

Desert Aerosols Transported by Khamsinic Depressions and Their Climatic Effects¹

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ABSTRACT

The mass of natural aerosol injected into the atmosphere by desert wind systems of the Khamsinic type has been estimated from long series of normal-incidence spectral solar radiation measurements and from synoptic analysis of the temporal and areal extent of such phenomena.

From these data the regional and global dust input into the atmosphere due to deserts has been inferred and compared to other estimates.

A tentative value for the effect of such desert aerosols on the average global albedo and surface temperature is also presented.

1. Introduction

Recently, the influence of aerosols on the thermal regime of a planetary atmosphere, both directly and by their effect on cloud properties, has become of much interest [e.g., McCormick and Ludwig, 1967; Peterson and Bryson, 1968; Budyko, 1969; Bryson and Wendland, 1970; Galindo and Muhlia, 1970; Robinson, 1970; *Man's Impact on the Global Environment* (S.C.E.P.), 1970; *Study of Man's Impact on Climate* (S.M.I.C.), 1971].

The division of the atmospheric aerosol load into a natural and a man-made part cannot be easily made, although some effects of man's activity can be already singled out (e.g., Cobb and Wells, 1970; Bryson and Wendland, 1970; S.M.I.C., 1971) at least on a regional basis.

Desert or savannah dust, especially that of Sahara or Libyan Desert origin, is an appreciable fraction of the atmospheric natural aerosol content over large regions of the globe. Wind-blown desert aerosol from North Africa is frequently discovered far away from its source (e.g., Mörkofer, 1941; Prospero, 1968; Stevenson, 1969; Bryson and Wendland, 1970; Volz, 1970; Volz and Sheehan, 1971; S.M.I.C., 1971; Goldberg, 1971).

Another major desert dust source, the Chinese deserts, shows similar seasonal behavior (T. Kitaoko, 1971, private communication). It is therefore of interest

to investigate its effect on the atmospheric energy balance. We believe that this may be done by studying the optical effects of Khamsins occurring over the Middle East area.

The Eastern Mediterranean is exposed during the April-June period to a weather phenomenon termed "Khamsin" or "Sharav." Hot and dry winds blow from the southwest, south or southeast, due either to thermal lows developing or intensifying over North Africa or to a deepening of the Sudanic trough. These winds transport large quantities of natural dust due to the nature and temperature of the desert surface and the characteristic unstable conditions over the latter. The travel time of such advective systems over the aerosol source region (which also includes part of Israel) is usually one day to a week during the months of April, May and June. The aerosol does not have much opportunity to mix with the other air masses so that its properties remain relatively pure. The process of formation of Khamsinic depressions and their associated weather systems were studied in detail by Elfandy (1940) and Sayed Ahmed (1949). Sudanic lows and some of their effects were treated by Ashbel (1938). A characteristic feature of a Khamsin is the milky-gray color of the sky. In some instances a bluish sun is observed.

One may assume that for the worldwide desert sources, the mechanisms of injection, transformation, transport and precipitation of the aerosols are similar in character to those of the Khamsinic aerosol.

In this study we therefore attempt to analyze quantitatively, as a first step in a program on the energetic

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effects of desert aerosols in the atmosphere, long series of solar spectral normal incidence measurements. These were taken by well and frequently calibrated Linke-Feussner actinometers at Jerusalem in the Judean Hills and at the Lydda Airport in the coastal plain.

2. Atmospheric turbidity associated with Khamsinic depressions

The dominant role of Khamsinic depressions in the seasonal pattern of turbidity and the latter's dependence on wavelength has been clearly identified recently (Joseph and Manes, 1971). Both the turbidity coefficients and their standard deviations reach their maximal values during the month of May, the period of highest frequency of occurrence of Khamsinic depressions. The minimum monthly average value of the wavelength exponent α also occurs during this period, showing that quite large particle sizes are frequent then. The turbidity maximum and the α minimum are superimposed, as shown in our previous work, on "typical" European patterns with extrema in summer (Ångström, 1964). The characteristic wide range of values attained by the turbidity parameters in cloud-free periods during Khamsinic weather is shown in Fig. 1, in contrast to other weather types (see also Joseph and Manes, 1971) as well as the relatively high values of atmospheric turbidity. The average value of the Ångström turbidity coefficient β_0 of the desert aerosol present during Khamsins is 0.13 and the standard deviation 0.06. This value is obtained when one discounts the extremely low turbidity frequently found at the onset of advective Khamsins. (During such periods the turbidity has a modal value of 0.015 or about half of that found during a clear winter day.)

