

## Application of Soil Dust Optical Properties in Analytical Models of Climate Change

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### ABSTRACT

Several analytical models of the radiative effects of aerosol layers on global climate provide the common result that the critical value ( $\rho_c$ ) of the ratio ( $\rho$ ) of aerosol layer absorption to hemispheric backscattering is given by

$$\rho_c = \frac{(1-A)^2}{2A},$$

where  $A$  is taken to be the albedo of the earth's surface, or of the present earth-atmosphere system. The models predict that introduction of a new aerosol layer with  $\rho > \rho_c$  will cause a decrease in system albedo, and a layer with  $\rho < \rho_c$  will cause an increase. In this paper we demonstrate this common result for  $\rho_c$  and then employ recently published data on the complex refractive index and size distribution of atmospheric surface layer soil particles to compute values of  $\rho$ . The resulting values ( $5 < \rho < 28$ ) are quite large compared to previous estimates. Together with the above model result they indicate that increased generation of such airborne soil particles will tend to increase the input of solar energy to the earth-atmosphere system. This "heating" effect results, in part, from the relatively large mean particle sizes used in the computations. The effects of particle asphericity on the computed  $\rho$  values are discussed.

### 1. Introduction

In recent years several authors (e.g., McCormick and Ludwig, 1967; Bryson, 1968) and groups (SCEP, 1970; SMIC, 1971) have expressed concern that increased aerosol pollution could modify the transfer of radiant energy between the sun and the earth in such a way as to significantly alter the earth's climate. To assess this possibility in a quantitative manner, a number of other investigators (e.g., Charlson and Pilat, 1969, 1971; Atwater, 1970; Ensor *et al.*, 1971; Rasool and Schneider, 1971; Schneider, 1971; Mitchell, 1971; Barrett, 1971; Neumann and Cohen, 1972; Yamamoto and Tanaka, 1972; Braslau and Dave, 1973a, b; Chýlek and Coakley, 1974; Shettle and Green, 1974) have developed models to describe radiative energy exchange in the earth-atmosphere system, and the effects that increased aerosol content have on this exchange. These models have ranged from very approximate formulations which provide simple analytical results to more detailed and realistic treatments which require complex numerical solution procedures. For the most part, however, application of all of these models to actual climatic prediction has been severely limited due to a lack of

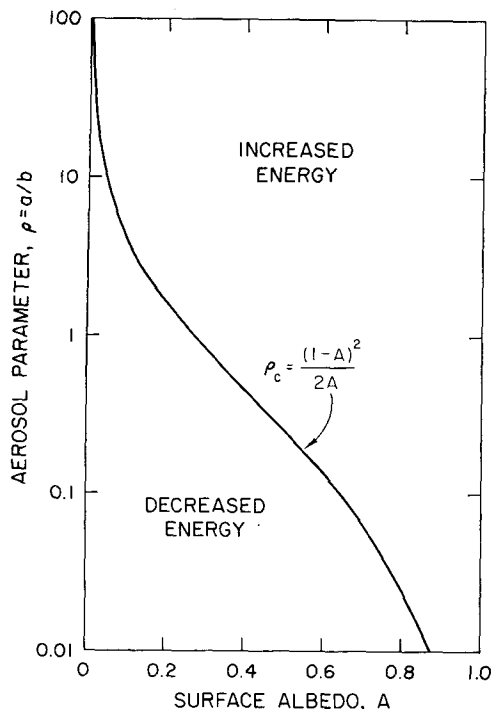
realistic input data that characterize the atmospheric aerosol.

The purpose of this paper is to point out a common result of some of the simpler models, and then to apply this result by using as input data some measurements of the complex refractive index and size distribution of atmospheric surface-layer soil particles recently reported by Grams *et al.* (1974).

