Results of a Comprehensive Atmospheric Aerosol-Radiation Experiment in the Southwestern United States*

Part I: Size Distribution, Extinction Optical Depth and Vertical Profiles of Aerosols Suspended in the Atmosphere

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ABSTRACT

An exploratory field experiment was undertaken to determine the practicality of a method specifically designed to obtain the optical properties of aerosols as they relate to the earth's radiation balance. The method requires a basic set of data consisting of the vertical distribution of aerosol concentrations, size distribution, optical depth, and net radiation fluxes. From these data radiation absorptions are determined, and effective aerosol refractive indices consistent with the actual absorption are deduced through the application of precision radiative transfer calculations. The results of 11 experiment episodes involving a combined aircraft and surface-based measurement system are described. The episodes took place in an arid desert region located near Blythe, California, and in a semiarid agricultural region located near Big Spring, Texas. Part I deals with the physical-numerical depiction of such aerosol properties as optical depth, size distribution, and vertical profiles of concentration. Part II will deal with the analysis of measurements of the radiation field leading to the deduction of the effective aerosol refractive index compatible with the absorption of solar radiation.

1. Introduction and background

The energy that is available to drive the circulation of the atmosphere is produced by the balance between solar shortwave radiation which impinges on the planet earth and the terrestrial infrared radiation which is emitted to space. A quantitative summary of the energy budget of the earth was compiled by Lettau (1954); a more recent study by Robinson (1970) indicates the total kinetic energy of the atmosphere to be about 6% of the solar power absorbed by the planet. Unfortunately, radiometric measurements are seldom more accurate than 1%. Long-range trends which are climatically significant may be caused by fractions of 1% change in the radiation budget, and therefore are difficult to establish directly.

The climatic impact of a variety of anthropogenic, volcanic or solar effects may be assessed by modeling the radiation budget. Crude calculations of the radia-

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* First GAARS field test.
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tion budget indicate that changes in the optical properties and distributions of aerosols may affect the net radiation balance in a manner which may cause a significant variation in the earth's mean climatic state. This is equivalent to changing the heat input to the dynamical atmosphere-earth-ocean system (Budyko, 1969; Charland and Pilat, 1969; Ensor et al., 1971; Yamamoto and Tanaka, 1972). Although these radiation balance calculations do not consider global variations in the distribution and composition of radiatively active constituents, they nevertheless point out significant sensitivity to changes in aerosol optical properties and concentrations. Atmospheric aerosol absorption can easily amount to a few percent of the solar flux (Robinson, 1966; Kondratyev et al., 1974) and this compares significantly to the energy in the atmospheric circulation.

Considerable progress has been made in theoretical research in the calculation of radiation transfer, but a great deal more work remains to be done in applying this research to the real atmosphere. The radiation balance affected by aerosols cannot be adequately modeled because sufficient data are lacking on the physical, chemical and optical properties of aerosols, including the spatial distribution of concentration.

Recognizing the need for a thorough knowledge of atmospheric aerosols and their radiative effects on a global scale, in the Spring of 1971 the authors outlined the requirements for a comprehensive research program directed to meeting this need. An outcome of their subsequent interactions was the recommendation for a Global Atmospheric Aerosol and Radiation Study (GAARS)6 to be structured as an open cooperative endeavor drawing upon the scientists and facilities of universities, NCAR, and such organizations as NOAA and NASA. The cooperation was suggested because of the magnitude and diversity of the expertise required, which no single institution could supply.

The underlying concept of the GAARS is the acquisition and application of a self-consistent set of measurements necessary for sound scientific interpretation. Based on the scheme illustrated in Fig. 1, a fundamental set of data to be obtained from an aircraft and surface-based measurement system was prescribed for an exploratory experiment. In May 1972 a first field test was undertaken to test the practicality of the GAARS concept. The results of the field test are reported herein. For the sake of brevity, whenever practical we reference available previous work to describe the specific details of the varied components that contribute to this paper.

For readers who are not familiar with the ambitious aerosol and radiation research program of the Soviet Union, Complete Atmospheric Energetics Experiment (CAENEX), we call attention to a reference on the formulation of the program (Kondratyev, 1973). Since its inception a number of experiments have been conducted. Although CAENEX and GAARS were formu-


![Fig. 1. Dependent sequence of inputs proceeding from aerosol chemical and physical properties to the determination of their radiative effects.](image-url)
lated independently, there exists, nevertheless, a remarkable degree of similarity between the two.

2. Objectives of the first field test

The first GAARS field test was aimed at the acquisition of a sufficient set of measurements of aerosol properties required as input for radiative transfer calculations. These measurements, comprising a subset of the requirements of Fig. 1, supply only preliminary data for examining the crucial aspects of the GAARS concept.

