The Ratio of Absorption to Backscatter of Solar Radiation by Aerosols during Khamsin Conditions and Effects on the Radiation Balance

Joachim H. Joseph[1]

National Center for Atmospheric Research,[2] Boulder, Colo. 80303

Noah Wolfson

Department of Geophysics and Planetary Sciences, Tel-Aviv University, Ramat-Aviv, Israel

(Manuscript received 14 March 1973, in revised form 25 March 1975)

ABSTRACT

Representative average values of the absorption to backscatter ratio of the aerosols in two typical synoptic conditions, namely cloudless khamsinic days and normal cloudless days, have been deduced from four years of data at two sites in Israel, one in the coastal plain and one in the Judean hills.

These values (15 for cloudless days and 11 for khamsinic days) have been compared to those, available in, or deducible from, published studies of the aerosol effects in various parts of the world.

It was found that the presence in the atmosphere of most of the aerosols induces a net heating effect, as far as the solar radiation balance is affected, over the regions in which they were measured.

We have further evaluated the complex part of the refractive index, averaged over the solar spectrum between 0.3 and 2.5 μm (0.03$\pm$0.02 for khamsinic days and 0.08$\pm$0.02 for normal days). This result is important because of the little data available for this parameter so far.

1. Introduction

Measurements conducted in the atmosphere always reveal the presence of aerosols (e.g., S.M.I.C., 1971; Porch *et al.*, 1970). These aerosols are optically active in the solar and infrared regions and may affect the radiation transfer in the thermal infrared. Their influence on the radiative balance of the earth-atmosphere system (EAS) might thus be appreciable. Surveys conducted at various sites around the world, and for long periods of time seem to indicate a slow average upward trend in the atmospheric load of aerosols, at least regionally (Manes, 1972; Joseph and Manes, 1971; Roosen *et al.*, 1973).

McCormick and Ludwig (1967) suggested that the rise of the atmospheric load of aerosols might be responsible for the cooling trend of the EAS, which has been observed since the 1940's (S.M.I.C., 1971). They proposed that the increase in the aerosol load of the atmosphere leads toward an increase of the backscattering of solar radiation to space and thus to a decrease in energy absorbed in the EAS. Charlson and Pilat (1969) drew attention to the fact that when trying to evaluate the effect of aerosols on the thermal balance, one should take into account that the aerosols absorb, as well as scatter, solar energy. They have presented a model for describing the solar energy balance under cloudless skies. It was found that the ratio a/b (a is the absorption of the aerosols, b is their "backscattering") and the surface albedo determine whether the aerosols will cause an increase or decrease in the solar energy absorbed by the EAS.

The present study concentrates on finding representative values for the ratio a/b in a semi-arid region—the Middle East. The spectrally averaged value (0.3–2.5 μm) of this parameter will be evaluated from available experimental data on solar radiation fluxes at two sites—Bet-Dagan near Tel-Aviv and Jerusalem for both cloudless days and cloudless khamsinic days (dry hot weather with desert winds).

The influence of various synoptic meteorological parameters on a/b has been examined. These various results are compared with similar data from many locations scattered over the globe.

The distinction between khamsinic and non-khamsinic cases is important because the dominant aerosol during the former is that of the desert. The desert aerosol is one of the most important natural aerosols and is effective on a global climatic scale (e.g., Junge 1972).

2. Theoretical background

From theoretical work (Mie, 1908; Van de Hulst, 1957; Ensor *et al.*, 1971), the ratio of the absorption co-

[1] On leave from the Department of Geophysics and Planetary Sciences, Tel-Aviv University, Ramat-Aviv, Israel.

[2] The National Center for Atmospheric Research is sponsored by the National Science Foundation.

efficient to the backscatter coefficient of an aerosol is given by

$$\frac{b_{abs}}{b_{bs}}=2\int_{r_1}^{r_2} p(a_t,b_t)n(r)dr \bigg/ \int_{r_1}^{r_2}\int_{\pi/2}^{\pi}[i_1(\theta)+i_2(\theta)]\sin\theta\, n(r)\,dr\,d\theta, \quad (1)$$

where a_t, b_t are the complex Mie amplitudes, $n(r)$ the size distribution of the aerosol, r_1, r_2 the limits of the optically active part of the aerosol distribution, and

$$\int_0^{\pi}[i_1(\theta)+i_2(\theta)]\sin\theta\, d\theta=\sum_{t=1}^{n}(2t+1)(|a_t|^2+|b_t|^2), \quad (2)$$

$$p(a_t,b_t)=\sum_{t=1}^{n}(2t+1)[\mathrm{Re}\,(a_t+b_t)-|a_t|^2-|b_t|^2]. \quad (3)$$

Thus in order to compute the ratio b_{abs}/b_{bs} it is necessary to know the complex refractive index and the size distribution. The size distribution is a variable quantity and the complex refractive index, and especially its imaginary part, is not very well known. The fact that these computations are strictly valid only for spherical aerosols is not a problem in the present context as the angularly integrated backscatter part of the phase function of real atmospheric aerosol is only about 30% smaller than that of a Mie sphere. The Mie sphere simulations are based on a sphere having the same surface area (Holland and Gagne, 1970, 1971).

Ensor *et al.* (1971) have published evaluations of b_{abs}/b_{bs} assuming a Junge-type size distribution of the aerosols and different refractive indices. We carried out a similar analysis using data on desert aerosols (Deirmendjian, 1969; Peterson and Weinman 1969; Lindberg and Laude, 1974; Dave and Braslau 1974).

