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Seasonal variability of optical properties of aerosols in the Eastern Mediterranean

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Abstract

The aerosol optical properties (scattering and absorption coefficients) were investigated at two remote locations in the Eastern Mediterranean in conjunction with aerosol ion composition measurements: Finokalia in the Crete Island in Greece (March 2001–June 2002) and Erdemli in Turkey (July 1999–June 2000). Ambient light-scattering coefficient ($\sigma_{sp-532\,nm}$) at Finokalia had a mean value of $50 \pm 23\,Mm^{-1}$ while at Erdemli this value was $90 \pm 160\,Mm^{-1}$, due to a severe dust event that occurred from 17 to 19 April 2000. Scattering coefficients up to $5000\,Mm^{-1}$ were encountered during the transition periods (spring and autumn) and were associated with dust storm events. During these events significant correlations were observed between dust and σ_{sp} and mass scattering efficiencies of 0.21 and $0.96\,m^2g^{-1}$ were calculated for dust for Finokalia and Erdemli, respectively. Significant correlations were also observed at both locations between non-sea-salt sulphate (nss-SO_4^{2-}); σ_{sp} and mass scattering efficiencies of 5.9 ± 1.8 and $5.7 \pm 1.4\,m^2g^{-1}$ were calculated for the nss-SO_4^{2-} at Finokalia and Erdemli, respectively. At Finokalia absorption measurements were also performed at the same time and the mean absorption coefficient ($\sigma_{ap-565\,nm}$) was found to be $5.6 \pm 3.6\,Mm^{-1}$. Maxima of absorption coefficient were associated with two distinct meteorological situations indicative of pollution transported from northern Europe and Saharan dust events. Saharan dust can therefore significantly contribute to both scattering and absorption of solar radiation, the latter due to its hematite content. Based on scattering and absorption measurements, an annual mean single-scattering albedo (ω) adjusted at 550 nm of 0.89 ± 0.04 was calculated for Finokalia. Finally, radiative forcing efficiency (RFE) over the sea at 550 nm induced by aerosols has been calculated for Finokalia. RFE follows a clear seasonal variation, with the lowest mean values during summer ($-73\,W\,m^{-2}$) and the highest during winter ($-30\,W\,m^2$). Using aerosol optical thickness measurements in the area, we obtain radiative forcing estimates at the top of the atmosphere (TOA) ranging from -12.6 to $-2.3\,W\,m^2$ for summer and winter, respectively. These values are up to five times higher than that induced by the greenhouse gases ($2.4\,Wm^{-2}$) but opposite in sign.
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Keywords: Optical properties of aerosols; Seasonal variation; Chemical composition; Radiative forcing; Eastern Mediterranean

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1. Introduction

The role of aerosols on climate is a major environmental issue nowadays. Contrary to the greenhouse gases, which are widespread and well mixed globally, the distribution of aerosols is quite patchy and depends on the presence of either natural (sea-water, dust, vegetation) and/or anthropogenic sources. The Mediterranean area has been repeatedly suggested as one of the areas where aerosols are expected to play a significant climatic role (Formenti et al., 2001a, b; Andreae et al., 2002; Lelieveld et al., 2002; Tragou and Lascaratos, 2003). During an intensive campaign performed in the Eastern Mediterranean (MINOS campaign) in August 2001, Lelieveld et al. (2002) reported that aerosols can reduce solar radiation absorption over the sea by about 10% and they can alter the heating profile of the lower troposphere (Markowicz et al., 2002). As a consequence, evaporation and moisture transport, in particular to North Africa and the Middle East, are suppressed. Therefore, a clear need exists for long-term measurements of aerosol physical, chemical, and radiative properties in the Eastern Mediterranean which is the cross-road of air pollution from Western and/or Eastern Europe (Lelieveld et al., 2002).

Although several studies on physicochemical characterisation of aerosols have been performed in the Eastern Mediterranean, most of them are limited to specific periods of the year, mainly during summer (Ichoku et al., 1999; Formenti et al., 2001a, b; Kouvarakis et al., 2002; Andreae et al., 2002; Sciare et al., 2005; Bryant et al., in press). Among these works, only one addressed the variability of chemical and optical characteristics of aerosols on a long-term basis and was performed at the Negev Desert (Andreae et al., 2002).

The current study has two aims:

(i) To present long-term (at least 1 year) measurements of chemical and optical properties of aerosols in different environments in the Eastern Mediterranean compared to the Negev Desert. For this reason, two coastal areas located at Erdemli (Turkey) and Finokalia (Crete, Greece) disposing long-term published data on chemical composition of aerosols were selected.

(ii) To estimate the radiative forcing efficiency (RFE) and radiative forcing estimates at the top of the atmosphere (TOA) induced by aerosols on a seasonal basis, based on scattering and absorption measurements conducted for more than a year at Finokalia.

2. Experimental

2.1. Location and sampling

Sampling was conducted at two remote coastal sites (Fig. 1), both located at the Eastern Mediterranean region. The first one is Finokalia station (35.3 °N, 25.7 °E) located on the north coast of Crete Island in Greece, and the second one is located at Erdemli in Turkey (36.6 °N, 34.1 °E) on the south coast of Turkey. A more detailed description on these sampling sites and the prevailing meteorological conditions are reported by Kubilay et al. (2000) and Kouvarakis et al. (2001).