Fig. 2 shows schematically the relationship between radiation measurements and the variables needed for radiative transfer calculations at any wavelength. As a minimum, measurements of the vertical upward and downward hemispheric fluxes are required at two levels in the atmosphere. The reflectivity of the surface is determined from the ratio of flux measurements at the lowest level.

It is useful, although not necessary, to have a measurement of the directly transmitted solar intensity \( I_i \) (see Fig. 2) at a number of levels \( i \) in order to obtain the aerosol extinction optical depth \( \Delta \tau_i \) between any two levels for use in the radiative transfer calculations. The relationship between \( I \) and \( \tau \) is \( \Delta \tau_i = -I_i \Delta \tau_i \), where \( \Delta \tau_i \) is the decrease in direct intensity and \( \Delta \tau_i \) the optical depth increment for all constituents at the \( i \)th level.

The aerosol scattering phase function \( P_i(\theta) \), which determines the angular distribution of radiant energy scattered by an ensemble of particles, can be either measured directly, or, as in our case, inferred by using size distribution measurements and Mie scattering theory for spherical particles.

Thus, the measurement requirements are 1) vertical hemispheric fluxes, 2) surface reflectivity, 3) optical depth (or vertical profiles of aerosols), 4) size distributions of aerosols, and 5) aerosol refractive indices. Some spectral resolution is needed for the radiation measurements in order to avoid the strong molecular absorption bands of water vapor and oxygen at wavelengths \( \gtrsim 0.7 \mu \text{m} \).

Nearly all of the instrumentation for the necessary measurements was already available from the ongoing research being conducted by the participants. This points to the advantage of cooperation between established organizations since the cost of the experimental results is dramatically reduced.

The research topics that are presented can be classified as 1) aerosol extinction and size distributions, 2) vertical profiles of aerosols, and 3) radiation fluxes and radiative transfer analysis. To provide a convenient presentation format which relates to topics previous works, our report is given in two parts. Part I covers categories 1 and 2, and Part II will cover category 3 and conclusions. Also, a special effort was made throughout to elucidate the relationships between the various measurements, and how they are used to achieve the basic aims of the first experiment as well as the longer range goals of GAARS.

3. Site survey and experiment period

Calculation of radiative transfer in this experiment is limited to a plane-parallel, vertically inhomogeneous atmosphere with molecular and aerosol absorption and with scattering by molecules and aerosols. The powerful light-scattering and inhomogeneous nature of clouds were purposely avoided in the first field test because their effects could not be accounted for accurately with the measurement and analysis techniques that were specified in the present experimental procedure. Another consideration was the variation in aerosol optical properties due to the influence of high humidities. In view of these limitations the southwestern region of the United States was selected for experiment sites as the experimental design fitted the features of a desert atmosphere. The month of May is a time of high turbidity in the desert (Flowers et al., 1969) and low probability of clouds in the southwestern desert region of the United States.

In late January 1972 a survey flight was undertaken to locate sites in the southwest for the first field tests

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\[ \Delta \tau = \text{Incremental extinction optical depth} \]
\[ E_3 = \text{Surface emissivity} \]
\[ P(\theta) = \text{Aerosol scattering phase function} \]
\[ R = \text{Reflectivity} \]
\[ q = \text{Absorption} \]
\[ e = \text{Extinction} \]
\[ x = \text{Scattering} \]
\[ F = \text{Hemispheric fluxes} \]
\[ I = \text{Intensity of the solar beam} \]

**Fig. 2.** Diagram showing measurement of radiation field variables (left-hand side) and of the variables required for the calculation of the radiation field variables (right-hand side).

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\( \sum_{i=1}^{4} \Delta \tau_i = \Delta \tau^a + \Delta \tau^e \)

\( \Delta \tau^e \) is the incremental extinction optical depth due to aerosols. The symbols are defined as follows:

- **E** = Surface emissivity
- **P(θ)** = Aerosol scattering phase function
- **R** = Reflectivity
- **q** = Absorption
- **e** = Extinction
- **x** = Scattering
- **F** = Hemispheric fluxes
- **I** = Intensity of the solar beam

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\footnote{The aerosol refractive index was not measured directly, but was inferred from measurements 1) through 4). Details of the procedure are given in Part II of this work.}
scheduled for the coming May. The survey chiefly covered the states of Arizona and New Mexico, but included the southeastern tip of California. Two experiment sites were selected: Blythe, Calif., and the nearby vicinity; and Big Spring, Tex., where aerosol research was already in progress (Gillette et al., 1974). The details and outcome of the survey were reported by Furukawa and DeLuisi (1972).\(^8\)

A total of 11 measurement episodes took place in an interval of less than two weeks, from 14–25 May. An NCAR Queen Air provided the aircraft support. Seven flights were made at the Blythe area and four at Big Spring, with supporting measurements made from the surface at both sites. Of these flights, seven were daytime and four were nighttime. Flight 4, a daytime flight, was terminated because of cirrus development when the experiment was about half completed.