In the present study, the ratio b_{abs}/b_{bs} is evaluated according to the technique suggested and developed by Robinson (1962) and used by him and others (Robinson 1962, 1966; Forbes and Hamilton, 1971; Drummond and Robinson, 1974).

Basically, the technique consists of measuring the total (global, G) and the diffuse downward (D) flux of solar radiation at the surface simultaneously with the upward flux of solar radiation above the aerosol layer. On subtracting the theoretically predicted values for a dustless atmosphere with the same temperature profile and amounts of gaseous absorbers, one is left with

$$\delta G=G(\text{clear})-G(\text{turbid}), \quad (4)$$

the deficit in global radiation, and

$$\delta D=D(\text{turbid})-D(\text{clear}), \quad (5)$$

the excess in downward (forward) diffuse radiation.

The deficit in global radiation, δG, is proportional to the true absorption a and the "backscatter" b of the solar flux due to the aerosol. The excess in diffuse radiation, δD, is proportional to the "forward scatter" f of the aerosol.

The ratio of the two measured quantities, δG and δD, is thus proportional to

$$\frac{\delta G}{\delta D}=\frac{a+b}{f}=\frac{b}{f}[(a/b)+1], \quad (6)$$

as shown in the literature (*ibid*).

Measurement of the upward flux U of solar radiation above the aerosol layer gives values of the backscatter b:

$$\delta U=U(\text{aerosol})-U(\text{clear})=b. \quad (7)$$

Knowledge of δG, δD and δU allows the evaluation of the absorption to backscatter ratio (a/b) and of the forward-to-backward scattering ratio (f/b), each as a function of solar zenith angle.

In the present case, only δG and δD were available. The ratio f/b, therefore, had to be evaluated separately. We compared experimental data (e.g., Robinson, 1962) with theoretical values computed by us for various types of aerosol.

The agreement between the measured and computed values is fairly good (Fig. 1). We decided, therefore, to use the experimental values of f/b as a function of solar zenith angle in order to be able to compare our results to previous studies (Robinson 1962, 1966; Drummond and Robinson 1974; Unsworth and Montieth, 1972).

Assuming f/b, one may derive the absorption to backscatter ratio, a/b, which for small optical depths becomes

$$\frac{a}{b}\approx\frac{b_{abs}}{b_{bs}}. \quad (8)$$

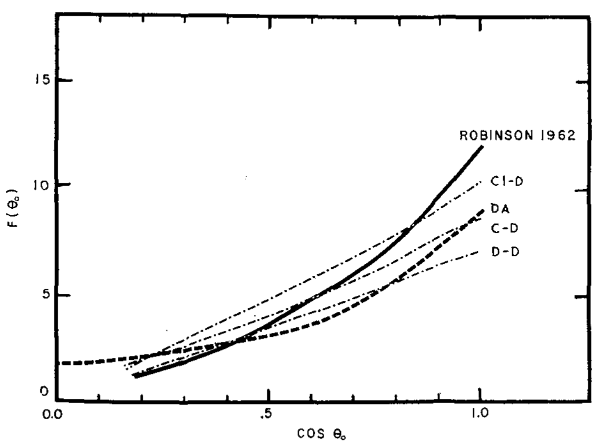

FIG. 1. The forward-to-backward scatter ratio: C–D, deduced from Type C (Dave and Braslau, 1974); C_1–D, deduced from Type C_1 (Dave and Braslau, 1974); D–D, deduced from Type D (Dave and Braslau, 1974); DA, desert aerosol (present); Robinson (1962) experimental data.

In the details of the calculations there are differences between this work and the original work of Robinson. In the calculation of the clear sky global radiation, our own model was used (Joseph, 1971) except for the computation of the ozone absorption which was done after Kennedy (1964).

3. Data selection and results

The inter-comparison of radiation measurements of the type used in this study is notoriously difficult. The reasons are that they are obtained under widely different conditions, calibrations, and, at times, by different instruments and observers (e.g., Drummond and Robinson, 1974). For instance, the presence of a thin cirrus deck, unobservable by a human observer, in an otherwise clear sky may seriously affect the results. In many cases, variations in humidity with time (solar zenith angle) will influence conclusions. Incorrect adjustment of or correction for the radiation obscured by the solar eclipse disk or incorrect positioning of the radiometer must be taken into account.

It is therefore imperative to examine very carefully the data to be used in a study such as the present one. For this study, data were available, not only on the global and diffuse, but also on the normal incidence (direct) and thermal infrared radiation at Bet-Dagan, and at Jerusalem, for the years 1967–1970. Both stations used an Eppley pyranometer for measuring the global radiation (G) and an Eppley pyranometer with a shading ring for measurement of the diffuse component of the solar radiation (D). The normal incidence (I) measurements were made with an Eppley pyrheliometer at Bet-Dagan and with one of the Linke-Feussner type at Jerusalem, and the thermal infrared fluxes with Gier and Dunkle radiometers. Enough data were thus available to select only those days and times on which all data were mutually consistent to within 5%.

Calibration of the systems is done once monthly at Bet-Dagan and once in three months at Jerusalem according to standard W.M.O. procedures.