### 2. Analytical models of aerosol climatic change

The analytical models of climate change due to aerosols begin by approximating the present (unperturbed) earth-atmosphere system as having a plane-parallel geometry with a system albedo  $A$ . A new plane-parallel aerosol layer is then introduced between the present system and the sun, and the albedo  $A'$  of the new system (including the new layer) is computed by some approximate means. (Alternatively,  $A$  is often taken to be the albedo of the earth's surface. The new aerosol layer is then introduced at an arbitrary height in the present atmosphere, and the present atmosphere is treated as perfectly transmitting. Mathematically, these two formulations are completely equivalent.) If  $A' > A$ , the prediction of the model is that the new aerosol layer will reduce the energy of the present earth-

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- FIG. 1. Relationship of critical aerosol parameter  $\rho_c$  to surface albedo  $A$  as determined by several analytical models of the radiative effects of an optically thin aerosol layer.

atmosphere system, whereas if  $A' < A$ , the net energy will increase. These results are generally referred to as "cooling" or "heating," respectively, although it is recognized that the energy change may include dynamic as well as thermal effects.

Many of the analytical models begin by characterizing the new aerosol layer by the macroscopic parameters  $a$  and  $b$ , which are, respectively, the fraction of incident radiation absorbed and that backscattered (i.e., scattered into the backward hemisphere) by the layer as a whole. In this approach the problem is effectively reduced to the familiar optics problem of evaluating the reflection coefficient of a system of two glass plates with individual reflection coefficients  $R = b$  and  $R_0 = A$ , and transmissivities,

$$t = 1 - a - b \quad \text{and} \quad t_0 = 0, \tag{1}$$

respectively. By summing the series of reflections between the two layers, as done, for example, by Rasool and Schneider (1971) and by Schneider (1971), one can readily show that the albedo of the perturbed system is given by

$$A' = b + A \left( \frac{t^2}{1 - Ab} \right). \tag{2}$$

Inspection of (2) reveals that the three conditions  $A' > A$ ,  $A' = A$ , and  $A' < A$  are all possible, depending on the aerosol layer characteristics  $a$  and  $b$  (hence  $t$ ) and the unperturbed albedo  $A$ . The borderline condition, in

which the aerosol layer will have neutral effect on system energy, is obtained by equating  $A'$  to  $A$  in (2). The resulting equation may then be solved to obtain critical parameters characterizing such a neutral aerosol layer. For this purpose it is useful to eliminate the aerosol layer parameters  $a$  and  $b$  in favor of two other parameters which also completely characterize the layer. These parameters are the layer transmissivity  $t$ , defined in (1), and the layer absorption-to-backscatter ratio  $\rho$ , defined by

$$\rho \equiv a/b. \tag{3}$$

The definitions (1) and (3) provide the substitution equation

$$b = \left( \frac{1-t}{1+\rho} \right). \tag{4}$$

By substituting (4) into (2), setting  $A' = A$  and  $\rho = \rho_c$  (the critical, or neutral, value), one obtains

$$\rho_c = \frac{(1-A)^2 - 2At + [(1-A^2)^2 + 4A^2t^2]^{\frac{1}{2}}}{2A(1+t)}, \tag{5}$$

in which the extraneous root ( $\rho_c < 0$ ) has been eliminated. Incremental analysis using (2) shows that a layer with ratio  $\rho > \rho_c$  will tend to increase the energy of the earth-atmosphere system (i.e.,  $A' < A$ ), whereas if  $\rho < \rho_c$ , the result will be the opposite, and a decrease of system energy will ensue.

Although (5) is valid for arbitrary layer transmissions  $t$ , we shall restrict our subsequent discussion to the case of an optically thin layer, for which  $t \rightarrow 1$ . In this limit, (5) reduces to

$$\rho_c = \frac{(1-A)^2}{2A}, \tag{6}$$

which indicates the significant result that the sign of the system albedo change caused by an optically thin aerosol layer is determined entirely by its ratio of absorption to backscatter and the value of the surface (or unperturbed system) albedo. The result (6) is plotted in Fig. 1, and values of  $\rho_c$  for some typical surface types, are listed in Table 1. To determine just how optically thin a layer must be for (6) to be accurate, we have evaluated (5) as a function of  $t$  for the values of  $A$  listed in Table 1. The values of  $\rho_c$  given by (5) increase as  $t$  decreases. However, for the range of surface albedos listed, (6) is accurate to within 10% for any layer with  $t > 0.9$ . Eqs. (4) and (2) show that such layers with  $0 < \rho \leq 10$  can change the earth-atmosphere albedo by several percent or more. Since it has been estimated (e.g. Sellers, 1969, 1973) that a 2% change in the amount of solar energy reaching the earth can cause changes of several degrees Celsius or more in global average temperature, and major alterations in global snow cover, it is evident that the simple result (6) is applicable to climatically significant aerosol layers. We

TABLE 1. Critical values of the ratio  $\rho_c$  of aerosol layer absorption to backscatter as obtained in several analytical models of climate change.

