are sufficient for the activation of dust sources. Confirmation of that point can be obtained once again by looking at the EMB surface charts, which report observations of dust uptake in some of the identified regions reported in Table 5. Moreover, Meteosat visible observation reveals that all along its displacement the cold front has been associated with dust outbreak inside the Mediterranean basin.

The dust layer situated ~5 km in Plate 1c is most likely due to dust advection in level 4. Indeed, despite the results reported in Table 5, it is evident in Plate 3c that level 4 back trajectory ending regions at 1200 UT on May 7 and 0000 UT on May 8 surround a likely dust source (northwest Algeria, 0.52°W, 36.20°N; north Tunisia, 10.83°E, 34.88°N), so that it is possible for level 4 air masses to be dust loaded between 1200 UT on May 7 and 0000 UT on May 8. This statement is backed up by the 0000 UT EMB chart of May 8, which reports dust uptake at two observation stations in the corresponding area.

Finally, from Table 5 one can see that level 1 is also likely to transport dust from south Libya. However, because of the height of this transport, dust particles advected above Thessaloniki will mix with the locally produced aerosols inside Thessaloniki PBL. Thus it is not possible to assess the presence of dust particles in
Plate 3. Meteosat noon infrared difference image and 5-day air mass back trajectories. The 0000 UT (crosses) and 1200 UT (plusses) back trajectories are computed in the different model levels (blue, level 4; red, level 3; yellow, level 2; white, level 1) for (a) May 13 (only level 2 and 3 back trajectories are plotted), (b) May 4 and (c) May 7.
Table 5. Same as Table 4 but for Event Ths-5

<table>
<thead>
<tr>
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</thead>
<tbody>
<tr>
<td>Level 1</td>
<td>0000</td>
<td>no</td>
<td>no</td>
<td>yes (21.69°E, 28.42°N)</td>
<td>yes (21.28°E, 29.72°N)</td>
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<tr>
<td></td>
<td>1200</td>
<td>no</td>
<td>no</td>
<td>yes (11.7°E, 29.19°N)</td>
<td>yes (12.99°E, 29.09°N)</td>
<td>yes (12.43°E, 31.49°N)</td>
</tr>
<tr>
<td>Level 2</td>
<td>0000</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>yes (16.05°E, 28.33°N)</td>
<td>yes (14.17°E, 32.31°N)</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>yes* (2.98°E, 32.57°N)</td>
<td>no</td>
<td>no</td>
<td>yes* (2.13°E, 29.03°N)</td>
<td>yes* (1.40°E, 29.96°N)</td>
</tr>
<tr>
<td>Level 3</td>
<td>0000</td>
<td>no</td>
<td>no</td>
<td>yes (4.55°W, 28.47°N)</td>
<td>yes (7.19°W, 28.37°N)</td>
<td>no</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>yes* (5.30°E, 33.7°N)</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>Level 4</td>
<td>0000</td>
<td>no</td>
<td>yes (4.41°W, 31.95°N)</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td></td>
<td>1200</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
<td>no</td>
</tr>
</tbody>
</table>

A "yes" means that the trajectory position coincides with a turbid area (position indicated in parentheses).

This means dust uptake is reported in the meteorological surface maps.

4. Sun Photometer and Lidar Synergy

From the above described study we have seen that dust-loaded air masses in the lower model layer (layer 1) (0.8+ 0.4 km) were likely to be advected above the measurement sites. It is indeed possible for that layer to be transported just above the marine boundary layer, whose top height is ~300-500 m, and to mix with the Thessaloniki PBL.

4.1 Respective Optical Thickness of Dust and Background Particles

Here we propose a procedure that uses Sun photometer measurements in order to apportion the respective contributions of large and small particles to the aerosol optical thickness. As dust particles are usually large and as Thessaloniki aerosol background is composed of small particles (see section 3.1), this will enable us to find out the proportion of dust particles that lies in the PBL.

Sun photometer measurements give access to the total atmospheric aerosol optical thickness (AOT_T). It can be written as the sum of the aerosol optical thickness below 700 m (region where our lidar is blind) and of the aerosol optical thickness in the rest of the atmosphere. In our case, where no significant stratospheric aerosol load has to be considered, the rest of the atmospheric contribution is due to aerosol in the free troposphere. The latter is mainly composed of dust particles. Let the dust optical thickness in the free troposphere be noted as \( \tau_{dT}(\lambda) \). Assuming that there are dust particles in the boundary layer, the total aerosol optical thickness is the sum of the dust optical thickness in the PBL (\( \tau_{PBL}(\lambda) \)) and of Thessaloniki’s aerosol background optical thickness (\( \tau_{bg}(\lambda) \)). Thus the optical thickness retrieved from the Sun photometer measurements can be expressed as:

\[
AOT_T = \tau_{PBL}(\lambda) + \tau_{bg}(\lambda) + \tau_{dT}(\lambda)
\]
Now it is possible to get another relation making use of the Sun photometer derived Angström exponent which is an equivalent Angström exponent. $(\bar{\alpha})$ that is characteristic of the mixture of all the aerosol that contributed to the measured signal: the dust particles and Thessaloniki’s aerosol background. Their respective Angström exponents $(\alpha_{\text{dust}}$ and $\alpha_{\text{bg}}$) are necessary to compute $\bar{\alpha}$ which can be expressed as:

$$\bar{\alpha} = \ln \left( \frac{1 + R}{R} \left( \frac{\lambda_1^{\alpha_{\text{dust}}} + \lambda_2^{\alpha_{\text{dust}}}}{(\lambda_1^{\alpha_{\text{bg}}} + \lambda_2^{\alpha_{\text{bg}}})} \right) \right)$$

$$\ln (\lambda_2^{\alpha_{\text{dust}}})$$

where $R$ is defined as

$$R = \frac{\tau_{\text{dust}}^{\text{TOT}} + \tau_{\text{dust}}^{\text{FT}}}{\tau_{\text{dust}}^{\text{BL}} + \tau_{\text{bg}}}$$

Thus provided that the equivalent Angström exponent is obtained from the Sun photometer, we are able to determine the value of the ratio.

Figure 5 shows the evolution of the equivalent Angström exponent as a function of $R$. It was calculated assuming $\alpha_{\text{dust}} = 0.0 \pm 0.1$, chosen according to measurements of dust events performed in rather clean atmosphere in the tropical Atlantic [Chiapello, 1996]. These measurements showed that during dust events the Angström exponent fluctuated between -0.1 and 0.1. A similar value has been reported in Crete [Dulac et al., 1997]. Regarding $\alpha_{\text{bg}}$, we have taken a value of 1.62 $\pm$ 0.26, where 1.62 is the yearly mean Angström exponent retrieved from the Sun photometer measurements, which we have supposed to be characteristic of Thessaloniki background aerosol conditions. and 0.26 is the associated standard deviation. Table 1 also gives the different $R$ that we have found for the corresponding dust events.

Because of the shape of the curves, the absolute error of $R$ resulting from the uncertainty of $\alpha$ will get dramatically bigger as $\alpha$ gets smaller. Thus the validity domain of such a procedure should be restricted to a range of $\alpha$ for which uncertainties are comparably small. This would ensure a stability of the solution found for $R$ (i.e., small variation in $\alpha$ will not generate large variations in $R$). We have calculated the uncertainties of $\alpha$ for May 8 and June 2 using (2). They are of $\approx$0.03 and 0.02, respectively. Using Figure 5 we have obtained the respective ratios: 0.52$\leq$$\alpha$$\leq$1.1 and 1.7$\leq$$\alpha$$\leq$3.1. These are used in the section 4.2 to retrieve the dust optical thickness in the PBL.

### 4.2. Dust Optical Thickness in the PBL

Once we have found the ratio $R$ between the total atmospheric dust optical thickness ($\text{DOT}_{\text{T}}$) and the background aerosol optical thickness, it is possible to derive the apportionment of $\text{DOT}_{\text{T}}$ between the PBL and the free troposphere from equations (5) and (8) using simultaneous lidar soundings. Indeed, $\tau_{\text{dust}}^{\text{TOT}}(\lambda)$ can be considered as being simply the dust optical thickness that we have retrieved from the lidar measurements. Simultaneous lidar and Sun photometer measurements are available for May 8 and June 2, 1997 (Ths-5 and Ths-9, respectively). Table 6 summarizes the results of the procedure applied to these days. It appears that a significant dust optical thickness is found in the PBL for Ths-5, while it is not the case for Ths-9. This seems to agree with the results of the back trajectory studies. Indeed, these have shown that for Ths-5, dust transport could take place in the air layer situated at 0.8 $\pm$ 0.4 km, while it could not for Ths-9.

### 5. Conclusion

We have reported on the vertical sounding of several North African dust outbreaks over the Mediterranean basin performed in the framework of the European MEDUSE project. These soundings were performed using two backscatter lidar systems operated at two stations (OHP, France, and Thessaloniki, Greece) in the western and eastern part of the Mediterranean basin from spring 1996 to mid 1997. The identifications and characterizations of these outbreaks were made using a synergy of lidar Sun photometer and satellite (Meteosat and TOMS) measurements.

Lidar soundings have revealed the multilayering of the dust transport above the Mediterranean basin. For each of the sounded events, several scattering structures were highlighted. These appeared to correspond to the layers in which dust particles were trapped and transported from turbid regions in North Africa. Their height ranged between 2.5 and 5 km. Our study has shown that this multilayering was associated with the complexity of the wind field (due to frontal conditions) that activated different dust sources. Frontal conditions are characteristics of the Mediterranean dust transport, and thus we believe that Mediterranean transport will be, in general, multilayered. This is just the contrary of the tropical Atlantic transport that is known to occur under trade winds inside one well-defined layer: the Saharan air layer.

### Acknowledgments

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