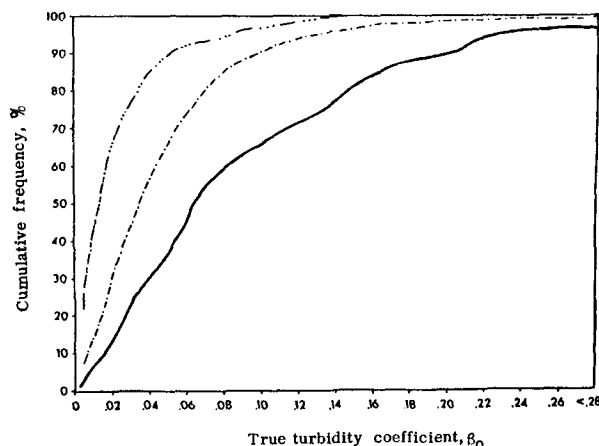


FIG. 1. Cumulative frequency distribution of the turbidity coefficient β_0 for various weather types in Jerusalem (1961-68): normal winter day (triple-dot curve), normal summer day (single-dot curve), advective-type Khamsin (solid curve).

A similar analysis of α for Khamsins in contrast to other typical weather systems is shown in Fig. 2. The modal value is 0.9, showing the presence of quite large particles during significant parts of the Khamsin phenomenon. Quite low values of α (<0.5) are also found. The latter may relate to the milky-gray color of the sky.

In order to use β_0 for an evaluation of the climatic effects of desert aerosol by using available theoretical studies (Yamamoto and Tanaka, 1972; Rasool and Schneider, 1971), it must be converted to the appropriate turbidity coefficient at $0.5 \mu\text{m}$ or to the appropriate optical depth at that wavelength as used in these studies. Since the optical depth at any wavelength λ (μm) may be written as $\tau = \beta_0 \lambda^{-\alpha}$, the optical depth for Khamsins at $0.5 \mu\text{m}$ is 0.24.

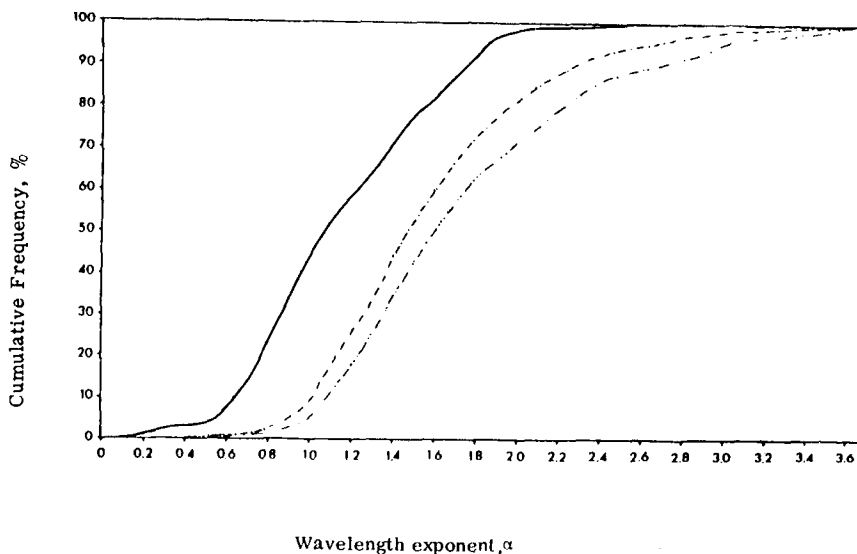


FIG. 2. As in Fig. 1 except for the wavelength exponent α .