Surface type	Albedo $A$	Sensible heating index $C$	Total system energy $\rho_c = \frac{(1-A)^2}{2A}$	Critical ratio $\rho_c = a/b$ for	
				Surface temperature (Mitchell) <sup>a</sup> $\rho_c = \frac{C(1-A)(1-KA)}{D(1+A) - C(1-A)}$	
Urban areas	0.20	0.80	1.6	$D=1.0^b$ 0.91	$D=0.75^c$ 1.97
Deserts	0.30	0.95	0.82	0.73	1.50
Prairies and farmlands	0.20	0.40	1.6	0.29	0.44
Forests	0.16	0.30	2.2	0.23	0.34
Oceans	0.08	0.09	5.3	0.08	0.10
Snowfields	0.70	0.09	0.064	0.005	0.006

<sup>a</sup>  $K$  is assumed equal to 1.

<sup>b</sup> Typical of an aerosol generated (e.g., by man) at the surface.

<sup>c</sup> Typical of a general tropospheric aerosol.

note that (6) was not explicitly derived by Rasool and Schneider (1971) or by Schneider (1971), but that it follows readily from their result (2).

A climatological model which differs somewhat from that of Rasool and Schneider was devised by Mitchell (1971). It considers individual energy absorption terms in the atmosphere and at the surface to describe the effect on a "synoptic" surface region caused by a layer of atmospheric aerosol directly overhead. The model takes into account the fact that radiation incident on a moist surface region results, at least partially, in evaporation and that the latent heat thus produced is transformed to sensible heat at another time or location or both.

Mitchell explicitly derives a critical ratio  $\rho_c$  which applies to the sensible heating or cooling of a synoptic region at the surface. His result, which is valid for optically thin layers, is

$$\rho_c = \frac{C(1-A)(1-KA)}{D(1+A) - C(1-A)}, \tag{7}$$

in which  $C$  is the ratio of sensible to total heating at the surface;  $D$  is the fraction of the aerosol layer in convective contact with the surface; and  $K$  is the ratio ( $b_s/b_i$ ) of the aerosol backscatter fraction  $b_s$  for sunlight reflected from the surface to the aerosol backscatter fraction  $b_i$  for sunlight incident from above. As in Mitchell's paper,  $K$  will be taken to be equal to unity in the following discussion; this assumption is also implicit in Rasool and Schneider's derivation of the effective albedo  $A'$ .

The critical ratios given by Mitchell's result (7) are listed in Table 1 for a tropospheric aerosol layer, with the previously listed values of  $\rho_c$  given by (6). Comparison of the results shows that, in considering the sensible heating of a surface region, inclusion of latent heating effects generally leads to lower values of  $\rho_c$  than are predicted by (6). However, as noted by Mitchell,

in considering the effect of an aerosol layer on total system energy, the proportions of latent and sensible heating are immaterial, and so is the amount of convective contact between the layer and the surface. Accordingly, in considering effects on the total energy of the earth-atmosphere system, we may take  $C = D = 1$  in (7), thus obtaining

$$\rho_c = \frac{(1-A)^2}{2A}. \tag{8}$$

This result, which was obtained by Mitchell, is identical to the result (6) which we have derived from the result (2) of Rasool and Schneider.