Approximately 90\% of the data from these flights are of a usable quality. The research described in the present paper includes the results of practically all of the individual preliminary GAARS experiments.

4. Aerosol optical extinction observations

The monochromatic extinction of electromagnetic radiation in the atmosphere can usually be represented by the Beer-Lambert law, stated as

\[
\frac{dI(\lambda)}{I(\lambda)} = \sum_{i=1}^{n} a_i(\lambda) \rho_i(x) dx,
\]

where \(I(\lambda)\) is the intensity of the radiation at wavelength \(\lambda\), \(a_i(\lambda)\) is the extinction (absorption, scattering, or both) cross section for the \(i\)th constituent of the medium \(x\), and \(\rho_i(x)\) is the concentration. The integral of the right-hand side of Eq. (1) over a finite path length is commonly called the optical depth in connection with usage in atmospheric problems. In our case \(n \leq 3\) which includes molecular scattering, water vapor absorption, and aerosol scattering and absorption. By itself, Eq. (1) cannot be used to separate the extinction effects of absorption and scattering. Since the method we use to obtain a separation is sensitive to variations in optical depth, it is mandatory that errors in optical depth measurement be minimized.

Aerosol extinction data were obtained in four ways. A multi-wavelength ground-based photometer obtained extinction optical depths in narrow wavelength bands centered at the wavelengths given in Table 1; a Volz (1959) photometer measured extinction optical depths at 0.5 \(\mu\)m; a pyranometer system on the aircraft was used to infer a change in extinction optical depth between the highest (3–5 km) and lowest levels (near the surface) flown; and size distributions measured in situ by the aerosol sizing impactor were used to calculate aerosol extinction, using Mie scattering theory. The two latter measurements provided information on the optical depth of the slab of atmosphere being studied while the first two provided information on the optical depth of the entire atmosphere as well as a cross check on the representativeness of the aerosol sampling results.

In the following subsections, each method that was used to obtain aerosol extinction information is described separately. After the last subsection, a summary discussion is given on aerosol optical depth. The subsection on impactor data includes the results of a special experiment that provided size distribution information from inverted solar aureole measurements.

a. Multi-wavelength photometer

Fig. 3 shows a schematic diagram of the filter photometer that was used to measure the optical depth of aerosols. It consists of a light-tight box containing two wheels which hold up to 12 interference filters, a motor for rotating the filters, and a silicon photodiode detector. The collimator consists of a tube 30 cm long containing a quartz lens, field stop and baffles. In order to reduce the contribution of forward scattering to the direct component of the intensity, the collimation half-angle was chosen to be 0.35\(^\circ\).

The photosensor was a UDT model PIN-10 silicon detector, with a usable range of 0.3 to 1.1 \(\mu\)m and an active area equal to 1.25 cm\(^2\). An opaque position in the filter wheel provided the dark current which is subtracted from the signal currents. The dark current is typically less than 0.1\% of the signal current.

Seven interference filters were used over the range of 0.42–1.01 \(\mu\)m. Their center wavelengths and bandpass characteristics are given in Table 1. The transmission characteristics of each filter were measured before and after the experiment with a Cary model 14 spectrophotometer.

In order to check the consistency of the data, it is desirable to have a relative overall instrument calibration that includes the filters, detector and optical system. The calibration with the instrument mounted on a precision optical bench and using three tungsten-iodine calibration lamps was performed once before and after each run. The overall variability under these circumstances was found to be approximately 1\%.\(^8\)

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\(^8\) Report available from the authors.
The electronic system consisted of a logarithmic amplifier driving a Hewlett-Packard voltmeter. Its output was used as a measure of the solar irradiance at the wavelength of measurement.

Additional control circuitry was employed to actuate driving motors for the filter wheel and to select the appropriate filters. In order to provide automatic tracking of the sun, the radiometer was installed on an equatorial telescope mount driven by a dc motor. Whenever necessary, manual corrections are made to maximize the readings, insuring precise orientation for direct viewing of the sun.

Values of log solar irradiance at each wavelength versus $\sec \theta$ were least-squares computer-fitted to a straight line, with the slope determining the atmospheric extinction coefficient. Corrections for Rayleigh scattering and ozone absorption were made to the measured extinction in order to extract the extinction due to aerosols.

For comparison Fig. 4 shows the best estimates of aerosol optical depth determined in four different ways. The abscissa gives the flight numbers of six daytime flights and the ordinate the aerosol optical depth either for the entire depth of the atmosphere as in the case of the Volz (dotted line) and multi-wavelength photometers (dashed line), or for the layer of atmosphere through which aircraft measurements were made during a particular measurement episode (dot-dashed line). The optical depths are for a wavelength of $\sim 0.5 \mu m$. The wavelength dependence of optical depth as measured by the multi-wavelength photometer is discussed in the section on impactor data.

b. Volz photometer

The Volz photometer (or versions of it) has been widely used for measuring the turbidity of the atmo-
sphere (Fischer, 1967; Volz, 1969; Flowers et al., 1969; Bach, 1971; and others). Although measurement uncertainties encountered with the Volz photometer have been reported (Laulainen and Taylor, 1974), careful usage of the instrument reduces error problems. Our Volz photometer measurements were calibrated to be consistent with the data supplied by the multi-wavelength photometer described in the previous section.