2.2. Aerosol sampling and analysis

At Finokalia, bulk aerosol samples were collected on 47 mm diameter, 0.45 μm porosity Gelman Zefluor PTFE filters. Samples were collected on 2–3 days basis from March 2001 to June 2002 except summer 2001 when samples were collected every 12 h. Aerosol sampling at Erdemli was performed during two periods with a daily sampling step. The first one lasted from January 1999 to June 2000 and the second one from January 2001 to November 2001. At Erdemli aerosol samples were collected on Whatman-41 filters (20 × 25 cm) using a high-

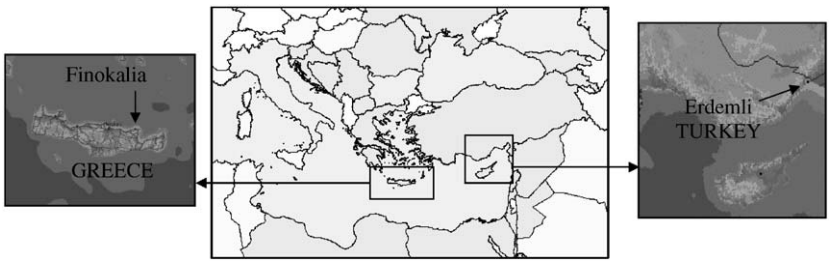

Fig. 1. Location of the Finokalia/Greece and Erdemli/Turkey stations in the Eastern Mediterranean region.

volume sampler operating at a flow rate of $50\,m^3h^{-1}$ with a sampling efficiency higher than 95%. In both stations, the aerosol samples were analysed for main ionic species by ion chromatography (IC). A comprehensive description on the IC measurements and the methodology can be found in Kouvarakis and Mihalopoulos (2002).

2.3. Optical measurements

2.3.1. Light scattering coefficient measurements

Aerosol light scattering coefficient (σ_{sp}) was measured at both locations with a mono-wavelength portable integrating nephelometer (M903, Radiance Research, Seattle, USA). This instrument measures the light scattering coefficient (σ_{sp}) at 532 nm. The optical and electrical background noise is sufficiently low to allow measurements of σ_{sp} (for particles) from less than 10% of air Rayleigh scattering ($\sigma_{sp} < 1\,Mm^{-1}$) to greater than $1000\,Mm^{-1}$.

Measurements were performed at ambient relative humidity (RH), without aerosol cut-off. Instruments were calibrated using CO_2 and zero checks were frequently performed by means of a HEPA filter. For the Finokalia nephelometer an intercomparison was performed in July 2001 with a second Radiance Research Nephelometer, operated by the Max-Planck Institute, and showed a good agreement with deviations below 5%.

2.3.2. Absorption measurements

A commercial instrument particle soot absorption photometer (PSAP), Radiance Research, Seattle, USA) was used for the measurement (in near real time) of the light absorption coefficient of ambient aerosols at Finokalia station. Total uncertainty of PSAP measurement at 60 s averaging time and typical atmospheric levels was estimated around 15% at the 95% confidence level (Bond et al., 1999). The PSAP measurement is derived by the manufacturer as the difference between light extinction and scattering (i.e. light absorption). The results presented here have been corrected for scattering effects using the equation proposed by Bond et al. (1999), which lowers the annual mean measured values by about 10%.

3. Results and discussion

The results of this work will be split in to four parts. The ionic composition of aerosols at the two stations will be presented in the first part. In the second and third parts, the optical properties of aerosols, namely light scattering and absorption coefficients, are presented and the factors controlling their levels are discussed. The role of chemical composition and especially the link between nss-SO_4^{2-}, dust content and optical parameters will also be addressed. Finally, the RFE and the radiative forcing estimates at the TOA induced by the optical properties of aerosols will be discussed in the fourth part.

3.1. Ionic composition

Based on studies conducted at both locations (Bardouki et al., 2003; Koçak et al., 2004), the mass associated with the main ions is a major fraction of the total aerosol mass (accounting for up to 50%). The following discussion will be limited to two ions: nss-SO_4^{2-} and nss-calcium (nss-Ca^{2+}). Nss-SO_4^{2-} is the main ion in the fine mode contributing up to 65% of the total ionic mass of this mode and it is frequently used as a tracer for anthropogenic activity, although in summer time marine biogenic contribution on nss-SO_4^{2-} is around 25% at both sites (Kubilay et al., 2002). Although nss-Ca^{2+} is responsible for only up to 12% of the coarse mode mass, it can be used as a dust load indicator. Indeed, Markaki et al. (2003) and Sciare et al. (2005), based on measurements conducted at both locations, showed that nss-Ca^{2+} correlates significantly with Al, a ubiquitous tracer of terrigenic influence (Wedepohl, 1995), with the relative slope ranging from 1.0 to 1.3. Thus, dust can be estimated using the observed nss-Ca^{2+} values and the relation between Al and dust ($[Al]\% = 7.09 \pm 0.79$ [dust]) (Guieu et al., 2002). Note that this indirect dust estimation may induce errors up to 40%.

