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Surface-based observations of volcanic aerosol effects

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ABSTRACT

Spectral optical depth measurements at Hamilton, Ontario in 1981 and 1983 are used to estimate stratospheric aerosol optical depths and a particle size distribution for the period of maximum stratospheric perturbation by aerosol due to the El Chichon eruption. Results are generally in good agreement with high altitude measurements and estimates. On average, stratospheric optical depth increased by about 0.08 with little spectral variation. Maximum aerosol enhancement occurred between January and May 1983. Mean optical depths for this period were inverted by the constrained linear inversion method of King. The resulting size distribution was similar to the haze *H* form of the incomplete γ distribution. It compared favourably with a distribution obtained by inverting NASA CV990 optical depth measurements and was similar to estimates based on direct aerosol measurements by aircraft and balloon.

1. Introduction

After El Chichon erupted in Mexico in March and April 1982, a stratospheric aerosol cloud spread over much of the northern hemisphere. In mid-latitudes, stratospheric aerosol densities increased throughout 1982 and reached maximum values in the first half of 1983 (Adraini et al., 1983; Hofman and Rosen, 1983; Hay and Darby, 1984). The radiative and climatic effects of this volcanic event are particularly important since the aerosol perturbation was probably the largest this century. The least ambiguous information on the aerosol's radiative effects and properties has been obtained from instrumented aircraft (e.g., Dutton and DeLuisi, 1983) or high altitude stations (e.g., King et al., 1984). Although analysis of surface-based radiometric measurements to show stratospheric aerosol changes is complicated by the presence of a highly variable tropospheric aerosol, broad-band surface radiation measurements have clearly shown turbidity changes due to stratospheric enhancement (Hay and Darby, 1984). In this paper spectral direct beam radiation measurements during 1981 and 1983 at McMaster University in Hamilton, Ontario (43.25°N, 79.87°W) are used to estimate both the effect of the enhanced stratospheric aerosol on optical depths and, by inversion, the corresponding aerosol size distribution.

2. Optical depth determinations

Measurements were made from a rooftop using filter-wheel radiometers designed and built at the University of Arizona and calibrated in Arizona by the Langley plot method in 1980 and 1982. Measurements for 7 wavelengths (400, 500, 610, 670, 780, 889, 1029 nm) with narrow waveband filters (7 to 15 nm) are used. These were made manually every 15–20 min during cloudless periods, between April and August in 1981 and between January and September in 1983. Reference will also be made to measurements between February and April, 1985.

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Spectral optical depth, $\tau(\lambda)$, for wavelength λ is defined by

$$I(\lambda) = I_0(\lambda) \exp[-\tau(\lambda) m], \tag{1}$$

where $I(\lambda)$ is the measured radiation intensity at the ground by a spectral radiometer, $I_0(\lambda)$ the corresponding intensity at zero air mass and m the optical air mass. The optical depth is the sum of components due to absorption and scattering. Measurements were made at wavelengths where water vapour absorption could be neglected. Since intensity is proportional to measured radiometer voltage, V, time-dependent aerosol spectral optical depths $\tau_a(\lambda)$ were calculated from

$$\tau_a(\lambda) = (1/m) \ln[V(\lambda)/V_0(\lambda)] - \tau_R(\lambda, p)\, p/p_0 - a(\lambda)\eta, \tag{2}$$

where $V_0(\lambda)$ is the extraterrestrial, or calibration, voltage; $\tau_R(\lambda, p_0)$ the Rayleigh optical depth at mean sea level pressure, p_0, calculated using Young's (1980) value for the depolarization factor; p the measured surface pressure; $a(\lambda)$ the spectral absorption coefficient for ozone (Vigroux, 1953) and η the ozone content (atm-cm) for an atmospheric column. Ozone data for Toronto were used.