Measurements with the Volz photometer were made at unspecified hours during the days when the experiment was in progress. Also, the locations where these measurements were made did not necessarily coincide with the experiment sites. Our interpretation of the results of the Volz photometer data is therefore limited to defining the mean aerosol optical depth of a region extending from the Blythe airport to the Quartzsite area, approximately 32 km, and along an approximate 15 km line extending from Big Spring northward.

Reexamination of Fig. 4 shows that, in general, the day-to-day variations in optical depth given by the Volz photometer (λ ≈ 0.50 μm) follow the trends shown by the other data in the same figure. The smaller day-to-day variations in the Volz data can be explained by the smoothing effect of averaging measurement results over greater time and space dimensions, compared to the results of the other measurements which were made within a few hours time and along an approximate 8 km path.

c. Impactor data

The impactor used to obtain size distribution data was described by Bifford and Ringer (1969). The range of measured aerosol particle radii extends from 0.4 to 12 μm. The impactor installed on the aircraft was used to sample aerosol size distributions at each predetermined flight level; one was also operated on the ground at Big Spring. Approximately three to five levels in the atmosphere between the surface and 5 km were sampled in the various episodes. Local flight restrictions in the Big Spring area prevented aircraft ascents above 3 km except during one nighttime episode.

Since our main interest in aerosol size distribution was for the purpose of determining their optical properties, we used Mie’s solutions for scattering of electromagnetic radiation by spherical particles (see Deirmendjian, 1969; van de Hulst, 1957). For each size distribution measured, a volume extinction cross section was calculated using a complex refractive index of 1.5–0.01 for wavelengths of 0.4, 0.5, 0.6, 0.7, 0.8, 0.9 and 1.0 μm (see, e.g., Herman et al., 1971; DeLuisi et al., 1972; Harris, 1973). Then, to estimate a total extinction optical depth for the layer of atmosphere sampled by the aircraft a midpoint integration was performed:

\[ \tau_d(\lambda) = \sum_{i=1}^{n} \delta_i(\lambda) \Delta z_i, \]

where \( \tau_d(\lambda) \) is the spectral aerosol extinction optical depth at wavelength \( \lambda \), \( \delta_i(\lambda) \) the aerosol volume extinction cross section, and \( \Delta z \) the finite geometric thickness of atmosphere containing aerosols having the size distribution measured by the aircraft impactor. The top and bottom layers in most cases are only half the normal thickness. This procedure follows the scheme depicted on the right-hand side of Fig. 2.

The calculated optical depths at 0.5 μm are shown in Fig. 4 for comparison with the other types of measurements. From the figure it is apparent that these calculated optical depths vary widely. A cause of this variability is suggested in the section on vertical profiles of aerosols where it is seen that there are cases where layers of high concentrations of aerosols can be missed (giving low calculated optical depths) by the sampling aircraft (for example, flights 1 and 11). Combined with this problem is the uncertainty in the measurement of size distribution due to the sampling of a small volume of air which is taken to be representative of a much larger volume, namely, the entire layer. However, the uncertainty in the Mie extinction cross sections calculated from the size distribution data is close to one-half the uncertainty in the size distribution data because the integration over the size distribution function reduces the effects of random error in each size interval, according to DeLuisi et al. (1974).

The effects of spatial variations in the aerosol concentrations and size distributions on radiation propagating through the atmosphere tend to be averaged out by the integrating nature of the scattering process. A size distribution of aerosols sampled by the aircraft at each level can be considered as a horizontal average since the flight path is on the order of 8 km. The size distributions were vertically averaged for each flight to obtain a single, optically effective aerosol size distribution. For each finite radial size interval \( j \) of size distribution \( i \) a vertically averaged, optically weighted \( dN/d \log r \) function is found by summing over \( n \) sampled layers:

\[ \frac{\Delta N_j}{\Delta \log r_j} = \frac{1}{\sum_{i=1}^{n} \Delta \tau_d \Delta N_{i,j}} \sum_{i=1}^{n} \Delta \log r_i, \]

where \( N \) is particle concentration and \( r \) is particle radius, and where \( \Delta \tau \) was calculated from Eq. (2) for a wavelength of 0.5 μm. Fig. 5 shows the average distributions for each of the three daylight measurement episodes (1, 5 and 7) at the Blythe site, and Fig. 6 shows the averaged size distributions for each of the three daylight measurement episodes (8, 9 and 11) at the Big Spring site. These size distributions seem to display two basic features. Distributions for flights 1, 8 and 11 have far fewer large particles >1–2 μm than the distributions for flights 5, 7 and 9.