The variability of nss-SO_4^{2-} (Fig. 2a) and nss-Ca^{2+} (Fig. 2b) for the studied periods is presented for Finokalia and Erdemli measuring sites. At Finokalia as well as at Erdemli, nss-SO_4^{2-} shows a well-defined seasonal variation with higher values during the dry period (May–October) compared to the wet one (December–March). The nss-SO_4^{2-} levels at both locations are in good agreement with previously published works in this area (see for example Kouvarakis and Mihalopoulos, 2002; Kubilay et al., 2002; Koçak et al., 2004). The nss-Ca^{2+} has a more uniform behaviour compared to nss-SO_4^{2-}, with the highest values occurring during the transition periods (spring and autumn), and

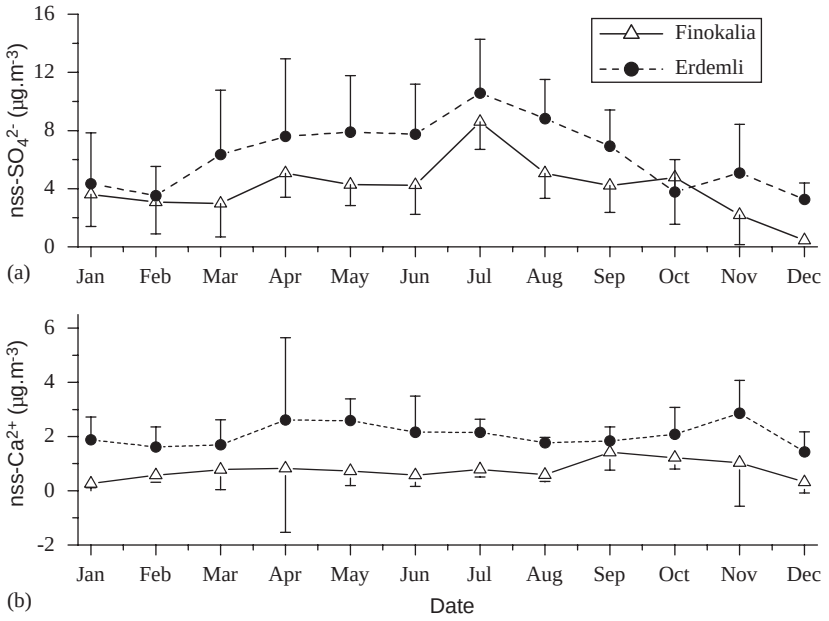

Fig. 2. Monthly mean values and standard deviations of (a) nss-SO_4^{2-} and (b) nss-Ca^{2+} at Finokalia for the period 2001–2002 and at Erdemli for the period 1999–2000.

coincide with dust events from arid areas of Northern Africa or Arabian Peninsula (Israelevich et al., 2003).

3.2. Aerosol light scattering coefficient (σ_{sp})

3.2.1. Variability of aerosol light scattering coefficient (σ_{sp}) at both locations

(a) *Finokalia:* The observations lasted for 17 months (July 2001–November 2002) and the σ_{sp} 5-min mean values ranged between 5 and 445 Mm^{-1} with a mean value of $50 \pm 23\,Mm^{-1}$ (Fig. 3). The highest values were linked to the presence of south winds with high dust load as observed from the air mass back-trajectories (see for instance the event of 15—16/4/2002). Trajectories (provided by www.arl.noaa.gov/ready/hysplit4.html) were calculated at 1000 m above sea level to avoid orographic interferences at lower heights and to be consistently within the boundary layer of the region which ranges between 1000 and 1300 m. Trajectories were produced for all cases with σ_{sp} exceeding twice the mean value of the corresponding month, and the results are depicted in Fig. 4. All events of high σ_{sp} are associated either with transport from regions affected (i) by enhanced anthropogenic influence (N sector, Europe; 49% of the cases), (ii) by high dust load (S sector; 17% of the cases) and (iii) pollution sources and high dust load (NE sector, former Soviet Union countries and Turkey; 34% of the cases) (Sciare et al., 2003a).

Aerosol scattering coefficient presents a distinct seasonal pattern (Fig. 5). Maximum values are found during the dry period and minimum values are encountered during the wet period as is in the case of nss-SO_4^{2-}. The summer maximum σ_{sp} is explained by the longer lifetime of aerosols during the dry period as well as the prevalence of winds from the N and NE sectors. The spiky pattern of the σ_{sp} observed during the transition periods (autumn and spring) was related to Sahara dust events. Moreover, the role of the NE sector is clearly depicted in October 2001, when σ_{sp} values were almost twice the values observed in October 2002 (Fig. 5). Trajectory analysis revealed that the occurrence of NE winds during October 2001 was higher than during October 2002 by a factor of six.

(b) *Erdemli*: σ_{sp} observations at Erdemli covered the period from July 1999 to June 2000 and ranged from 0.7 up to 4860 Mm^{-1} (Fig. 6). As at Finokalia, the highest value (observed in April 2000) was associated with an important dust event. Kubilay et al. (2003, 2005) have shown that Saharan-based dust outbreaks predominantly take place at Erdemli station from mid-March to the end of August, with the most intensive events observed in April–May.