Finally, we mention a third analytical model of aerosol climatological effects. Chýlek and Coakley (1974) also consider a plane-parallel aerosol layer introduced above a surface with albedo  $A$ . Rather than working with macroscopic properties of the isolated layer, Chýlek and Coakley explicitly solve the equation of radiative transfer within the layer by using microscopic aerosol properties and a two-stream approximation to describe internal intensities. In this manner they derive, as a critical ratio for unchanged system albedo,

$$\frac{1-\omega}{\omega\beta} = \frac{(1-A)^2}{2A}, \tag{9}$$

where  $\omega$  is the single-scattering albedo and  $\omega\beta$  is the fraction of radiation scattered into the backward hemisphere. The parameters  $\omega$  and  $\beta$  are descriptive of a single-scattering polydispersion of the particles which comprise the layer, and do not necessarily describe the layer as a whole. However, as noted by Chýlek and Coakley (1974), the result (9) is accurate only for optically thin layers. For such a layer, the macroscopic layer parameters  $a$  and  $b$  are obtained by direct integration of single-particle absorption and hemispheric

backscatter cross sections over the layer polydispersion. Thus

$$\frac{a}{b} = \frac{1-\omega}{\omega\beta}, \quad (10)$$

and the two-stream result for the critical ratio of an optically thin layer becomes

$$\rho_c = \frac{(1-A)^2}{2A} \quad (11)$$

which is identical to (6) and (8).

### 3. Absorption-to-backscatter ratios for soil-dust aerosols

A recent publication by Grams *et al.* (1974) presents optical properties of airborne soil particles determined by analysis of simultaneous light-scattering and particle-sampling measurements made in the atmospheric surface layer. The determined optical properties included the real and imaginary parts of the complex refractive index

$$m = n_{\text{Re}} - n_{\text{Im}}i, \quad (12)$$

as well as the size distribution of the sampled particles. The size distributions, as determined by laboratory sizing of collected particles (Blifford and Ringer, 1969; Blifford and Gillette, 1971, 1973; Gillette and Blifford, 1971), were found to have a log-normal form, i.e.,

$$n(r) = \frac{0.434N_p}{r(2\pi)^{\frac{1}{2}} \log\sigma_g} \exp\left[-\frac{(\log r - \log r_g)^2}{2 \log^2\sigma_g}\right], \quad (13)$$

where  $n(r)dr$  is the number of particles with radii between  $r$  and  $r+dr$ ,  $N_p$  is the total number of particles, and  $r_g$  and  $\sigma_g$  are the geometric mean and standard deviation of the size distribution, respectively.

The value of  $n_{\text{Re}}$  was determined by standard microscopic analysis of the sampled particles. Although a wide range of values was observed,

$$n_{\text{Re}} = 1.525 \quad (14)$$

was found to be representative. The value of  $n_{\text{Im}}$  was determined from measurements of the angular variation of light scattered by the particles in the following manner. By applying Mie scattering theory to each of the observed distributions of particle size, the expected angular variation of the intensity of the scattered light was calculated for a fixed value of  $n_{\text{Re}}$  and for a wide range of values of  $n_{\text{Im}}$ . For each set of simultaneous light-scattering and particle-collection measurements, the value of  $n_{\text{Im}}$  was taken to be that value which provided the best fit to the experimental data. Thus, while it was often found that the sampled particles were not the homogeneous spheres frequently assumed in radiative transfer calculations, Mie scattering computations of the angular variation of light scattered by soil-dust

aerosols using the complex refractive index determined by this procedure would be consistent with the experimental data. For the complete set of particle samples, a range of values of  $n_{\text{Im}}$  was found. The geometric mean of this set was

$$n_{\text{Im}} = 0.005, \quad (15)$$

with an uncertainty factor (geometric standard deviation) of about 2. Within the range of observed  $n_{\text{Re}}$  values,  $n_{\text{Im}}$  was not very sensitive to the assumed  $n_{\text{Re}}$ .

The complex refractive index given by (14) and (15), taken with a particle size distribution and an incident solar wavelength,  $\lambda$ , provide all of the information required to compute the absorption and backscattering parameters  $a$  and  $b$  for an optically thin aerosol layer. The parameters are given by