Since the variations in size distributions of aerosols are often related to variations in optical effects such as spectral extinction, we refer to the optical data pro-
Fig. 5. Moderate radial resolution aerosol size distributions for slabs of atmosphere approximately 4 km thick. \( N \) is number density (cm\(^{-3} \)) and \( r \) the particle radius (\( \mu \)m). Data are for the Blythe, Calif., area.

Fig. 6. As in Fig. 5 except for the Big Spring, Tex., area.

Fig. 7. Average aerosol extinction optical depth versus wavelength for six measurement episodes: flights 1, 5, 7, 8, 9 and 11. Solid line is estimated from impactor size distribution data using Mie scattering theory; dashed line is aerosol optical depth plus molecular absorption; circles with dots are aerosol optical depth containing no molecular absorption.

Inferred from theory. The slope of the extinction optical depth versus wavelength or spectral extinction cross section is strongly dependent on the slope of the size distribution of aerosols, primarily in the highly active particle radial size region 0.1–10 \( \mu \)m. With this relationship in mind, we further examined the results of Fig. 7 in an effort to resolve the disagreement.

If the data for Fig. 7 are averaged according to the two size distribution categories, three for the size distributions having an excess of large particles and three for size distributions having fewer large particles, then the spectral extinction optical depths separate into two curves; Fig. 8 shows this separation. The average for flights 5, 7 and 9 on this figure show a spectral extinction that is neutral with wavelength implying that a Junge (1958) type size distribution having an equivalent optical effect would have a power of 2 or

\[
\frac{dN}{d \log r} = a r^{-\gamma}, \quad \gamma = 2,
\]

where \( a \) is a constant depending on total concentration. This result stems from the relationship \( a = \gamma - 2 \), where \( \gamma \) is the slope of the spectral extinction cross section when it is expressed as

\[
\tau(\lambda) \propto \lambda^{-\alpha}.
\]

In contrast to the neutral extinction of flights 5, 7 and 9 the data for flights 1, 8, 11 show a distinct variation with wavelength, where now the power of the Junge size distribution would be approximately 2.6 (i.e., now
\( \alpha = 0.6 \). Although these relationships between aerosol size distribution and spectral extinction optical depths are treated quite simply here, they are nevertheless meaningful and are sufficient to use as cross checks for comparing aerosol data with optical depth (Quenzel, 1970; Deluissi et al., 1972; Shaw et al., 1973). It is to be noted that each of the three spectral extinction measurements that was used for an average curve in Fig. 8 displayed features in wavelength variation similar to the average. For example, the average of flights 5, 7 and 9 giving neutral extinction did not consist of single curves that varied significantly from the mean.

In regard to the individual aerosol size distributions used to obtain the spectral extinction curve in Fig. 7, it is of interest to note that there was not a single case of calculated Mie extinction that showed a decrease with increasing wavelength. This result has led us to surmise that either the impactor is giving systematically slightly lower concentrations of small sized particles with respect to the larger sizes or that the effects of the smaller particles not measured by the impactor are significant, or perhaps both. The aerosol particle size region less than a few tenths of a micron can be optically important if concentrations are high (Deluissi et al., 1972).

In spite of the disparity in the wavelength variation of optical extinction between aerosol and spectral optical data, the absolute magnitudes were quite satisfactory. At 0.5 \( \mu m \) the photometer extinction data in Fig. 7 are about 20\% higher than the calculated extinctions. The lidar data, which are described in the next section, indicated that the concentrations in the aerosol layers being sampled were rapidly decreasing at a kilometer or so above the highest levels flown by the aircraft, the decrease generally occurring at about 4–6 km. Indeed, from this information, a 20\% difference is about what could be expected as a result of the remaining aerosols extending to the top of the atmosphere. Because the aerosol impactor data agree so well on the average with the photometer data, we have accepted the average aerosol optical depth inferred from the impactor data as representative of the average of all slabs of atmosphere sampled by the aircraft. However, we chose to employ the aircraft radiometer system to provide a measure of aerosol optical depths for the individual measurement episodes (see the summary discussion of aerosol optical depth below).

We now refer to a scanning photometer experiment that was designed to give information on the size distribution of aerosols throughout the entire depth of atmosphere. Since most of the aerosols are concentrated in the lowest levels of the atmosphere, the size distribution data given by the scanning photometer can be compared with the size distribution data obtained from the multi-wavelength photometer, and from the impactor.

A scanning photometer similar to the instrument shown in Fig. 9 was used to measure the intensity of the sun’s aureole. Fig. 10 shows a measured angular scan through the sun and in the solar plane. The measurement was made during flight 4 which was termi-
nated because of increasing development of cirrus overhead. This particular measurement is shown because it illustrates the effect of cirrus (~12° below the center of the sun), and hence the sensitivity of the aureole intensity to clouds too thin to be visible to the eye.