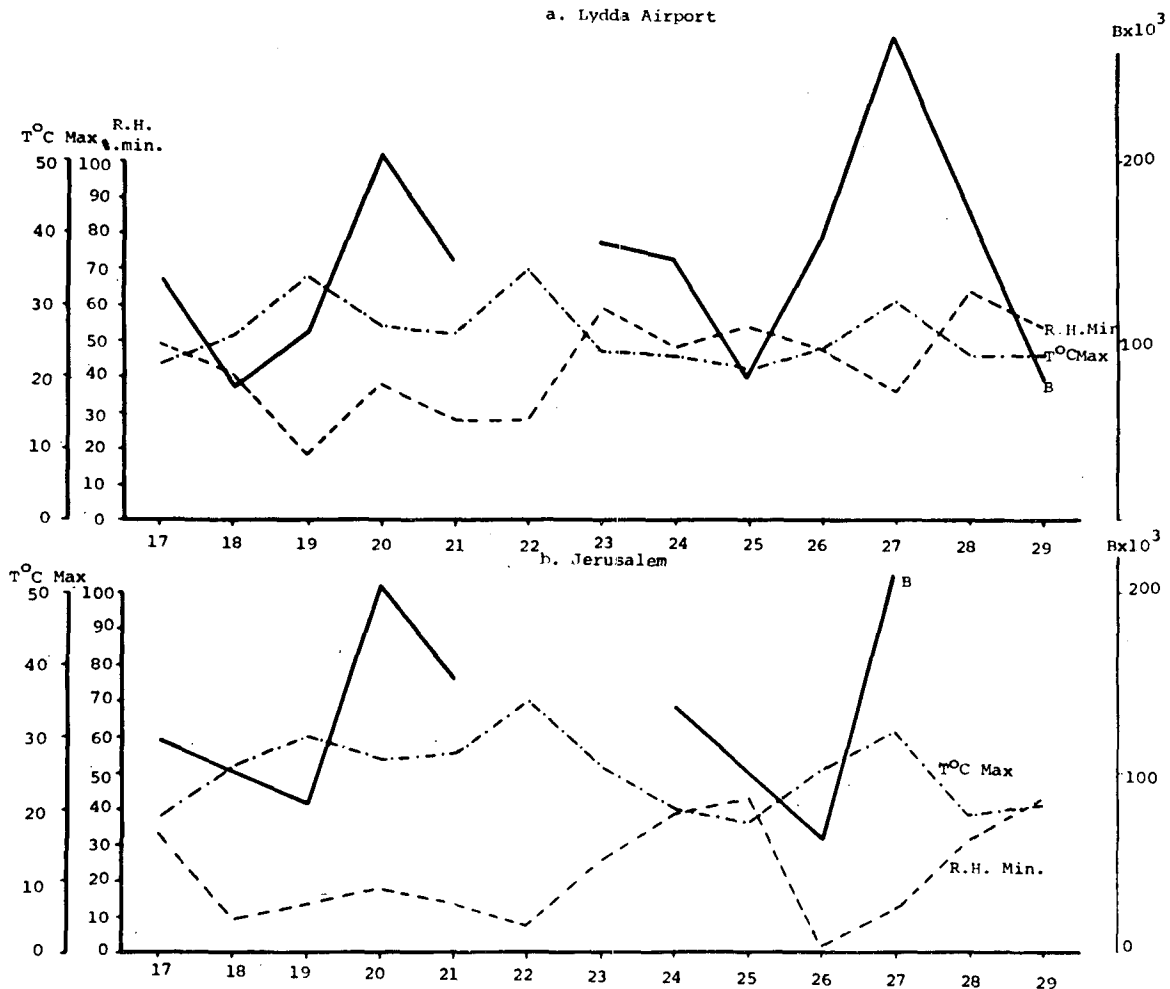


FIG. 3. Variations of the daily averages of the decadic turbidity coefficient B , relative humidity and temperature during the period 17-29 April 1959 at the Lydda Airport (a) and at Jerusalem (b).

3. The temporal variation of atmospheric turbidity in Khamsinic conditions

A characteristic set of daily averages of turbidity and other data during the passage of a series of Khamsins is shown in Fig. 3. The Khamsinic depressions on 18-20, 22-23, and 27 April 1959 resulted in a high and variable atmospheric turbidity throughout most of the period. The average monthly turbidity in April is 0.06. Values of turbidity of up to seven times the monthly average were reached and the average value of the turbidity during the Khamsin period is 0.17, or almost three times the monthly average which, of course, already includes the effect of Khamsins.

The synoptic weather situation for this period is described in the Israel Meteorological Service Monthly Weather Report for April 1959.