The first stage of the selection of cloudless days was done by requiring that no clouds at all be reported in the synoptic observations at Bet-Dagan between 05 and 15 GMT. In this way 72 days were selected during the period 1967–1970 from several hundred available. Further selection of the cloudless days on which b_{abs}/b_{bs} was to be calculated from the radiation data for Bet-Dagan was carried out by examining the data in the following ways.

The global radiation was computed using

$$G' = I \cos\zeta + D. \tag{9}$$

This quantity is then compared with G, the global radiation measured by the pyranometer.

Whenever $|(G-G'/G)100|$ for any particular day was greater than 5% for more than half of the hourly measurements taken between 07 and 17 local time, that day was rejected. Thus, in this study, use was made of data from only 22 days from Bet-Dagan and 17 from Jerusalem from the available 72 cloudless days during the 4-year period. The 17 days at Jerusalem are included in the 22 days chosen for Bet-Dagan. Finally, only morning data were actually used (solar zenith angle between 80° and 55°). These measurements are most internally consistent owing to technical reasons.

The earth surface albedo was 0.259 for Bet-Dagan as measured by Dr. J. Stanhill (private communication, 1972) and 0.25 for Jerusalem as measured by Prof. D. Ashbel (private communication, 1972). Water vapor amounts in the atmosphere were either taken from radiosonde measurements taken at Bet-Dagan or, whenever these data were not available, computed by use of the following equation:

$$q(p) = aq_g\left(\frac{p}{p_g}\right)^b, \tag{10}$$

where $q(p)$ is the mixing ratio at pressure level p, q_g the ground level mixing ratio, a, b correlation coefficients, and p_g ground level pressure.

The correlation coefficients were calculated from ten years of available radiosonde data for each month.[3]

4. Results

A representative part of the results of our analysis of the radiation data are shown in Table 1. Our theoretical evaluations of a/b versus solar zenith angle are shown in Fig. 2 on the background of the experimental curves of a/b.

The variation in solar zenith angle is relatively small in our own data set. [Eighty percent of the data fall between $(0.70<\cos\zeta_0<0.90)$.] This is due to the fact that all measurements analyzed in the present study were taken between 07 and noon during late spring and summer at a subtropical latitude. The variability in a/b at each zenith angle from day to day is quite large in our data and also in those from other stations. The range of functional dependences of a/b on zenith angle as shown in Fig. 2 is typical. The three theoretical curves show the variation in a/b as computed by us for increase in absorption of the same desert aerosol model (to be published). The absorption is indicated by the values of the imaginary part of the refractive index ($K=0.01$, 0.02 and 0.2).

The heavy curve (present) is the best-fit a/b of all our observational data. The parts of this curve for $\cos\zeta_0<0.65$ and >0.95 are based on very few points and are therefore dashed. One standard deviation is indicated by the limits of the vertical line at $\cos\zeta_0=1$.

The best-fit curve is between the $K=0.01$ and $K=0.02$ lines between $0.65<\cos\zeta_0<0.93$. This is the region where most of our data are located. Considering

[3] J. Joseph, to be published.