$$a = \int_0^\infty \pi r^2 Q_a(m, r/\lambda) n(r) dr, \quad (16)$$

$$b = \int_0^\infty \pi r^2 Q_b(m, r/\lambda) n(r) dr, \quad (17)$$

where  $Q_a$  and  $Q_b$  are the Mie-scattering efficiencies for absorption (van de Hulst, 1957; Kerker, 1969; Deirmendjian, 1969) and hemispheric backscattering (Chýlek, 1973; Chýlek *et al.*, 1975), respectively. Using (16) and (17), we computed  $a$  and  $b$  for a wide variety of log-normal distributions, as given by (13). The results are summarized in Fig. 2, which shows isopleths of constant values of the ratio  $\rho = a/b$  as a function of the geometric mean radius  $r_g$  and the geometric standard deviation  $\sigma_g$ . The range of values from  $\rho = 0.25$  to  $\rho = 35$  was calculated, using the refractive index  $m = 1.525 - 0.005i$  and the wavelength  $\lambda = 0.5 \mu\text{m}$  (mid-visible sunlight). Different curves would result from other choices for  $m$  and  $\lambda$ . To indicate the effects that the soil-dust particles studied by Grams *et al.* (1974) have on the input of solar radiation, we also plotted the values of  $r_g$  and  $\sigma_g$  determined for each of the 14 collected samples.

The curves in Fig. 2 show that the observed aerosol size distributions result in aerosol parameter values that fall in the range of  $5 < \rho < 28$ . We note that these observed values are quite high compared to the numerical estimates of previous authors (e.g., Ensor *et al.*, 1971). Fig. 1 shows that all of the experimentally determined  $\rho$  values exceed the critical ratio  $\rho_c$  for all surface albedos larger than  $A = 0.09$ . Thus, according to the models described in Section 2, all of the sampled aerosols would lead to increased energy of the earth-atmosphere system.

We emphasize that the relatively large values of  $\rho$  shown in Fig. 2 are, in part, due to the relatively large mean sizes of the sampled particles, which were collected at a height of only 1.5 m in an agricultural setting before the growing season. Fig. 2 shows that  $\rho$  is quite sensitive

to particle size, and that polydispersions with smaller  $r_g$  and  $\sigma_g$  would have smaller values of  $\rho$ .

#### 4. Conclusions

The soil-dust complex refractive index and particle size distributions determined by Grams *et al.* (1974) produce aerosol layer absorption-to-backscatter ratios  $\rho$  which exceed the critical values  $\rho_c$ , obtained as the common result of several analytical models of aerosol radiative effects. These models therefore predict that any activities that would lead to an increase, on a global scale, in surface layer soil-dust particles, such as the increased soil erosion associated with agriculture, would result in increased solar energy input to the earth-atmosphere system.

We feel, however, that this conclusion must be accompanied by two reservations. We have already noted above that the large values of  $\rho$  were due, in part, to the large mean particle sizes reported by Grams *et al.* (1974). In that study, particle size distributions were determined by phase contrast microscopy. In the range of particle sizes for which optical microscopic measurements are applicable ( $r > 0.3 \mu\text{m}$ ), the method gave reliable results (see Table 1 of Grams *et al.*, 1974), and it had the added advantage of permitting direct visual examination of each particle. However, additional observational data for smaller particles could modify the values of  $\rho$  plotted in Fig. 2. Thus, it would be desirable to obtain additional size distribution measurements on soil-dust aerosols, using techniques that are accurate throughout the entire range of radii from 0.1 to  $20 \mu\text{m}$ . If differing size distributions were so determined, the techniques outlined in this paper and the results shown in Fig. 2 could be used to obtain appropriate values of  $\rho$ . However, it should also be mentioned that laboratory drying of hygroscopic particles can tend to reduce particle sizes and, thus, accurate *in-situ* size measurements in humid (e.g., tropical) regions could yield mean particle sizes even larger than those of Grams *et al.* (1974). In this connection, we feel that an important result of the present study is to emphasize the importance of particle size distribution in determining the radiative effects of aerosols. Several recent numerical modeling studies of aerosol radiative effects (Yamamoto and Tanaka, 1972; Braslau and Dave, 1973a, b), while considering effects of different aerosol absorptivities (or  $n_{Im}$  values), have used only a single particle size distribution, i.e., the Junge (1963) power law distribution function. The present results suggest the need to repeat some of these computations, to ascertain whether aerosol size effects are as important in the more realistic numerical models as they are in the simple analytical models. The numerical solutions of Herman *et al.* (1971) indicate that particle size effects are important in determining the amount of forward-scattered sunlight.