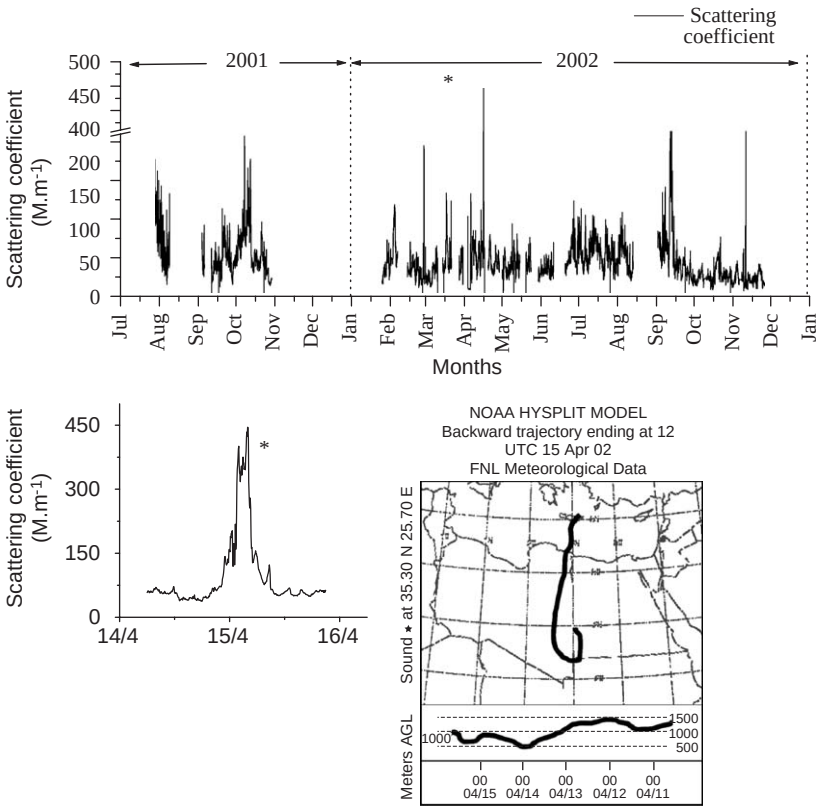

Fig. 3. Variability of aerosol light scattering coefficient at Finokalia during 2001–2002. The * depicts an intensive dust event that occurred in April 2001.

The mean σ_{sp} for the whole period was $45 \pm 22\,Mm^{-1}$, whereas when April's 2000 dust event was included, the annual average at Erdemli was found to be $90 \pm 160\,Mm^{-1}$. The annual variation of σ_{sp} at Erdemli is presented in Fig. 7.

The σ_{sp} values observed at both locations are in accordance with values reported for the Eastern Mediterranean region. In particular, Gerasopoulos et al. (2003) reported a mean scattering coefficient of $65 \pm 40\,Mm^{-1}$ in northern Greece, mainly resulting from continental pollution. The corresponding levels at Erdemli can be directly compared to the ones reported by Andreae et al. (2002) in Israel. In that work, a mean scattering coefficient of $87 \pm 54\,Mm^{-1}$ was observed as a result of the combination of continental pollution and dust events.

3.2.2. Mass scattering efficiencies (a_λ) at both locations:

(*a*) *Mass scattering efficiency for nss-SO_4^{2-} ($\alpha_{nss\text{-}SO_4}$):* The calculation of the mass scattering efficiency for nss-SO_4^{2-} was deduced from the slope of the total scattering coefficient versus the nss-SO_4^{2-} mass after integrating the scattering coefficient to the nss-SO_4^{2-} sampling interval. The calculations were performed for each season separately and statistically significant correlations were observed for all seasons at both locations leading to $\alpha_{nss\text{-}SO_4}$ values with uncertainty lower than 20% (Fig. 8).

Both at Erdemli and Finokalia, $\alpha_{nss\text{-}SO_4}$ was found to be higher during the cold and wet seasons (autumn and winter), while during the warm and dry seasons (spring and summer), $\alpha_{nss\text{-}SO_4}$ dropped significantly (Fig. 8). This can be mainly attributed to changes in RH which increased from about 60% in summer to about 70% in winter. Note that when the RH increases from 60% to 70%, the scattering enhancement factor f(RH) increases by about 40% (http://vista.cira.colostate.edu/improve/).

Preliminary field measurements were conducted at Finokalia using a tandem nephelometer (a drier was installed before one of the nephelometers while

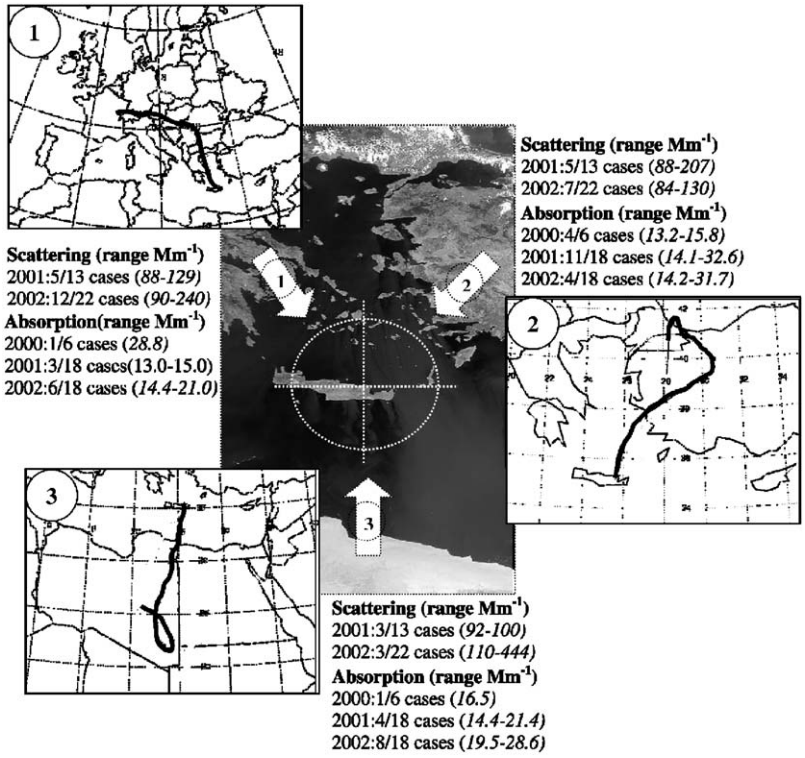

Fig. 4. Trajectory analysis at Finokalia for the aerosol scattering coefficient (2000–2001) and absorption coefficient (2000–2002) events exceeding the mean value of the corresponding month by at least a factor of two. The thick black line represents the mean trajectory at 1000 m height. Under each class of air masses, the year, the number of exceedences relative to the total number of cases and the range of scattering and absorption coefficients are presented.