TABLE 1. Sample of radiation data and results.*

Location	Date	$\cos\zeta_0$	G_C	G_M	D_C	D_M	$\lvert\delta G\rvert$	δD	b	s	a	a/s	a/b	Rayleigh	Water	f
JM	3/23/68	0.72	0.860	0.785	0.064	0.166	0.095	0.102	0.016	0.118	0.079	0.669	4.94	0.582	0.278	6.4
		0.82	0.874	0.790	0.059	0.190	0.084	0.131	0.016	0.147	0.068	0.463	4.250	0.592	0.282	8.2
		0.85	0.879	0.804	0.058	0.181	0.075	0.123	0.014	0.137	0.061	0.445	4.36	0.596	0.283	9
BD	3/23/68	0.72	0.838	0.749	0.089	0.067	0.124	0.057	0.008	0.063	0.081	1.28	10.12	0.575	0.264	6.4
		0.82	0.852	0.769	0.083	0.063	0.125	0.062	0.007	0.069	0.076	1.10	10.85	0.585	0.267	8.2
		0.85	0.857	0.784	0.073	0.061	0.120	0.059	0.006	0.065	0.067	1.03	11.16	0.589	0.269	9.0
JM	5/15/68	0.85	0.829	0.659	0.059	0.183	0.170	0.124	0.014	0.138	0.156	1.130	11.14	0.595	0.234	9.6
		0.94	0.841	0.661	0.056	0.165	0.180	0.109	0.010	0.119	0.170	1.429	17.00	0.604	0.237	10.6
		0.97	0.852	0.672	0.054	0.159	0.180	0.105	0.009	0.114	0.171	1.500	19.00	0.614	0.238	11.4
BD	5/15/68	0.84	0.811	0.669	0.142	0.061	0.147	0.086	0.009	0.095	0.133	1.400	14.77	0.588	0.223	9
		0.94	0.822	0.707	0.115	0.059	0.144	0.085	0.087	0.092	0.108	1.17	15.42	0.600	0.227	10.8
		0.97	0.826	0.733	0.093	0.058	0.139	0.081	0.007	0.088	0.086	0.97	12.28	0.600	0.227	11.4
JM	8/11/68	0.83	0.803	0.620	0.059	0.145	0.183	0.086	0.010	0.096	0.173	1.802	17.3	0.593	0.210	8.6
		0.92	0.815	0.653	0.056	0.131	0.162	0.075	0.007	0.082	0.155	1.890	22.14	0.602	0.213	10.4
		0.96	0.819	0.673	0.055	0.136	0.146	0.091	0.008	0.099	0.138	1.394	17.25	0.604	0.215	11.2
BD	8/11/68	0.83	0.790	0.651	0.062	0.175	0.139	0.113	0.013	0.126	0.126	1.00	9.69	0.586	0.203	8.6
		0.92	0.801	0.690	0.059	0.169	0.111	0.110	0.009	0.119	0.102	0.85	11.33	0.595	0.206	10.4
		0.96	0.805	0.708	0.058	0.163	0.097	0.105	0.009	0.114	0.088	0.77	9.77	0.600	0.208	11.2
JM	3/7/70	0.52	0.770	0.495	0.078	0.165	0.275	0.087	0.024	0.111	0.251	2.261	10.45	0.553	0.217	3.6
		0.66	0.797	0.597	0.071	0.156	0.200	0.085	0.016	0.101	0.184	1.822	11.50	0.573	0.224	5.5
		0.75	0.813	0.603	0.063	0.159	0.210	0.096	0.014	0.110	0.196	1.782	14.00	0.585	0.228	7
		0.79	0.819	0.650	0.061	0.173	0.169	0.112	0.014	0.126	0.155	1.230	11.07	0.589	0.230	7.8
BD	3/7/70	0.52	0.754	0.580	0.080	0.217	0.174	0.137	0.033	0.175	0.136	0.78	3.58	0.547	0.208	3.6
		0.66	0.783	0.651	0.070	0.217	0.132	0.147	0.022	0.173	0.106	0.61	4.08	0.567	0.246	5.5
		0.75	0.798	0.696	0.066	0.217	0.102	0.151	0.018	0.172	0.081	0.47	3.86	0.578	0.220	7.0
		0.79	0.803	0.738	0.064	0.220	0.065	0.156	0.017	0.176	0.045	0.25	2.25	0.582	0.221	7.8
JM	4/30/70	0.51	0.749	0.434	0.078	0.217	0.315	0.139	0.039	0.178	0.276	1.55	7.08	0.548	0.201	3.6
		0.69	0.786	0.561	0.066	0.224	0.225	0.158	0.027	0.185	0.198	1.07	7.33	0.576	0.210	5.9
		0.83	0.808	0.665	0.059	0.197	0.143	0.138	0.016	0.154	0.127	0.825	7.94	0.592	0.216	8.6
		0.92	0.821	0.655	0.056	0.187	0.166	0.131	0.013	0.144	0.153	1.063	11.77	0.602	0.219	10.4
		0.95	0.824	0.702	0.055	0.180	0.122	0.135	0.012	0.147	0.110	0.748	9.17	0.604	0.220	11.0
BD	4/30/70	0.51	0.738	0.551	0.187	0.080	0.204	0.124	0.027	0.158	0.153	0.91	5.66	0.540	0.143	3.6
		0.69	0.772	0.648	0.124	0.068	0.198	0.130	0.018	0.152	0.102	0.67	5.66	0.571	0.202	5.9
		0.83	0.793	0.686	0.107	0.062	0.190	0.128	0.019	0.142	0.093	0.65	4.89	0.586	0.207	8.6
		0.92	0.805	0.721	0.084	0.059	0.171	0.112	0.008	0.132	0.074	0.56	9.25	0.594	0.211	10.4
		0.95	0.809	0.738	0.071	0.058	0.165	0.107	0.007	0.126	0.062	0.49	8.86	0.598	0.212	11.0

* JM—Jerusalem; BD—Bet-Dagan; ζ_0—solar zenith angle; G_C—global calculated; G_M—global measured; D_C—diffuse calculated; D_M—diffuse measured; $\lvert\delta G\rvert$—deficit in global radiation; δD—excess in diffuse radiation; b—backscatter; s—total scatter; a—absorption; f—forward to backscatter ratio.

the large standard deviation, each of these values is about equally possible.

The distribution of our data in the a/b, $\cos\zeta_0$ plane is such that we tried also to evaluate an average value of a/b, the absorption-to-backscatter ratio at an average solar zenith angle of $\cos\zeta_0=0.864$, where the experimental forward-backward scatter ratio is 9.

The average values of the absorption-to-backscatter ratio for cloudless days in Bet-Dagan is

$$\frac{b_{abs}}{b_{bs}}=15\pm4,$$

and

$$\frac{b_{abs}}{b_{bs}}=11\pm3$$

for khamsinic cloudless days.

The results for Jerusalem are 13 ± 4 and 9 ± 3, respectively. The error in the estimates is based on errors of 15% in f/b, 5% in G and 10% in D.

The influence of local meteorological conditions on the ratio a/b was examined. It was found that low humidities (absolute and relative) lead toward lower values of a/b. Examining the effects of the wind direction, it was found that easterly winds (dry winds in our region) lead toward lower values of a/b.

By making use of Ensor's (Ensor *et al.*, 1971) calculation and assuming a Junge type size distribution for the aerosols

$$\frac{dn(r)}{dr}=Cr^{-(\gamma+1)},$$

where C and γ are constants and r is the radius of the aerosols, it is also possible to estimate the complex refractive index of the aerosols.

The value of 1.5 for the real part of the refractive index, n, of the desert aerosol, as used by Ensor *et al.*, is usually accepted (Bullrich, 1964; Guetta, 1973; Lindberg and Laude, 1974).