Typical weather maps for a Khamsinic period are shown in Fig. 4. A depression formed over western North Africa and traveled rapidly eastward over the desert. The Khamsin reached its maximal development

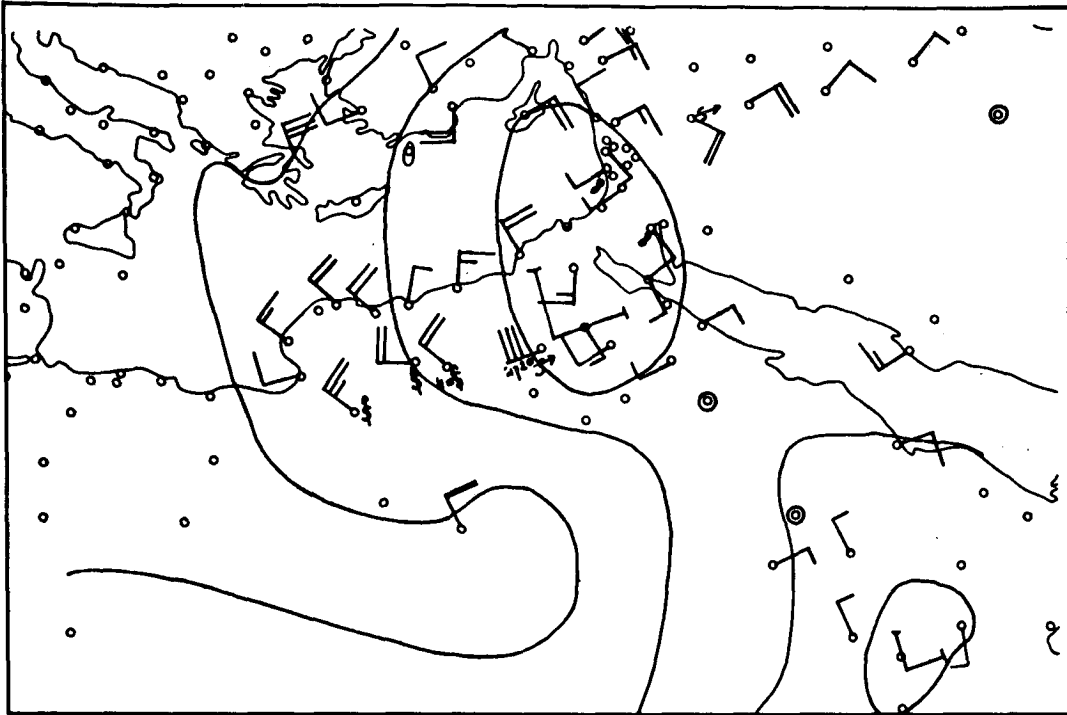
on 27 April 1959 and dust storms were reported throughout the affected region. The large desert area swept out by the depression is clearly evident as are the strong winds.

Satellite photographs can also indicate the presence of the dust cloud associated with Khamsinic depressions as shown by Ganor and Yaalon (1971). Typical diameters of such depressions are of the order of 500 km and they sweep over 1000-2000 km (Sayed Ahmed, 1949) so that a source area of 500,000 to 1,000,000 km² is indicated. The unstable conditions and high winds over the deserts are highly conducive to this whole area becoming a source for the traveling dust cloud. The duration of Khamsins is 1-7 days.

4. An estimate of the dust mass transported by Khamsins

According to Volz (1962), the mass M_1 of aerosol particles in the radius range $0.1 \lesssim r \lesssim 1.0 \mu\text{m}$, in an atmospheric column of 1 cm² cross section (whenever