the other one operated at ambient RH). These measurements showed a significant influence of RH on the scattering coefficient. More specifically, scattering coefficients increased almost exponentially above RH of 60% and work is under progress to compare the empirical formula for f(RH) with experimental data.

The annual mean $\alpha_{\text{nss-SO}_4}$ at Finokalia and Erdemli have been calculated to be equal to 5.9 ± 1.8 and $5.7 \pm 1.4\,\text{m}^2\,\text{g}^{-1}$, respectively, which are in good agreement with the values reported in the literature (Leaderer et al., 1981; Quinn et al., 1996; Meszaros et al., 1998; Ichoku et al., 1999; Formenti et al., 2001a, b; Bryant et al., in press) (Fig. 8).

(*b*) *Mass scattering efficiency for dust (α_{dust}):* We have seen that dust events can significantly contribute to scattering variability at both locations. Calculating α_{dust} requires caution since i) α_{dust} is significantly lower (by almost a factor of 10) than that of $\alpha_{\text{nss-SO}_4}$ (Andreae et al., 2002) and (ii) dust is a medium that favours the absorption of acidic species (like SO_2 and HNO_3), thus increasing nss-SO_4^{2-} levels, which in their turn increase the total scattering. Therefore, the analysis has been limited only to significant dust events (above $50\,\mu\text{g.m}^{-3}$) with low nss-SO_4^{2-} levels (below $3\,\mu\text{g.m}^{-3}$). Even though this criterion decreases the number of cases studied, significant correlations are revealed between scattering coefficient and dust content with slopes equal to 0.21 ($r^2 = 0.96$, $n = 5$) and 0.96 ($r^2 = 0.81$, $n = 17$) at Finokalia and Erdemli, respectively. Note that when the regression between total scattering coefficient and dust content is applied to all dust events (identified via back-trajectories), much higher α_{dust} is calculated (1.43 and 1.57 at Finokalia and Erdemli, respectively). This fact denotes the critical role of nss-SO_4^{2-}, i.e. even in small amounts in the coarse mode, they still contribute significantly to the ability of aerosols to scatter the light (Mihalopoulos et al., unpublished data).

The uncertainties of the α_{dust} calculations could be as high as 50% due to uncertainties in the dust

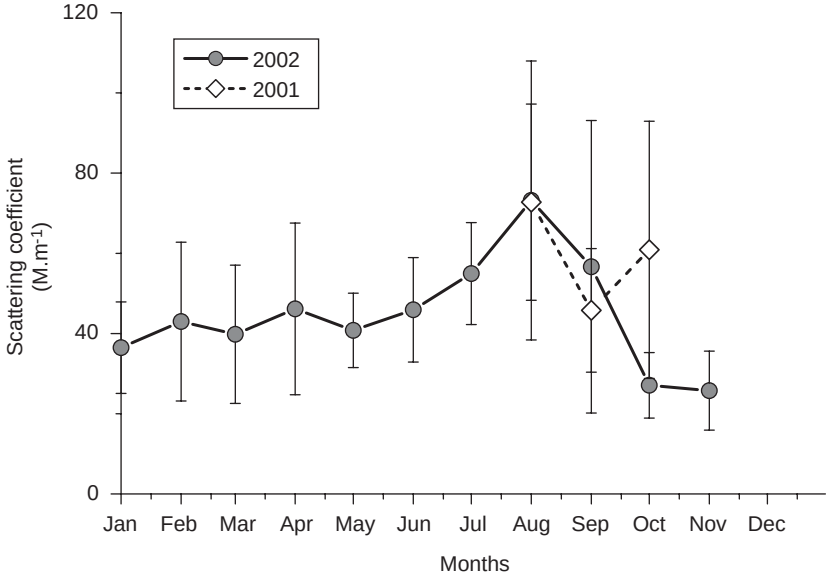

Fig. 5. Monthly mean aerosol scattering coefficient at Finokalia during the period 2001–2002.