3. Selection of data

Spectral optical depth frequency distributions were determined for 615 observations in 1981 and 1090 in 1983 (Fig. 1). These show systematic increases in 1983 as expected. Since optical depth varies with aerosol transport into an area, atmospheric diffusion of locally produced particulates and particulate growth under favourable humidity conditions, data were selected to minimize effects of these sources. Contributions from local aerosol sources are greatest under light winds while greatest aerosol transport is expected from neighbouring urban areas. To determine directional dependence of turbidity, mean optical depths for 670 nm were calculated for 30° wind direction sectors. Optical depth at this wavelength approximates a spectrally-averaged optical depth (Blanchet, 1982; Davies and Stewart, 1984). Few measurements were made for wind directions between 70° and 210°, which effectively eliminates significant biases due to aerosol contributions from Buffalo to the southeast and Toronto to the northeast. A "clean" sector between 240° and 360° was defined. Results for this sector in 1981 are used as a reference against which 1983 values can be compared to estimate the effect of El Chichon's aerosol. To minimize effects from local pollution sources data collected when wind speeds were less than 1 m/s were rejected, and to minimize effects from both local and imported aerosol data for visibilities less than 24 km were excluded. Our data did not show a dependence of optical depth on relative humidity.

4. Optical depth results

Median values for the 1983 data were larger at all wavelengths by about 0.08 (Table 1). These systematic differences represent an order of magnitude increase in stratospheric optical depth from an estimated value of 0.005 for the unperturbed stratosphere (Toon and Pollack, 1976). Their magnitude is similar to the stratospheric optical depths of 0.08–0.1 determined from aircraft in late 1982 by Spinhirne (1983) and Dutton and DeLuisi (1983). The aircraft determinations also show small spectral variation but are characterized by maximum optical depth values between 500 nm and 600 nm.

Median optical depths for 1985 are systematically larger than those for 1981 by 0.02 to 0.03, indicating a detectable effect from the volcanic aerosol two years after the eruption.

The estimated enhancement of stratospheric optical depth by 0.08 determined from the clean air sector should apply to all the Hamilton data at all visibilities. This was examined by relating optical depth (670 nm) to visibility, V (km). We

Table 1. *Median optical depth values for the clean sector*

λ (nm)	1981	1983	1985	1983–1981	1985–1981
400	0.1398	0.2353	0.1757	0.0955	0.0359
500	0.1187	0.2047	0.1444	0.0860	0.0257
610	0.0845	0.1788	0.1070	0.0943	0.0225
670	0.0805	0.1668	0.1082	0.0863	0.0277
780	0.0630	0.1504	0.0850	0.0874	0.0220
869	0.0648	0.1419	0.0838	0.0771	0.0190
1029	0.0355	0.1187	0.0623	0.0832	0.0268
N	245	475	488		