The value of 2 for γ is chosen in the khamsinic case because it leads to a neutral wavelength dependence of the optical depth (Junge, 1972; Joseph *et al.*, 1973; Carlson *et al.*, 1973). In the clear case, we use $\gamma=4$, which leads to a (-2) dependence of the optical depth on wavelength (Joseph and Manes, 1971).

The theoretical values are shown in Fig. 3 for Junge size distributions with $\gamma=2$, 3 and 4.

The following values were found for the imaginary part of the refractive index using these theoretical curves and our a/b ratios:

$$\kappa=\begin{cases} 0.03\pm^{0.02}_{0.01}, & \text{for khamsinic cloudless days} \\ 0.07\pm^{0.03}_{0.02}, & \text{for cloudless days.} \end{cases}$$

The values for the khamsinic case are not very different from what is indicated by Fig. 2.

The same procedure was followed for other size distributions for the same value of the real part, $n=1.5$ (silicate haze M, Deirmendjian, 1969; desert aerosol, Linberg and Smith, 1974; Lindberg and Laude, 1974). The data and the theoretical curves of a/b vs κ, the imaginary part of the refractive index at the peak of the solar spectrum, are shown in Fig. 3. It may be seen in this figure that the determination of the imaginary part of the index of refraction in this way is relatively insensitive to the differences between these same distributions, given the natural spread in the data.

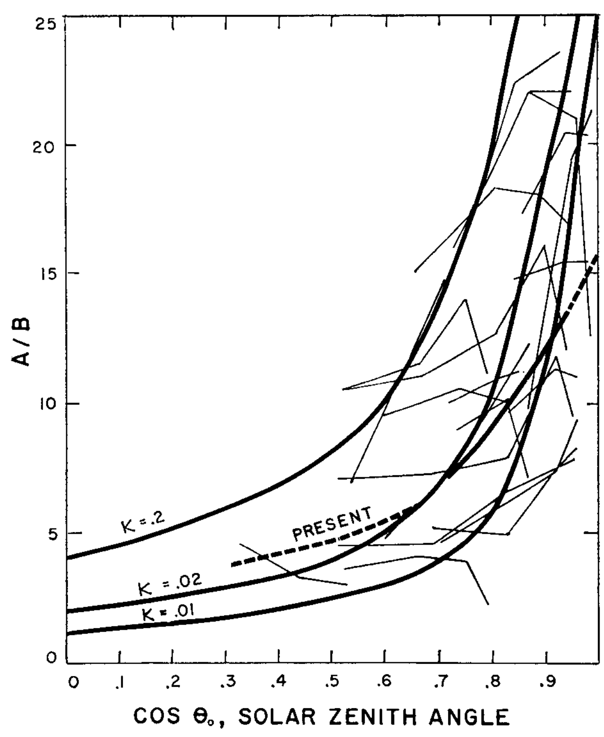

Fig. 2. Ratio a/b versus $\cos\zeta_0$, the solar zenith angle. The present best fit experimental results and theoretical results on a background of the experimental data.

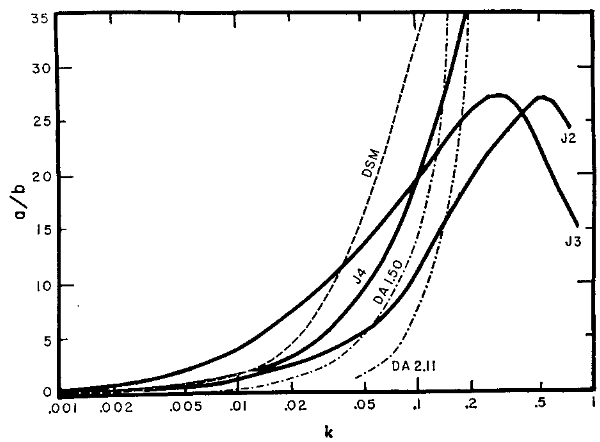

Fig. 3. Ratio a/b versus κ, the absorption index. J2, J3, J4 are Junge-type aerosol size distributions with $\gamma=2$, 3, 4. DSM, Deirmendjian silicate haze M; DA 1.50, DA 2.11, desert aerosol, $n=1.50$, 2.11.

All the experimental values of κ derived here are averaged over the solar spectrum between 0.3 and 2.8 μm, the spectral range of the global and diffuse radiation measurements. This is the reason they are slightly higher than the values found in the solar visible spectrum for the desert aerosol (e.g., Lindberg and Smith, 1974). Approximately the same ratio between the κ in the dusty and clear cases is found in the U.S. (Lindberg, 1974, private communication). The value deduced for khamsinic aerosols is probably more realistic as the relative humidity during such synoptic occurrences is well below 50% and the aerosol should act optically as a dry one. The value of κ found in the clear case is influenced by air pollution. The station at Bet-Dagan is then downwind of Tel-Aviv.

The determination of κ, the imaginary part of the refractive index, using the theoretical data in Figs. 2 and 3, is strictly valid only for solar radiation at 0.5 μm. Our results serve thus merely as an indication of a way to determine κ as a function of wavelength by future spectral determinations of a/b as a function of solar zenith angle. This is, in our opinion, the more preferable method, because the ultimate use for κ is in atmospheric energy balance modeling.