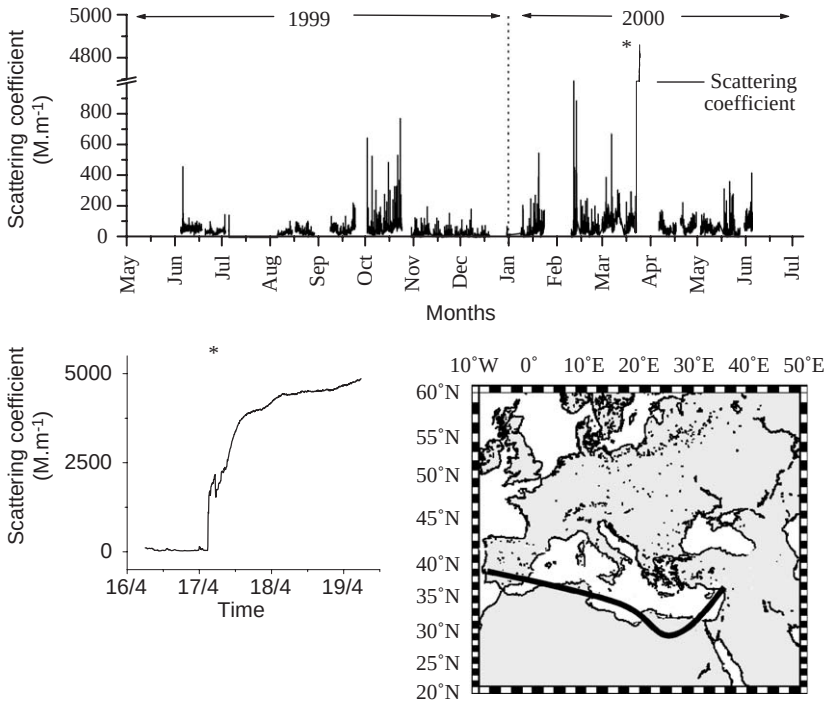

Fig. 6. Aerosol scattering coefficient variation at Erdemli as calculated for the period 1999–2000. The * depicts an intensive dust event that occurred in April 2000.

mass calculations (up to 40%) and uncertainties in the regression slopes (10%). Thus, the α_{dust} calculated for Finokalia and Erdemli (0.21–0.96) are in relatively good agreement with the value of $0.52\,m^2\,g^{-1}$ proposed by Andreae et al. (2002) based on measurements conducted at Negev Desert in Israel.

(*c*) *Specific mass scattering efficiency for fine mass* (α_{fm}): The α_{fm} was calculated only for the summer 2001 (MINOS campaign) at Finokalia. The

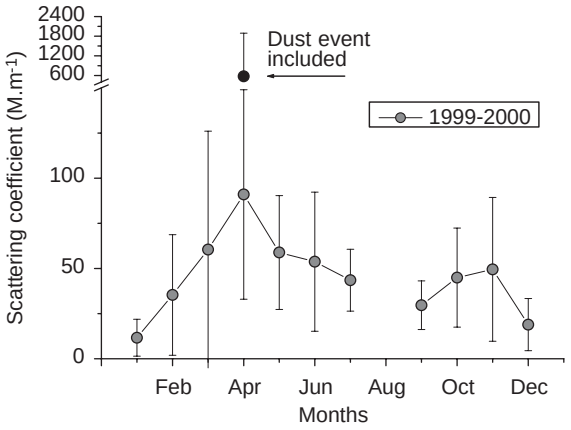

Fig. 7. Monthly mean aerosol scattering coefficient at Erdemli for the 1999–2000 period. The black dot includes the April 2000 dust event.

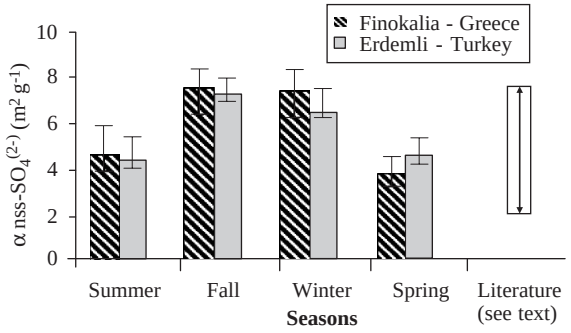

Fig. 8. Seasonal variation of the mass scattering efficiency of nss-SO_4^{2-} calculated for Finokalia and Erdemli and comparison to the literature.

scattering coefficient of the fine mass was calculated, considering the mass measured at the back filter of a Stack Filter Units (SFUs) with a cut-off of the order of $1.2 \pm 0.1\,\mu m$ being representative of the fine mass fraction (Sciare et al., 2005). A significant correlation has been observed between fine mass and scattering coefficient with a slope of $3.1 \pm 0.5\,m^2\,g^{-1}$ ($r^2 = 0.44$, $n = 45$), comparable to the values proposed by White et al. (1994) and Andreae et al. (2002).

3.3. Absorption coefficient (σ_{ab})

The main species that absorb radiation are black carbon and hematite found in dust aerosols. Alfaro et al. (2004) indicated that the effect on light absorption of iron oxides, which are the main absorbing minerals in desert dust, could be of the same order of magnitude as black carbon (BC).

Absorption measurements were performed for nearly the same sampling period as the scattering coefficient measurements at Finokalia during the period 2000–2002 (Fig. 9). Absorption coefficients ranged between 0.1 and $33\,Mm^{-1}$ with a mean value of $5.6 \pm 3.6\,Mm^{-1}$. In August 2001, the significantly higher value ($11.2 \pm 6.0\,Mm^{-1}$, Fig. 9b) compared to that measured during the same month in 2000 was attributed to biomass burning activities in the area (Salisbury et al., 2003; Sciare et al., 2003b).

High absorption coefficients are divided in two subcategories symbolized as (I) and (II) (Fig. 9). The first one (I) corresponds to air masses from the NW–NE directions and the second (II) corresponds to south winds. It is well known that black carbon is mainly an anthropogenic pollutant. Northwest, north and northeast winds are expected to be rich in BC as they pass through central and eastern Europe (Sciare et al., 2003a). An explanation for absorption coefficient peaks during the second subcategory (south winds) is the hematite (Bohren and Huffman, 1983; Pinnick et al., 1993; Alfaro et al., 2004). Its characteristic property of absorbing in the visible region of the solar spectra due to its reddish colour leads to a positive artefact. A detailed trajectory analysis was performed for all cases with absorption coefficient exceeding twice the mean value of the corresponding month, and the results are presented in Fig. 4. It is clear that all absorption events, as for scattering, are associated either with transport from regions affected by strong anthropogenic influence (N sector, 24% of the cases), by high dust loading (S sector, 31% of the cases) or both (NE sector, 45% of the cases).