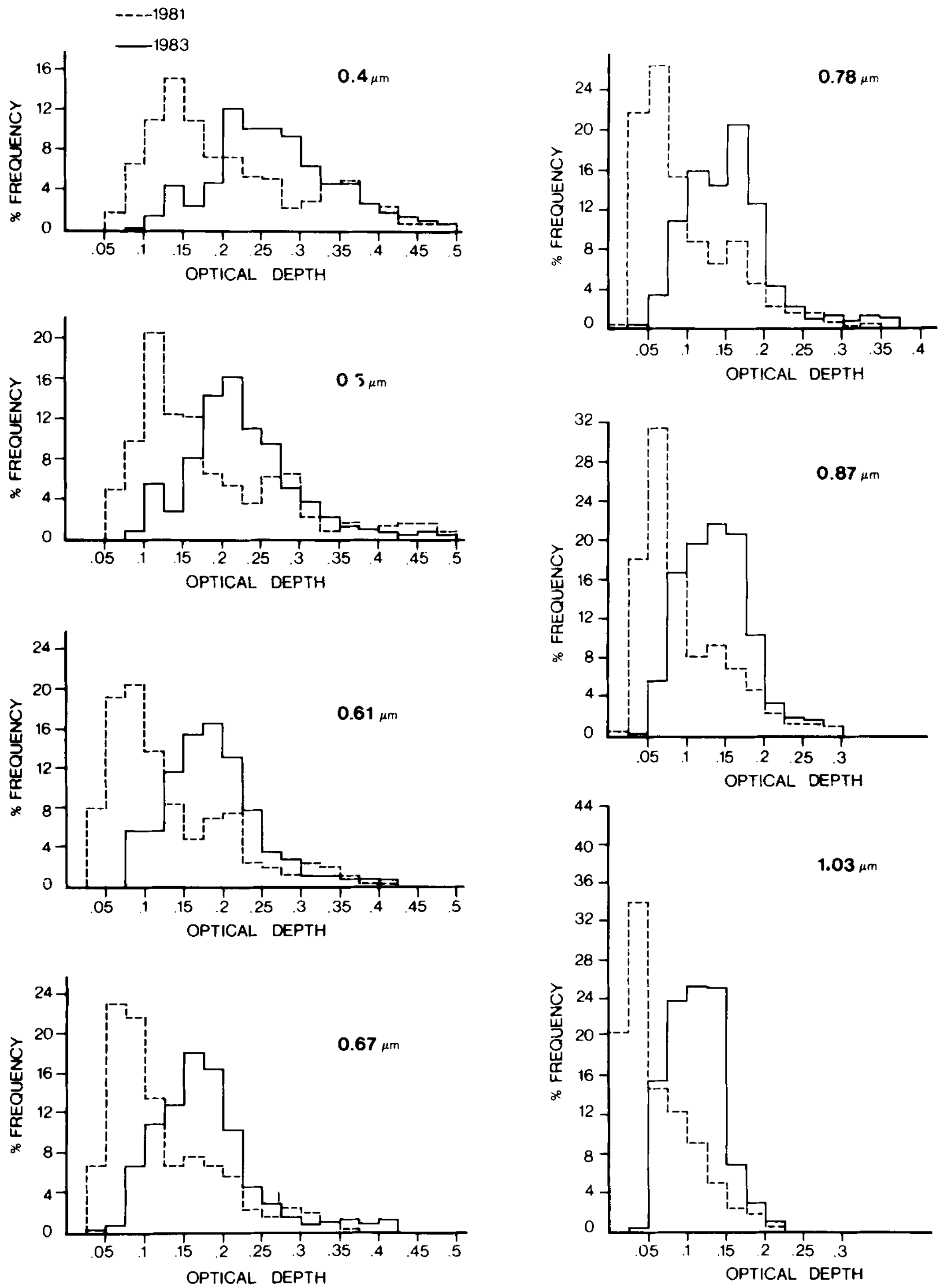

Fig. 1. Frequency distributions of spectral optical depths for 1981 and 1983.

define a scattering (Rayleigh and Mie) coefficient $\beta(\lambda, 0)$ at the surface (subscript 0) as

$$\beta = \beta_R(\lambda, 0) + \beta_M(\lambda, 0). \tag{3}$$

Then,

$$\tau(\lambda) = \beta_R(\lambda, 0)\, H_R + \beta_M(\lambda, 0)\, H, \tag{4}$$

where H_R and H are the scale heights for the molecular atmosphere (8 km) and for aerosol. Using the Koschmieder formula (McCartney, 1976):

$$V = 3.912/\beta(\lambda, 0), \tag{5}$$

$$\tau_a(\lambda) = H(3.912/V - \tau_R(\lambda)/H_R). \tag{6}$$

Mean optical depth values for different visibility classes in 1981 and 1983 are quite well fitted by eq. (6) (Fig. 2) using $H = 0.6$ km. Considering the 1983 optical depths as combinations of a tropospheric component, which is similar to 1981 and 1985 values, and a stratospheric component, which is independent of visibility, the 1983 data can be fitted by adding

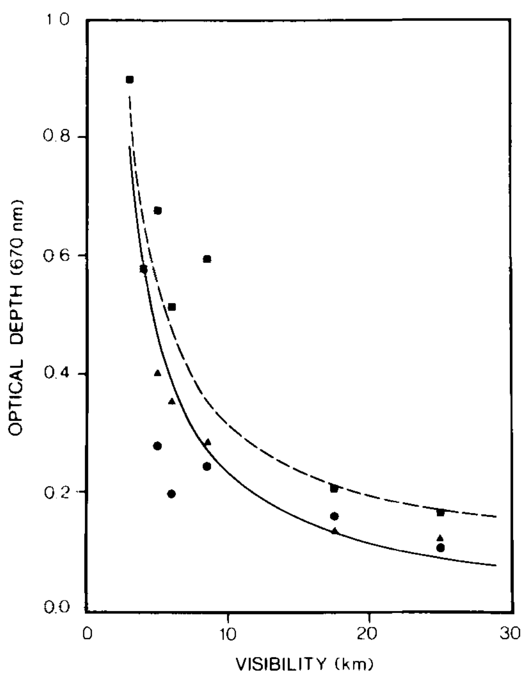

Fig. 2. Correlation between $\tau(670)$ and visibility. The solid line represents eq. (6) and the dashed line eq. (6) corrected for increased stratospheric aerosol. The dots, squares and triangles represent data for 1981, 1983 and 1985 respectively.