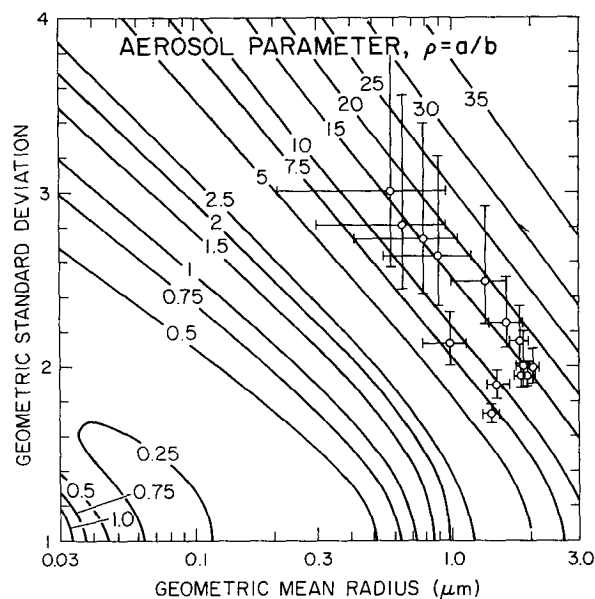


FIG. 2. Values of the aerosol parameter  $\rho$  (ratio of aerosol layer absorption to hemispherical backscattering) as a function of geometric mean radius  $r_g$  and geometric standard deviation  $\sigma_g$  for a log-normal size distribution,  $m = 1.525 - 0.005i$ , and  $\lambda = 500 \text{ nm}$ . Values of  $r_g$  and  $\sigma_g$  are plotted for 14 size distributions for the airborne soil particles collected by Grams *et al.* (1974).

The second reservation concerns the use of the value  $n_{Im} = 0.005$  obtained by Grams *et al.* (1974). As noted above, use of this value in computing the hemispheric backscattering fraction  $b$  is justified because  $n_{Im}$  was chosen as the value which gave computed Mie scattering intensities that best duplicated the angular variation of the actual scattering measurements. However, use of this same value of  $n_{Im}$  to compute the absorption fraction  $a$  may possibly be questioned. This follows from the fact that Holland and Gagne (1970) have shown that light intensities scattered in the backward direction by non-absorbing, irregularly shaped particles are considerably less than those scattered by equivalent non-absorbing spheres. This is qualitatively similar to the effect caused by increasing the  $n_{Im}$  values of the spheres [see Grams *et al.* (1974) for illustration]. Thus, if the angular dependences of the intensities measured by Grams *et al.* were due in part to particle shape effects, their technique would yield a value of  $n_{Im}$  which, while producing correct angular distributions, could overestimate absorptivities. Thus, the value of  $n_{Im} = 0.005$ , when used in (16) to compute the absorption  $a$ , should probably be looked upon as an upper limit.

We note, however, that in computing Mie intensities, increasing the value of  $n_{Im}$  also causes significant intensity reductions in the angular region between  $20^\circ$  and  $90^\circ$  (Figs. 1 and 8 of Grams *et al.*, 1974). The measurements of Holland and Gagne, on the other hand, show that asphericity does not cause similar intensity

reductions in this forward region. In fact, aspherical non-absorptive particles must scatter slightly *more* light (relative to spheres) into the forward hemisphere to compensate for the reduced backscattering. (This compensation is of course not required for absorbing spheres.) Since the optimum  $n_{Im}$  values determined by Grams *et al.* (1974) often improved (relative to the case  $n_{Im}=0$ ) the fits in the  $20^\circ$  to  $90^\circ$  region, as well as in the  $150^\circ$  to  $170^\circ$  region, it is highly likely that the measured angular scattering diagrams differed from those of non-absorbing spheres principally because of absorption effects, rather than shape effects. Thus, although the value  $n_{Im}=0.005$  is, strictly speaking, an upper limit when used for computing  $a$ , the above evidence indicates that the actual value is probably close to this limit.