a. Surface map



b. 850 mb

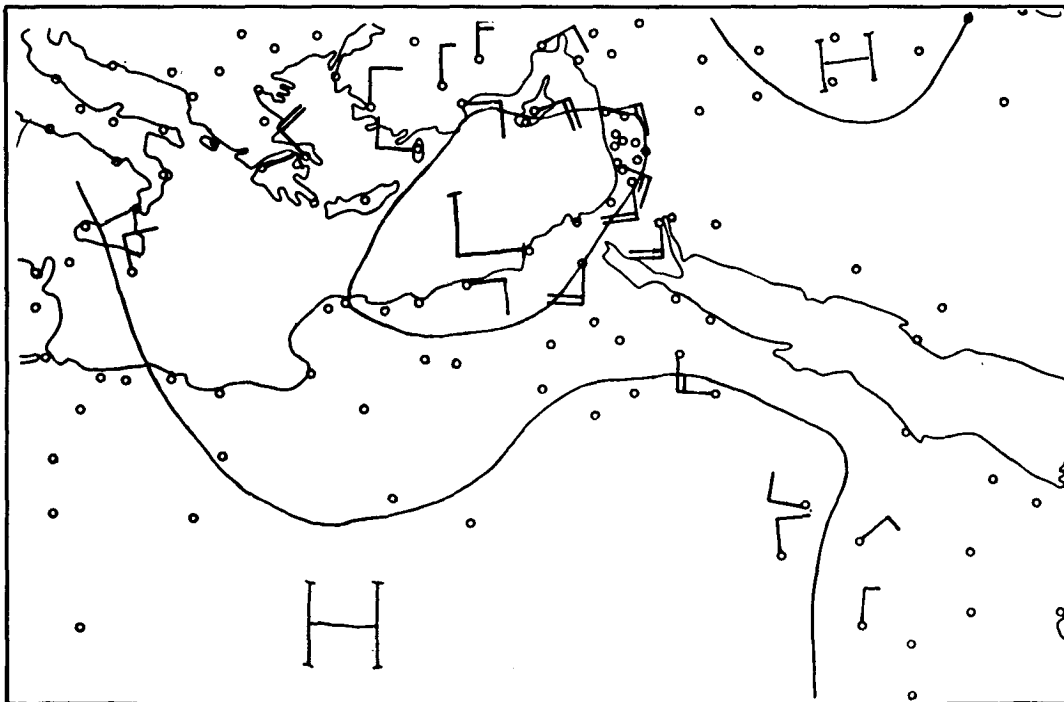


FIG. 4. Surface and 850-mb map for 1200 GMT 27 April 1959 (Eastern Mediterranean region).

the wavelength exponent α is close to unity, as in our case) can be determined by the linear relationship

$$M_1 = 55 \times 10^{-6} B \text{ [gm cm}^{-2}\text{]},$$

where B is the decadic turbidity coefficient.

Using this method, the total amount of particulate matter injected into the atmosphere by a Khamsinic depression during its life span can be estimated. Assuming the area swept out by a typical Khamsinic dust storm to be $\sim 1,000,000 \text{ km}^2$, we estimate that about 0.07 ± 0.03 million metric tons of particles within the range $0.1 \leq r \leq 1.0 \mu\text{m}$ are transported by a Khamsin. When one uses a Junge distribution indicated by the values of α during Khamsin periods, namely $dn/dr \propto r^{-4}$, and applies it between the sizes of 0.1 and $20 \mu\text{m}$ (S.M.I.C., 1971), the total mass M_{20} of such particles can be estimated as

$$M_{20} = \int_{0.1 \mu\text{m}}^{20 \mu\text{m}} dM = \int_{0.1 \mu\text{m}}^{20 \mu\text{m}} \frac{4\pi}{3} \rho r^3 dn(r),$$

where ρ is the mass density of the dust. Assuming the total mass between 0.1 and $1 \mu\text{m}$ to be given by M_1 , one finds that the total mass M_{20} between 0.1 and $20 \mu\text{m}$ is

$$M_{20} = \int_{0.1 \mu\text{m}}^{20 \mu\text{m}} dM = \frac{\ln\left(\frac{20}{0.1}\right)}{\ln\left(\frac{1}{0.1}\right)} M_1 = 2.3 M_1.$$

This gives about 0.16 ± 0.08 million metric tons in this size range. Particles with radii $> 20 \mu\text{m}$ are quickly removed from the atmosphere. Neglect of the giant particles here assumes no local dust contributions and might therefore result in an underestimate of the dust content by 20–30%. If one assumes 10 Khamsin episodes per year (Sayed Ahmed, 1949), one finds that the Mediterranean dust source area yields about 1.6 ± 0.8 million metric tons annually.

On taking the global areas of arid zones to be $40,000,000 \text{ km}^2$ (Flohn, 1971), or 8% of the earth's surface, and on assuming similar dynamic-synoptic processes to be responsible for the injection of dust into the atmosphere, one can deduce a dust input of 64 ± 32 million metric tons annually of particles $\leq 20 \mu\text{m}$. This is smaller than the value of 200 ± 100 as estimated by Flohn (1971), than the 500 estimated by Goldberg (1971) from the rates of accumulation in snow fields and from dust loads in winds and continental atmospheres, and also, less than the estimate of Peterson and Junge (1971).