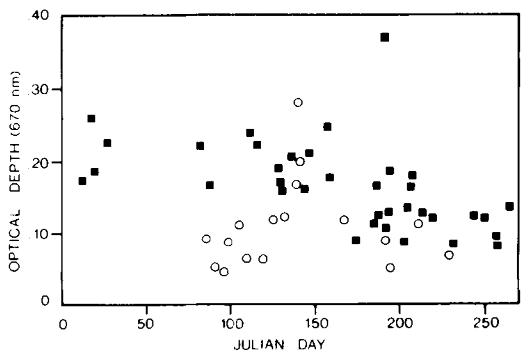

Fig. 3. Variation of $\tau(670)$ in 1981 (circles) and 1983 (solid squares).

a constant (0.08) to eq. (6). The dashed curve in Fig. 2, which represents this modification, provides a reasonable fit to the 1983 data.

The variation of mean daily optical depth for 670 nm in 1981 and 1983 is shown in Fig. 3. There were two regimes in 1983: larger values (0.2 on average) between January and early June which gave way abruptly to smaller values (0.136 on average) for the rest of the measurement period. Differences between optical depths for the two years reach a maximum of 0.127 between March and early May. This value corresponds with the maximum 3° latitudinal daily average optical depth for 675 nm from NASA CV990 flights in late April and early May over the western United States for similar latitudes (data provided by Dr. DeLuisi). The CV990 results also showed little spectral variation nor, in most cases, maximum optical depth in the visible portion of the spectrum. We conclude that our ground-based determinations of stratospheric optical depths are consistent with determinations from aircraft flights at the same latitude.

5. Aerosol size distributions

Since the effect of El Chichon was most apparent in spring 1983, results from inverted size distributions mainly in April for both years will be compared to estimate the stratospheric aerosol size distribution.

The columnar aerosol size distribution, $n_c(r)$, defined as the number of particles per unit radius interval in an atmospheric column of unit cross sectional area, can be obtained from spectral optical depths by inverting the equation:

$$\tau_a(\lambda) = \int \pi r^2 \, Q_{ext}(r, \lambda, \kappa) \, n_c(r) \, dr, \qquad (7)$$

which expresses optical depth as the total particulate optical cross section. Here r is particle radius, $Q_{ext}(r, \lambda, \kappa)$ is the Mie extinction efficiency factor and κ is the complex refractive index for the aerosol. Size distributions were calculated primarily by constrained linear inversion (King, 1982) using Q_{ext} values determined for $\kappa = 1.5 - 0.001i$. The method was selected because an a priori distribution type is not assumed (King et al., 1978; King, 1982; King et al., 1984).

The wavelength range of the optical depth measurements places limits on the size distribution information that can be obtained by inversion. Thomalla and Quenzel (1982) showed that wavelengths smaller than 1.029 μm provide no information on particles with radii greater than about 1 μm. King (1982) also showed that reliable inversion results could not be obtained for radii smaller than 0.08 μm. Within this range size distributions equivalent to an analytical form of 2 or 3 parameters may be derived. Kaufman and Fraser (1983) showed that optical depths calculated from multimodal size distributions yielded Junge size distributions upon inversion. Trakhovsky et al. (1982) question whether this wavelength range can produce forms other than Junge. These limitations are minimized in our application since stratospheric aerosol size distributions are mainly unimodal with radii smaller than 1 μm.

An analysis suggested by Thomalla and Quenzel (1982) showed that relatively few of the wavelengths used in this study would contribute significantly to the total information content. Rewriting eq. (7) in terms of the logarithmic volume distribution $V(r)$ (i.e., $dV/d \log r$):

$$\tau_a(\lambda) = \int \pi (3/4r) \, Q_{ext}(r, \lambda, \kappa) \, V(r) \, d \log r, \qquad (8)$$

plots of the kernels, $(3/4r)\, Q_{ext}$, showed that four kernel maxima overlapped indicating redundancy. Optical depths at 400, 610 and 869 nm contain the essential information in these measurements. Results from constrained linear inversion applied to optical depths for these wavelengths are presented here. They are very similar to results obtained using all wavelengths.