5. Application to the radiation balance and discussion

Determination of the influence of aerosols on the radiation balance in the shortwave region can be done by comparing the computed (a/b) or (b_{abs}/b_{bs}) with $(b_{abs}/b_{bs})_{crit}$, (where crit denotes the equilibrium ratio) as given by various models (Atwater, 1970; Charlson and Pilat, 1969; Schneider, 1971). Fig. 4, which is based

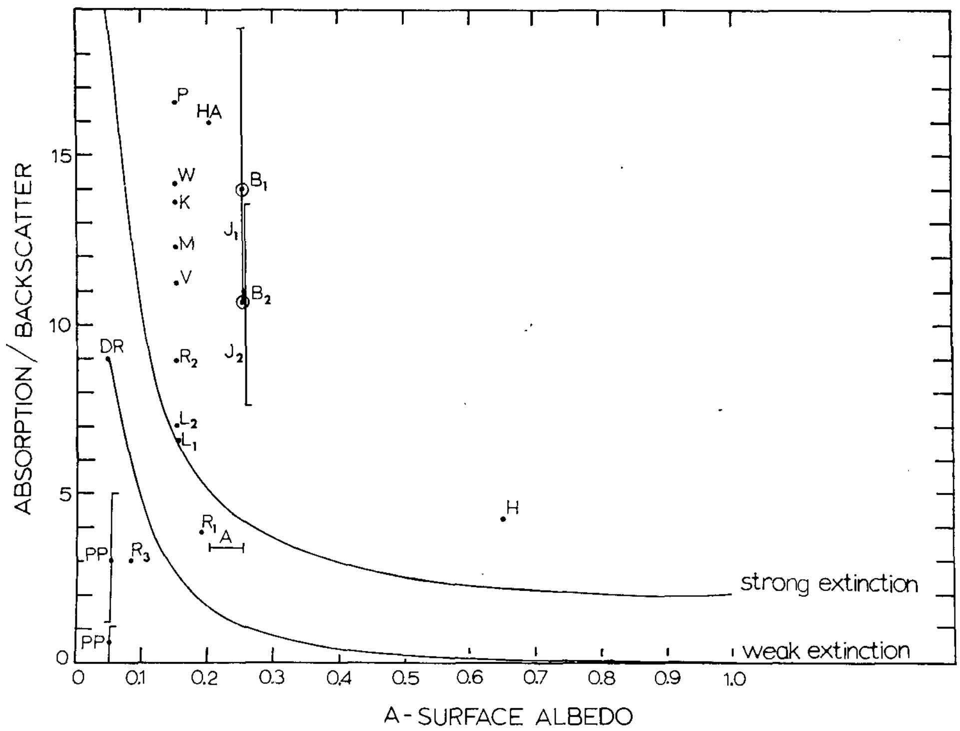

FIG. 4. Influence of b_{abs}/b_{bs} on the shortwave radiation balance as a function of the local surface albedo. (The equilibrium lines are deduced from Schneider, 1971).

A	Arizona, U. S. A. (Idso, 1972)—arid desert	M	Malta—arid
B_1	Bet-Dagan normal clear days—arid+urban	P	Pretoria—arid+urban
B_2	Bet-Dagan khamsinic clear days—desert	PP	Australia (Paltridge and Platt, 1973)—arid+agricultural
H	Halley Bay—desert—background	R_1	South England (Robinson, 1962)—urban
HA	Waldram, 1945—(Robinson, 1962)—urban	R_2	Robinson 1966 (hazy day)—urban
J_1	Jerusalem normal clear days—arid+urban	R_3	English Channel (Robinson, 1962)—urban
J_2	Jerusalem khamsinic clear days—desert	V	Vienna—urban
K	Kew—urban	W	Windhoek—arid+urban
L_1	Lerwick (Robinson, 1962)—maritime background	DR	West-Indies (Drummond and Robinson, 1974)—maritime+desert
L_2	Lerwick (Forbes and Hamilton, 1969)—maritime background		

on the model advanced by Schneider, enables one to decide in which way a certain aerosol, with a given b_{abs}/b_{bs}, will influence the radiation balance, given the surface albedo. The two curves for $(b_{abs}/b_{bs})_{crit}$ are for strong extinction (the upper) and for weak extinction (the lower) respectively. Strong extinction here means $b_{abs}+b_{bs}\approx 1$. The two curves serve to outline the range of possibilities for the position of the curve for $(b_{abs}/b_{bs})_{crit}$ in any particular case.

Fig. 4 presents results from Bet-Dagan and Jerusalem together with similar data from other parts of the world. It is seen clearly that most of the aerosols represented in this figure, including those of Bet-Dagan and Jerusalem, cause additional heating. However, if they were to move unchanged to regions with a lower surface albedo, the aerosols would induce a net cooling effect. High priority should thus be given to measuring the transport of various typical aerosols to oceanic regions and to estimating their effect on the radiation balance there. Only then would it be possible to make clearer statements on the global climatic effects of aerosols. In this context, it is interesting that the absorption-to-backscatter ratios measured by Robinson (1962, 1966), R_1, R_2, R_3 in Fig. 1, show a heating effect over land and a cooling effect over the English Channel. Also Paltridge and Platt's (1973) observations of a mixture of continental and sugar canefire-induced aerosols over the ocean most probably show a net cooling effect. Idso's (1972) measurements (A in Fig. 4) of the radiation fluxes over a desert region during a dust storm seem to indicate a neutral effect in the solar radiation balance but he claims that an important heating effect throughout the event (day and night) might occur in the thermal infrared as the downward radiation flux in the latter spectral region was increased by 15%.