The absorption coefficient data were also analysed on a seasonal basis (Fig. 10). Excluding August 2001 data, no clear seasonal signal was observed. Thus, although the sources influencing aerosol absorption change seasonally, aerosol absorption at Finokalia seems to remain almost constant throughout the year.

3.4. Radiative forcing

In this section, the direct aerosol impact on the radiative forcing is calculated based on the observations of optical properties at Finokalia.

One of the most important properties of aerosols is the single-scattering albedo—ssa (ω), defined as the ratio of the scattered radiation (σ_{sp}) to the total

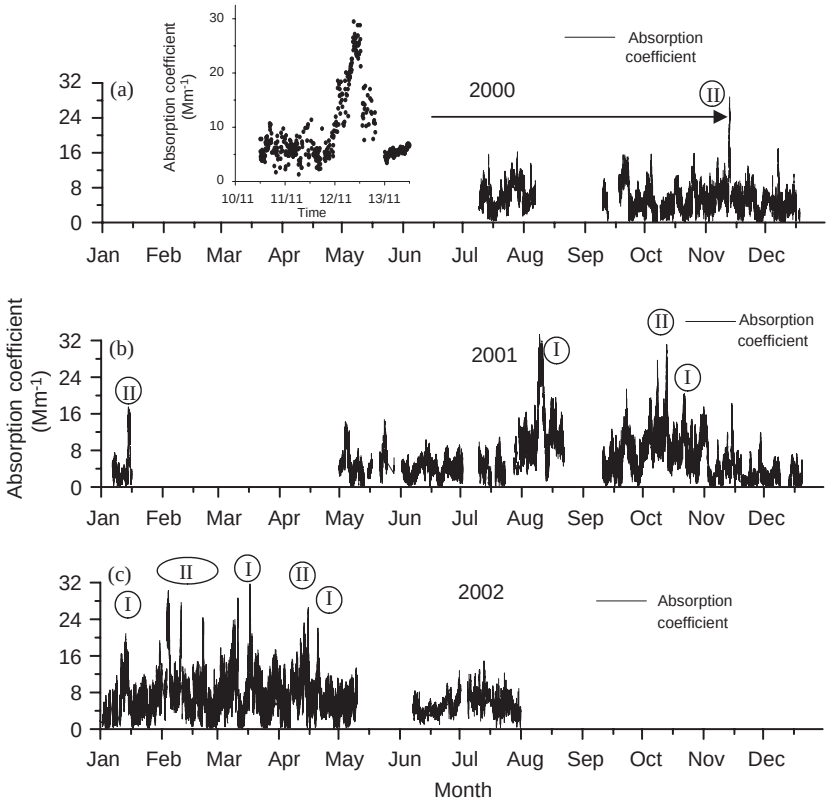

Fig. 9. Aerosol absorption coefficient at Finokalia as observed during the periods 2000 (a), 2001 (b) and 2002 (c).

radiation ($\sigma_{sp}+\sigma_{ab}$). The average calculated ω for the total period was 0.89 ± 0.04 and the lower and upper limits were 0.74 and 0.98, respectively. The lowest values of ω coincide with increased absorption coefficients occurring when air parcels were affected by biomass burning events in the neighbouring countries in conjunction with strong north winds (i.e. 8–13 August 2001) (Sciare et al., 2003b).

Another important aspect of the climate–aerosol interaction is the RFE, which is the amount of the radiative forcing that increases the optical thickness by one unit. In order to have a rough estimation of radiative forcing produced by anthropogenic aerosols, the RFE was calculated using the following equation (Haywood et al., 1995; Anderson et al., 1999; Andreae et al., 2002):

$$\mathrm{RFE} = SD(1-A_c)T^2(1-R_s)^2\left[2R_s\frac{1-\omega}{(1-R_s)^2}-\beta(b)\omega\right],$$

where S is the solar constant equal to $1370\,\mathrm{W\,m^{-2}}$, D the daylight fraction (changes seasonally), A_c the mean fractional cloud based on observations in the area and changes seasonally, T the atmospheric transmissivity (0.76), R_s is the surface albedo (0.07 for sea and 0.34 for land), $\beta(b)$ the average upscatter fraction derived from the backscatter ratio (of the order of 13% for our region) (Formenti et al.,

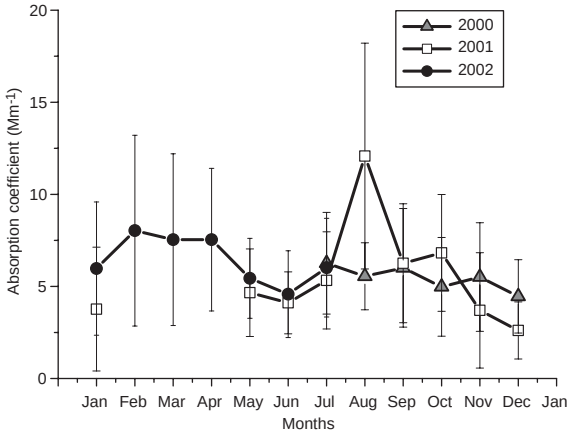

Fig. 10. Monthly mean aerosol absorption coefficient at Finokalia during the sampling period (2000–2002).