Typically, a measurement free of cirrus effects is inverted in the manner described by Twitty (1975), and by Weinman et al. (1975) to derive aerosol size distributions. The inverted size distribution shown in Fig. 11 (nearly straight, solid line) is compared with the concurrently observed impactor size distribution for flight 5 (curved solid line). The inverted distribution has been arbitrarily normalized to the impactor distribution to facilitate comparison. It is noted that the inverted distribution has a slope of $-2$ in the Junge (1958) form. We have seen in the section on impactor data that the size distribution of flight 5, by virtue of its calculated and observed equivalent optical extinction, is effectively similar to a slope of $-2$ in the Junge form.

Also, on board the aircraft was a Bausch and Lomb optical particle counter that was used to measure aerosol size distribution at the same time as the impactor. The radial resolution and range of these data are insufficient for Mie scattering calculations, but we have plotted data for flight 5 on Fig. 11 (dashed line) to compare with the other size distributions. Overall, the agreement is good.

Fig. 10. Intensity distribution of the solar aureole for a vertical angular scan through the sun at a 43° zenith angle. The anomaly at about $\Delta \theta = 12°$ is due to the presence of cirrus invisible to the eye.

Fig. 11. Comparison of aerosol size distributions as measured by impactor (flight 5) and by a Bausch and Lomb optical counter (dashed curve), and as deduced from aureole data (solid, nearly straight line). These data were obtained concurrently by the scanning photometer at the ground and with the aerosol counters on aircraft, sounding from near the surface up to about 5 km.
since the optical equivalences of the size distributions are not in serious disagreement.

An improved and more extensive airborne scanning photometer experiment was carried out a year later (Twitty et al., 1975) and a similar comparison was made with impactor data. Concentrations of small particles measured by the impactor were slightly lower than the concentrations deduced from the scanning photometer measurement; however, it is to be noted that the intensity of the solar aureole is less sensitive to smaller particles.

d. Optical depth by aircraft radiometer

Our aircraft radiometer system was limited to measurement of the vertical hemispheric fluxes having wideband wavelength resolution. Detailed characteristics of the system will be given in Part II. Our interest here is with the use of the system to provide a measure of the aerosol optical depth for the slabs of atmosphere being investigated. The bandwidth of greatest importance is obtained by the 0.32–0.685 μm filters, thus providing the best estimate of optical depth consistent with the radiative absorption measurements for the same slab of atmosphere and spectral bandwidth.

As a first-order estimate, the radiation incident at the top of the slab is nearly all direct (i.e., \( I_d \) in Fig. 2), especially if few aerosols remain overhead. Using the Beer-Lambert law, the directly transmitted radiation at the bottom of the slab is given by

\[
I_{\delta\lambda}(\mu_0) = \mu_0 I_{o,\delta\lambda} \exp[-\tau_{\delta\lambda}/\mu_0],
\]

where \( \tau_{\delta\lambda} \) stands for the optical depth for a spectral bandwidth \( \Delta \lambda \) and \( \mu_0 \) is the cosine of the solar zenith angle. \( I_{o,\delta\lambda} \), normally the intensity of the extraterrestrial flux at the top of the atmosphere, is used here to denote the intensity at the highest level flown. Since the diffuse component of the total radiation is very small compared to the directly transmitted radiation, we assume for this calculation that the total radiation at all levels is entirely direct.

The solution for \( \tau_{\delta\lambda} \) in Eq. (5) is obtained by using the radiation measurements at the top and bottom of the atmospheric slab. In the absence of molecular absorption \( \tau_{\delta\lambda} \) represents extinction for molecular scattering and aerosol. Since the hemispheric viewing radiometers also receive the diffuse downward flux, \( \tau_{\delta\lambda} \) tends to be underestimated. By calibrating the atmospheric extinction optical depth results obtained by the aircraft radiometer against the impactor aerosol Mie optical depth results for a wavelength at 0.50 μm, the radiometer optical depths are made to be consistent with our best estimate of the average aerosol optical depth of all slabs of atmosphere sampled. Further refinement of the radiometer optical depth results, such as making allowance for the Rayleigh scattering optical depth, is not feasible because uncertainties due to measurement noise become too great.