In this connection we note that an approximate value of 0.005 for  $n_{Im}$  has also been reported by Ward *et al.* (1973), who made atmospheric light-scattering measurements on a university campus in Florida. However, those investigators point out that their determination of  $n_{Im}$ , which uses methods similar to those of Grams *et al.* (1974), may also be influenced by effects of nonsphericity. [In future light-scattering measurements the effects of nonsphericity and absorption may possibly be separated by measuring the rotation of the plane of polarization of the scattered light (Ward *et al.*, 1973).] It should be emphasized, however, that similar values for  $n_{Im}$  have recently been determined by other investigators using entirely different methods. For example, Lindberg and Laude (1974), using measurements of diffuse reflectance from bulk samples of desert dust, have determined the value  $n_{Im}\approx 0.007$  for visible radiation. Still another method was used by Ivlev and Popova (1973), who studied the fractional makeup of minerals in soils and employed Kramers-Kronig analyses to determine the optical properties of these minerals. They summarized their studies by specifying the refractive index of a synthetic aerosol using a volume-weighted average of the optical constants of each component; the value of  $n_{Im}$  thus obtained was constant and equal to 0.005 for visible radiation.

Again, we note that, although the Grams *et al.* (1974) value of  $n_{Im}=0.005$  should be taken as an upper limit in computing  $a$ , it should be taken as representative in computing  $b$  because this value of  $n_{Im}$  produces the best agreement between computed and measured angular scattering diagrams. The large values of  $\rho$  shown in Fig. 2 are due as much to small values of  $b$  as to large values of  $a$ . The maximum effect, on computed  $\rho$  values, of reducing  $n_{Im}$  in computing  $a$  (but not  $b$ ) may be estimated from the results of Chýlek *et al.* (1974). Those results show that, for almost all particle sizes, when  $n_{Im}$  is varied in computing *both*  $a$  and  $b$ , the change in  $\rho$  is proportional to the change in  $n_{Im}$ . Thus, for example, a 10% reduction in the value of  $n_{Im}$  produces approximately a 10% reduction in  $\rho=a/b$ . This reduction in  $\rho$  results in part from a reduction in  $a$

and in part from an increase in  $b$ , and thus gives the maximum possible change in  $\rho$  which would result from reducing  $n_{Im}$  in computing  $a$ , but holding it constant in computing  $b$ .

We finally note that although this study applies only to surface-layer soil-dust aerosols, it is possible that on a global basis such particles may constitute a substantial fraction of the total aerosol optical depth. Studies of the chemical composition and height distribution of rural (not necessarily "background") aerosols will be required to ascertain the validity of this possibility. In this connection we note that Bryson (1974) has recently presented evidence that increased concentrations of *both* volcanic particulates and airborne soil dust have apparently been correlated with significant surface temperature changes in the past. While this indicates that such particles can be of climatic significance, the *sign* of the observed temperature change appears to contradict the present results, because increased aerosol concentrations have been associated with *reduced* surface temperatures.

There are, however, two possible resolutions of this apparent contradiction. First of all, the effects of volcanic aerosols must be considered separately from the results of the present study, because the composition and size distribution of volcanic aerosols almost certainly differ from those of the soil dust aerosols that we have considered. The dominant constituent of the volcanic particles which form or persist in the atmosphere is sulfate, usually in the form of aqueous sulfuric acid or ammonium sulfate (Cronin, 1971; Friend *et al.*, 1973; Castleman *et al.*, 1973). A recent study by Neumann (1973) has shown that neither of these compounds is absorptive in the visible. Thus particles having this composition would have  $a=\rho=0$ , giving rise to a cooling effect of volcanic aerosols which is consistent with the results of the simple models.

On the other hand, for soil dust aerosols, it must be remembered that while the results ( $\rho>\rho_c$ ) of the present study imply increased energy for the earth-atmosphere system as a whole, they imply increased surface temperatures only if the soil dust aerosol layer has sufficient convective contact with the surface. Any aerosol layer (regardless of the amount of absorption  $a$ ) causes a reduction of solar radiation at the surface, so that unless layer heating is somehow transferred to the surface, cooling of the surface must result, independent of the value of  $\rho$ . Thus the soil dust layers cited by Bryson (1974) may simply have had very small contact with the surface—much less, in fact, than the amounts considered typical by Mitchell (1971) and listed in Table 1. This also is certainly true of stratospheric particulates, which represent the most persistent aerosol component resulting from major volcanic eruptions.

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