The Middle Eastern aerosols, the influence of which on the radiation balance was evaluated in this study, are not of common origin. In addition to aerosols which enter the atmosphere due to local activities of man and nature, or by dust brought in aloft by desert winds (in which we were most interested), there might be aerosols of European origin in the atmosphere. Therefore, one has to understand the results given here as a comparison between the relative influence of the various types of aerosols, present on cloudless days and on khamsinic cloudless days, on the solar radiation balance.

From our results for Bet-Dagan, it seems that the atmospheric aerosol population during khamsinic days leads toward less absorption of solar energy than that during normal cloudless days. This conclusion is reinforced by our results for Jerusalem. As a result of less local urban activity in Jerusalem and its distance from the sea, the quantity of local urban aerosols is less than at Bet-Dagan. As a result of this, the ratios b_{abs}/b_{bs} for both heavy and light khamsins are smaller than those for Bet-Dagan. (Because of the small distance, 70 km, between both stations, it is possible to assume that higher-level aerosols, different from local desert and local urban, are present in the same quantities at both stations.) The difference between the two stations is in the population at the lower levels. It seems therefore that the urban or man-made agricultural aerosol raises the ratio b_{abs}/b_{bs}. This trend can also be traced in the data reported by Idso (1972), by Robinson (1962, 1966), by Paltridge and Platt (1973) and by Drummond and Robinson (1974), which are also presented in Fig. 4.

The fact that b_{abs}/b_{bs} is lower during khamsinic days is also connected to the fact that aerosols contain water. Covert *et al.* (1972) have shown that Q_{bs} (Q_{bs} is the backscattering efficiency factor) of most aerosols is almost independent of the relative humidity up to values of the latter of 60%. Hanel (1972) found that Q_{ext} (Q_{ext} is the extinction efficiency factor) does not change much with relative humidity until the latter reaches about 60%. Both Q_{abs} and Q_{bs} (Q_{abs} is the absorption efficiency factor) grow with rising relative humidity but Q_{abs} grows faster. Thus, in order to examine the optical behavior of the dry aerosol, one has to eliminate the influence of the relative humidity, if it rises above 50–60%. It seems even possible that an aerosol population over a desert region will exhibit a diurnal cycle of b_{abs}/b_{bs} as a result of a pronounced diurnal cycle of relative humidity.

Therefore the values computed for the complex part, κ, of the refractive index ($0.03^{+0.02}_{-0.01}$ for khamsinic aerosol and $0.07^{+0.03}_{-0.02}$ for cloudless days) are the maximal (for dry atmospheric spherical aerosols) possible ones, especially since the available data are for the whole integrated solar spectrum. This latter fact, however, makes the values found for κ in the present study of much practical interest for climatic studies as they indicate the actual physical effect of the aerosol.

It is instructive to mention here that in a recent theoretical study (Yamamoto and Tanaka, 1972) it is concluded that a Junge-type aerosol with its optical depth wavelength coefficient α equal to 1, with a real part of the refractive index equal to 1.50 and located in a Rayleigh atmosphere, leads to a net heating effect, if the imaginary part of the refractive index is larger than 0.05 over the solar spectrum, and if the results are suitably weighted for the global distribution of surface albedo.

The values of the absorption to backscatter, as summarized in Fig. 4, show that many of the aerosols represented there, if located over surfaces with a low albedo, should bring on a net cooling if their total optical depth is large enough. Almost all of the aerosols, if located over land, would bring on a net heating. The khamsinic desert aerosol might, if the spectrally averaged value of the imaginary part of its refractive index is larger than 0.05, be neutral in its global heating effect.

The data used for this study of desert aerosol in the Middle East were carefully chosen from two independent stations active over a long period of time. It seems possible, therefore, to accept the results as representative for the aerosol population in our region. However, one must remember that there is a bias in the results toward aerosols which are present in special weather conditions.

In order to understand the climatic influence of the aerosol and its effect on surface and air temperature, it is not enough to know their influence on the radiative shortwave balance. The aerosol is also active in the windows of the thermal infrared spectrum. Furthermore, in the atmosphere there are many feedback processes and the interaction between the radiative and hydrodynamical processes is very complex. In order to evaluate the aerosol effects, one has to use either global or regional climatic models which include both hydrodynamical and radiative heat transfers.

Acknowledgments. The radiation data supplied by Prof. Dov Ashbell are gratefully acknowledged.

REFERENCES

Atwater, M. A., 1970: Planetary albedo changes due to aerosols. *Science*, **170**, 64–66.

Bullrich, K., 1964: Scattered radiation in the atmosphere and the natural aerosol. *Advances in Geophysics*, Vol. 10, New York, Academic Press, 99–260.

Carlson, T. N., J. M. Prospero and K. J. Hanson, 1973: Attenuation of solar radiation by windborne Saharan dust off the west coast of Africa. NOAA Tech. Memo. ERL WMPO-7, 27.

Charlson, R. J., and M. J. Pilat, 1969: Climate: The influence of aerosols. *J. Appl. Meteor.*, **8**, 1001–1002.

Covert, D. S., R. J. Charlson and N. C. Ahlquist, 1972: A study of the relationship of chemical composition and humidity to light scattering by aerosols. *J. Appl. Meteor.*, **11**, 968–976.