Fig. 4 shows the optical depths estimated from the aircraft radiometer system (dot-dashed line). These results follow the general trend of the other measurements, but with less variation from flight to flight. The aerosol optical depths used in the analysis of radiation observations in Part II of this paper are those determined from the aircraft radiometer measurements. Uncertainties in these optical depths are on the order of \( \sim 10–20\% \).

e. Summary discussion on aerosol optical depth

Because our method of analysis of radiation data relies on a measurement of the optical depth of aerosols we have made a special effort to obtain representative estimates of this quantity. Normally, with measurement of the directly transmitted solar radiation (involving a collimating device) made at the top and bottom of a slab and the use of Beer's law a straightforward determination of the aerosol extinction optical depth can be made. In our case, the measurements were not available so we were forced to examine some alternate methods. From the individual investigation of each method for obtaining the optical depth of aerosol we have drawn the following conclusions:

1) Aerosol optical depths based on Mie calculations and measurements of particle size and number suffer from limitations in sampling a large volume of atmosphere, and in recording only a finite number of radius intervals.

2) Aerosol optical depths deduced from ground-based transmission measurements generally cannot be used to represent optical depths of a slab of atmosphere without ancillary information (such as can be provided by lidar) for the unsampled upper part.

3) Within limitations, it is possible to verify by means of atmospheric optical measurements (such as spectral extinction and solar aureole) the gross features of aerosol size distributions obtained concurrently by direct sampling methods.

5. Vertical optical structure of aerosols

The vertical optical structure of aerosols in the slabs of atmosphere have been determined by three methods: total-scatter integrating nephelometer measurements by aircraft platform, ground-based lidar, and Mie scattering calculations using the aircraft impactor data. The nephelometer and impactor measurements were made concurrently, while lidar operation (because of its limitation to nighttime operation) usually did not coincide in time with the other two. Each of the three measurement techniques provide unique information. Comparisons between the nephelometer and impactor results turned out to be especially useful.
a. Integrating nephelometer

The integrating nephelometer measures the volume scattering cross section of aerosols and air, giving a measurement in terms of a ratio of the excess scattering with respect to clean air (Ahlquist and Charlson, 1967). Measurements with the nephelometer provided data on the fine details of the vertical profile of aerosols as well as a qualitative check on the aerosol optical depths as determined by the radiometer and impactor size distribution data.

Analog output from the nephelometer was recorded on the aircraft’s data acquisition system at a rate of about 10 counts per second. The wavelength of maximum response of the nephelometer is near 0.525 μm.

The last level of the atmosphere sampled by the aircraft instrumentation was usually the highest. Immediately after sampling the last level, the aircraft made a sounding by slowly spiralling downward to within a few hundred feet or less above the surface. The nephelometer results for these soundings are shown in Fig. 12. Since a measurement episode took place in the order of a few hours, the final vertical sounding portrays an “instant” profile which could differ from the measurements obtained during a 2 h ascent. Concentrations of aerosol layers appeared to vary quite frequently in space and time. This variability has been observed with lidar by Fernald et al. (1972), and is evident from our own lidar data.

Using the aerosol size distribution measured by the impactor at the sampling levels flown by the aircraft, the equivalent nephelometer response was calculated using Mie’s solution for spherical particles having a refractive index of \( m = 1.5 - 0.01i, \) and for wavelengths ranging from 0.4–1.1 μm. The same computer program that was used to calculate aerosol optical depth with the impactor data was made to output results in terms of the nephelometer response, expressed in the form

\[
\beta_C(\lambda, z) = \frac{\beta_M(\lambda, z) + \delta(\lambda, m, dN(z)/d\log r)}{\beta_M(\lambda, z)},
\]

\( \delta \) This represents a guess on the basis of the works of other investigators, and is not a solution.
where $\beta_M$ is the Rayleigh volume scattering cross section at altitude $z$.

The results of the Mie calculations are shown in Fig. 12 as light straight lines connecting $\beta_C$ at the sampled levels. These can be directly compared with the integrating nephelometer results. Note that for flight 7 the abscissa is double those in the other figures.

Close inspection of the profiles in Fig. 12 reveals the high variability of concentrations of aerosols with altitude, obtained from the nephelometer results. It also indicates that impactor samples at discrete levels in the atmosphere can sometimes miss important layers, such as shown by flight 1. Thus, sampling only several levels in the atmosphere can lead to serious error when an attempt is made to assess total aerosol loading. An integrating nephelometer operating continuously at the surface also revealed a large temporal variation with diurnal cycles in aerosol concentration (Porch, 1974).

Further inspection of the individual profiles in Fig. 12 show that although the relative variations of each type of measurement correlate quite well, the absolute magnitudes do not. Furthermore, the absolute magnitudes vary from flight to flight. Returning to the optical depth data in Fig. 4 (flights 1, 5, 7, 9 and 11), it is noted that there are cases where the impactor data give calculated optical depths that are too high and that the corresponding calculated nephelometer profiles in Fig. 12 are also high; the converse also occurs. The discrepancies might have been caused by a calibration problem with the laboratory analysis of the impactor slides and possibly by the neglect of particles of radius <0.3 $\mu$m. Temporal and spatial variations in aerosol concentrations and uncertainties in the nephelometer and optical depth data undoubtedly make up a part of the differences observed in these comparisons.