Our best estimate of the stratospheric aerosol size distribution for this period and this latitude was obtained by inverting the differences between mean optical depths for 8 days between March 27 and May 6 in 1981 and means for 9 days between 12 January and 9 May in 1983 (Table 2). The resulting size distribution using optical depths for 400, 610 and 698 nm is shown in Fig. 4a. The indicated errors are maxima and minima obtained from 10 random perturbations of the mean optical depths within the limits of their standard errors.

This distribution is similar in form to the haze H variant of Deirmendjian's (1969) modified gamma distribution (Fig. 4a), which was fitted by the Box and Lo (1976) method using optical depths for all wavelengths.

This estimated size distribution is compared with two independent estimates. Firstly, a size distribution was obtained by constrained linear inversion using NASA CV990 flight data (J. DeLuisi, private communication) for 46°N given in Table 2 (Fig. 4b). Except for the smallest radius the agreement between the two distributions is surprisingly good. Secondly, the Hamilton distribution is compared with two sets of stratospheric aerosol measurements; average wire impactor data from NASA U-2 aircraft flights (K. Snetsinger and G. Ferry, private communication) for latitudes between 37°N and 49°N during March to May, 1983, and balloon-borne optical particle counter data at Laramie, Wyoming (41.33°N) on 12 February (Rosen and Hofman, 1986). In both cases measurements were for one height. Without information on the vertical variation of aerosol through the stratosphere the columnar size distribution could not be esti-

Table 2. *Optical depths used in inversions*

Hamilton				CV990	
Wave-length	1981	1983	1983–1981	Wave-length	46°N 4/4/83
400	0.1484	0.2821	0.1337	380	0.1290
500	0.1198	0.2493	0.1295	500	0.1300
610	0.0801	0.2193	0.1392		
670	0.0770	0.2057	0.1287	675	0.1270
780	0.0557	0.1830	0.1273		
869	0.0611	0.1722	0.1111	862	0.1140
1029	0.0366	0.1417	0.1051		

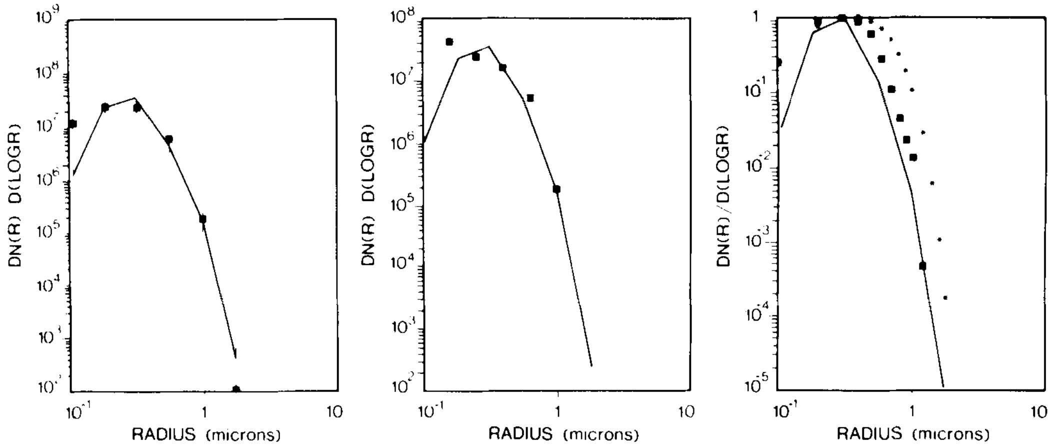

Fig. 4. Stratospheric particle size distribution estimates. In all diagrams the solid line represents the distribution for Hamilton obtained by inversion.
(a) Hamilton —haze *H* —square.
(b) CV990 —square.
(c) U-2 —square.
Balloon —star.

mated from these measurements. Instead each distribution was normalized by its maximum value. Fig. 4c shows that the shapes of the distributions are similar.

6. Conclusion

It is possible to estimate large perturbed stratospheric spectral optical depths and aerosol size distributions from surface spectral direct beam radiation measurements. Such estimations could usefully augment high altitude measurements by providing greater spatial and temporal resolution.

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