Dave, J. V., and N. Braslau, 1974: Effect of cloudiness on the transfer of solar energy through realistic model atmospheres. I.B.M. Res. Rept. 4869, pp. 55.

Deirmendjian, D., 1969: *Electromagnetic Scattering of Spherical Poly-Dispersions*. New York, American Elsevier, 290 pp.

Drummond, A. J., and G. D. Robinson, 1974: Some measurements of the attenuation of solar radiation during BOMEX. *Appl. Opt.*, **13**, 487–492.

Ensor, D. S., N. W. Porch, M. J. Pilat and R. J. Charlson, 1971: Influence of the atmospheric aerosol on albedo. *J. Appl. Meteor.*, **10**, 1303–1306.

Forbes, A. J., and R. A. Hamilton, 1971: Clear sky measurements at Lerwick. *Quart. J. Roy. Meteor. Soc.*, **97**, 99–102.

Guetta, D., 1973: The optical depth of atmospheric aerosols and water vapor under two cloudless weather conditions in an area far from cities. M. S. thesis, Tel-Aviv University, 84 pp.

Hanel, G., 1972: Computation of the extinction of visible radiation by atmospheric aerosol particles as a function of the relative humidity, based upon measured properties. *Aerosol Sci.*, **3**, 377–386.

Holland, A. C., and G. Gagne, 1970: The scattering of polarized light by polydisperse systems of irregular particles. *Appl. Opt.*, **9**, 1113–1123.

——, and ——, 1971: Comment on: The scattering of polarized light by polydisperse systems of irregular particles. *Appl. Opt.*, **10**, 1173–1174.

Idso, S. B., 1972: Radiation fluxes during a dust storm. *Weather*, **27**, 204–208.

Joseph, J. H., 1971: On the calculation of solar radiation fluxes in the troposphere. *Solar Energy*, **13**, 251–261.

——, and A. Manes, 1971: Secular and seasonal variations of the atmospheric turbidity in Israel at Jerusalem. *J. Appl. Meteor.*, **10**, 453–462.

——, —— and D. Ashbel, 1973: Desert aerosols transported by khamsinic depressions and their climatic effects. *J. Appl. Meteor.*, **12**, 792–797.

Junge, C. E., 1972: Our knowledge of the physics of aerosols in the undisturbed marine environment. *J. Geophys. Res.*, **77**, 5183–5200.

——, 1974: The possible influences of aerosols on general circulation and climate and possible approaches for modelling. *The Physical Basis of Climate and Climate Modelling*, Report of the GARP Study Conf., Stockholm, 29 July–10 August, 1974 (in press).

Kennedy, J. S., 1964: Energy generation through radiative processes in the lower stratosphere. Planetary Circulations Project Report No. 11, Dept. Meteor., M.I.T.

Lindberg, J. D., and L. S. Laude, 1974: Measurement of the absorption coefficient of atmospheric dust. *Appl. Opt.*, **13**, 1923–1927.

——, and M. S. Smith, 1974: Visible and near infrared absorption coefficients of kaolinite and related clays. *Amer. Miner.*, **59**, 274–279.

Manes, A., 1972: Atmospheric turbidity over Jerusalem. M. Sc. thesis, Tel-Aviv University, The Israeli Meteorological Service, Ser. A, Meteorological Notes No. 26, 6, 7, 13, 61.

McCormick, R. A., and J. H. Ludwig, 1967: Climate modification by atmospheric aerosols. *Science*, **156**, 1358–1359.

Mie, G., 1908: Beitrag sur Optik trueber Medien. *Ann. Phys.*, **25**, 377–445.

Paltridge, G. W., and C. M. Platt, 1973: Absorption and scatter of radiation by an aerosol layer in the free atmosphere. *J. Atmos. Sci.*, **30**, 734–737.

Peterson, J. T., and J. A. Weinman, 1969: Optical properties of quartz dust particles at infra-red wavelengths. *J. Geophys. Res.*, **74**, 6947–6952.

Porch, W. M., R. J. Charlson and L. F. Radke, 1970: Atmospheric aerosol: Does a background level exist? *Science*, **170**, 315–317.

Robinson, G. D., 1962: Absorption of solar radiation by atmospheric aerosol, as revealed by measurements from the ground. *Arch. Meteor. Geophys. Bioklim.*, **B12**, 19–40.

——, 1966: Some determinations of atmospheric absorption by measurement of solar radiation from aircraft and at the surface. *Quart. J. Roy. Meteor. Soc.*, **92**, 263–269.

Roosen, R. G., R. J. Angione and C. J. Klemcke, 1973: Worldwide variations in atmospheric transmission: Baseline results from Smithsonian observations. *Bull. Amer. Meteor. Soc.*, **54**, 307–316.

Schneider, S. H., 1971: A comment on "Climate: The influence of aerosols." *J. Appl. Meteor.*, **10**, 840–842.

S.M.I.C., 1971: *Study of Man's Impact on Climate.* Cambridge, The M.I.T. Press, pp. 308.

Unsworth, M. H., and J. L. Monteith, 1972: Aerosol and solar radiation in Britain. *Quart. J. Roy. Meteor. Soc.*, **98**, 778–797.

Van de Hulst, J. C., 1957: *Light Scattering by Small Particles.* Wiley, 114–130.

Yamamoto, G., and M. Tanaka, 1972: Increase of global albedo due to air pollution. *J. Atmos. Sci.*, **29**, 1405–1412.