Fig. 13 shows averages of impactor and nephelometer results which can be compared level by level. The number of samples for each average is given on the right-hand side of the figure. On the average, the results agree quite well. No further analysis has been made in regard to the discrepancies at the highest and lowest levels.

Vertical soundings of temperature were obtained by the aircraft during the times of the other observations (nephelometer, radiation, etc.). Variations in vertical temperature profiles existed at the boundaries and sometimes within the prominent aerosol layers.

b. Lidar soundings

The vertical variation of aerosol concentrations was also measured by lidar, which makes use of backscattering by a laser pulse directed vertically upward. The laser system used in the present experiment is a prototype of the system used by Fernald et al. (1975). This system was compact and completely mobile, being housed in a small 6 m air-conditioned travel trailer. Power was supplied by a diesel generator situated in the bed of the pick-up truck that was also used to haul the trailer to the experiment site.

The lidar system consisted of a $\frac{1}{2}$ J tunable dye laser operating at a wavelength of 0.585 $\mu$m. A 38 cm Fresnel lens was used in the light-collection telescope. Pulse length of the laser was 300 ns and spatial resolution was approximately 150 m. The data acquisition system has been described by Frush (1975).

Fig. 14 shows five aerosol profiles derived from averages of about 100 shots per profile. The ordinate is the ratio of Mie and Rayleigh backscatter divided by the Rayleigh backscatter which is similar to the nephelometer. All lidar soundings were obtained at night. The soundings start at about $\frac{1}{2}$ km above the surface because of beam-receiver parallax. Altitudes are with respect to sea level. The gaps in the profiles at about 5 km are due to the time required for automatic gain switching in the acquisition system. Each profile is normalized to a value of 1 (for Rayleigh backscatter) at the point of lowest magnitude of the profile, and at some altitude below 10 km.

The lidar soundings are out of phase with the actual aircraft flights by approximately 4–12 h. The flight numbers shown on Fig. 14 correspond to the flights nearest in time to the lidar profiles. Measurement uncertainties in the profiles are on the order of 50%, being mostly due to the normalization process, which is in the form of a constant bias in magnitude over the entire profile.

A very important bit of information derived from the lidar profiles is the fact that the aerosol layers having high backscatter signals were confined to altitudes ranging from the surface to within a few kilometers above the highest altitudes flown by the aircraft. At Big Spring some layers were noted at 6 and 7 km.
Aerosol layers at 6–7 km were not detected in the Blythe area. Like the aerosol profiles derived from the aircraft data the lidar profiles show a good deal of variability in layering.

Flight 2N was reasonably close in time with a lidar sounding. In Figs. 12 and 14 the three types of measurement show increasing aerosol concentrations with increasing altitude. The two lidar profiles for flight 3N are 5 h apart. A comparison of the two demonstrates the rather rapid change in the vertical structure of the profiles between times 2345 and 0445. It appears that the level of low concentration at 4 km has moved downward about 1 km during the 5 h time period (advective of aerosol layers at two levels can produce the same effect). Also, aerosol concentrations appear to be building up in the lowest levels. It is conceivable that the descending layer can be related to a “rainout” of large particles during the time when turbulent diffusion is nearing a minimum.

In contrast to the average of aerosol profiles obtained by direct sampling with the aircraft systems, the average of the lidar aerosol profiles decreased monotonically with altitude: the average value at the lowest level is 1.8 while at 4 km the average is 1.3. Since the lidar observations were made during nighttime, we may be seeing a real aerosol diurnal change due to the stability of the lower atmosphere.

6. Summary

By use of direct and indirect sampling methods we succeeded in constructing physically consistent, vertically inhomogeneous models of the aerosol characteristics of a turbid atmosphere over a desert and an agricultural region. The individual measurements acted in some cases to verify each other, and in other cases, weaknesses were discovered which would not otherwise have been noted from isolated measurements. Although the number of measurement episodes was small, the derived information was of sufficient quality for use in interpreting the radiative behavior of turbid atmospheres in the absence of clouds.

Data provided by the integrating nephelometer and lidar clearly indicated a high degree of variable vertical structure in aerosol concentration which no doubt relates importantly to the transport of these aerosols both in the vertical and horizontal directions.

We have found very satisfactory agreement on the average between spectral extinction calculated from direct measurements of aerosol size distribution using Mie theory and transmission measurements of the direct solar beam. Comparison of the spectral variation of optical depth of the two types of measurements has resulted in an evaluation of the aerosol sampling instrumentation on an absolute scale as well as particle concentration as a function of size. We conclude that the mean aerosol size distribution in these tests can be represented by a Junge distribution with slope $\gamma = 2$.

Part II of this paper will describe the application of the results of Part I to the interpretation of the radiation flux measurements in terms of the absorption due to the refractive index of aerosols, and to absorption by water vapor.

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