The primary reason for this seems to be the simple extension of the Junge distribution from 1 to $20 \mu\text{m}$. Very preliminary data on the size distribution of aerosols during one Khamsin show that the number of

particulates with sizes $> 1 \mu\text{m}$ is at least two or three times that indicated by the assumption used here, and that the size spectrum peaks there. A very recent and excellent summary of aerosol size distributions (Junge, 1972) shows that the Saharan dust component in the oceanic aerosol has sizes ranging between 0.1 and $20 \mu\text{m}$ with a strong peak in its size distribution between 1 and $10 \mu\text{m}$. The number density at the peak is more than an order of magnitude larger than at either 1 or $10 \mu\text{m}$. If one uses these data as an indication of the real ratio of the mass between 0.1 and $20 \mu\text{m}$ compared to that between 0.1 and $1 \mu\text{m}$, one finds a value of the order of 2 larger. It may therefore be stated that a realistic lower bound on the mass of dust put into the atmosphere by Khamsinic phenomena is 128 ± 64 million metric tons annually.

5. Possible global effect on albedo and surface temperature

It might be informative to use the data presented in this study to get an approximate idea of the effect of this type of natural dust on the global albedo and therefore on the earth's average surface temperature. This can perhaps be done in the following manner because a large fraction of this type of aerosol will be distributed by the prevailing wind systems throughout all latitudes (Goldberg, 1971). If one assumes that only 8% of the areas of the globe is covered by such dust layers, then one can estimate an "average" global turbidity coefficient or optical depth at $0.5 \mu\text{m}$ due to such dust as 0.02 by multiplying the average β_0 at $0.5 \mu\text{m}$, i.e., 0.24 , by 0.08 .

Let us assume that the optical depth of such dust at $0.5 \mu\text{m}$ effectively determines its albedo. Following a current model computation (Yamamoto and Tanaka, 1972), if one assumes the dust to have a real refractive index of 1.5 and to be distributed with height in the atmosphere according to Elterman (1964), then the global albedo would be increased by ~ 0.002 due to the introduction of the desert aerosol into a clean ($\beta_0 = 0$) atmosphere with normal cloudiness overlying a surface where the continents have an albedo of 0.15 and the oceans 0.05 . In this case, the global average surface temperature could be decreased by about 0.7 K (Rasool and Schneider, 1971; Yamamoto and Tanaka, 1972), if there are no other phenomena to counteract this reduction. This estimate is obviously very uncertain and should be applied only in an informative way, pending its verification through the inclusion of this type of aerosol transport in numerical models of climate.

6. Conclusions

This analysis of long series of normal-incidence solar radiation data in Israel with particular emphasis on the yearly periods of Khamsins (or Sharav) dust storm occurrences, has brought into focus characteristic pat-

terns of atmospheric turbidity in the Eastern Mediterranean area. These include very high values of the turbidity parameters and low values of α , the wavelength exponent, indicating the presence of large particles. The yearly maximum of turbidity is during the months of frequent occurrence of such weather phenomena (spring). The areal extent of such weather systems is of the order of 1,000,000 km². These data enabled us to estimate the amount of dust injected into the atmosphere by such dust storms in one year. The lower bound on this quantity may be taken as 3.2 ± 1.6 million metric tons per year from the Mediterranean dust sources.

By assuming the same type of weather phenomena to be responsible for injection of dust into the atmosphere with about the same frequency, in all similar arid areas of the world, we estimated the yearly global total mass input of desert dust into the atmosphere to be at least 128 ± 64 million metric tons.

A tentative and approximate estimate of the global effect of such dust storms on the albedo of a dust-free but cloudy earth-atmosphere system gave an increase of 0.002, leading to a decrease of the average surface temperature by about 0.7K.

Another conclusion from this study is that for desert aerosol, the extension of a Junge distribution beyond 1 μm is not valid and leads to underestimates of the aerosol mass. It is hoped that a future "complete" study of the spectral radiation balance during Khamsin phenomena will result in data of sufficient detail and quality to be used in numerical simulations of climate.

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