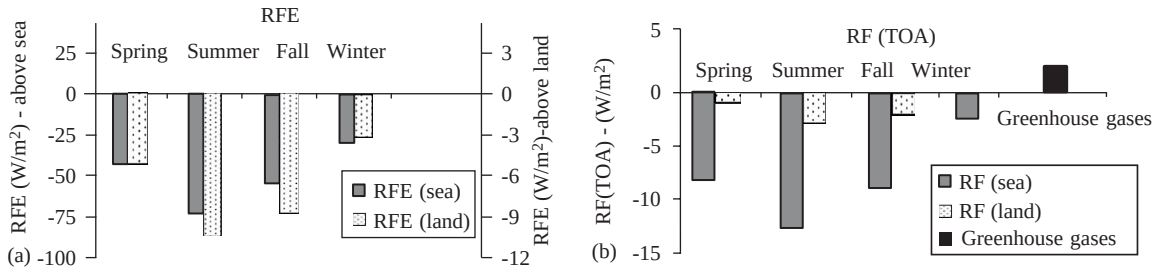

Fig. 11. Seasonal variation both above the sea and land of the (a) radiative forcing efficiency (RFE) and (b) radiative forcing (RF) at the top of the atmosphere (TOA).

2001b; Bryant et al., in press), and ω is the single-scattering albedo in the area.

The RFE is irreversibly correlated to ω and thus positively correlated to absorption. It ranges from -99 up to $-26\,\mathrm{W m^{-2}}$ above sea and from -31 up to $32\,\mathrm{W\,m^{-2}}$ above land. Negative values correspond to cooling effects while positive ones are attributed to the heating. From the data analysis, RFE at Finokalia showed a clear seasonal trend (Fig. 11a). Above the sea, the mean summer values had an average of $-73\,\mathrm{W m^{-2}}$ whereas during winter the mean value was $-30\,\mathrm{W m^{-2}}$. Spring and autumn had presented intermediate values equal to $-43\,\mathrm{W\,m^{-2}}$ and $-54\,\mathrm{W\,m^{-2}}$, respectively. To obtain radiative forcing at the TOA, RFE values were multiplied with the aerosol optical thickness provided by the AERONET site at Crete (http://aeronet.gsfc.nasa.gov/) (annual mean value for 2002–2003 of 0.158 ± 0.0456). Radiative forcing at the TOA ranged from $-2.3\,\mathrm{W m^{-2}}$ (winter) up to $-12.6\,\mathrm{W\,m^{-2}}$ (summer) (Fig 11b). These values are up to five times but opposite in sign to the forcing induced by the greenhouse gases and estimated to be equal to $2.4 \pm 0.2\,\mathrm{W\,m^{-2}}$ (Houghton et al., 1995) and shows the important direct role of aerosols in the area almost year-round. For comparison Markowicz et al. (2002), using a detailed radiative transfer code calculated for Finokalia during summer 2001 a mean radiative forcing at TOA equal to $-8.3 \pm 2.7\,\mathrm{W\,m^{-2}}$, which is close to the results found during this study and for the same season ($-12.6\,\mathrm{W\,m^{-2}}$).

4. Conclusion

The main optical properties of aerosols, namely scattering and absorption coefficients, define the level of warming and/or cooling of the atmosphere from the aerosols. In this study, these two aerosol properties were studied at two regions in the Eastern Mediterranean: the Finokalia station (north coast of Crete, Greece) and the Erdemli station (south coast of Turkey). The specific area is of highly interest as it is a crossing point of masses of different origin (Lelieveld et al., 2002).

Ambient scattering coefficients at Finokalia show a slight seasonality with maximum values during summer. High values have been also observed during the transition periods mainly due to dust events. The mean scattering coefficient at Finokalia ($50\,\mathrm{Mm^{-1}}$) is comparable to that obtained in northern Greece (Gerasopoulos et al., 2003) while the average scattering coefficient at Erdemli ($90\,\mathrm{Mm^{-1}}$) agrees well with the value found in Israel (Andreae et al., 2002). The Turkish station is closer to dry desert areas, thus it is affected to a greater extent by dust events. As a result there is no clear seasonal pattern for this region.

The nss-SO_4^{2-} contributes to scattering, and its mass scattering, efficiency $\alpha_{\mathrm{nss\text{-}SO_4}}$ showed a seasonality possibly due to the changes in the RH. The calculated scattering efficiency for dust α_{dust} falls in the range of the values reported by Andreae et al. (2002) at Negev Desert.

The absorption levels recorded at Finokalia did not show any significant seasonality. Thus, although the sources influencing aerosol absorption change seasonally, aerosol absorption at Finokalia seems to remain almost constant year-round.

Based on the measured scattering and absorption coefficients, a single-scattering albedo (ω) of 0.89 ± 0.04 at 550 nm was calculated for Finokalia. The RFE over the sea induced by the measured optical properties of the aerosols was found to vary seasonally with a mean summer minimum of $-73\,\mathrm{W m^{-2}}$. Finally, the radiative forcing at the TOA ranged from -12.6 to $-2.3\,\mathrm{W\,m^2}$ for summer and winter, respectively, values which are up to five times but opposite in sign to the forcing induced by the greenhouse gases.

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