Comments on a Paper by A. J. Dyer, 'Anisotropic Diffusion Coefficients and the Global Spread of Volcanic Dust'

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In a recent paper *Dyer* [1970] uses measurements of the decrement in direct solar radiation to trace the global spread of volcanic dust from the 1963 Mt. Agung eruption. He assumes that this dust is carried from a tropical stratospheric reservoir to middle and high latitudes by the stratospheric circulation system. The purpose of this letter is to point out that there is a large discrepancy between the magnitude of the extinction reported by Dyer and the aerosol scattering cross sections measured by the laser radar technique.

The index of attenuation that Dyer uses reduces to ln (I'/I) for an overhead sun, where I' is the value for the direct solar beam at the earth's surface before the eruption and I is the corresponding value in the presence of volcanic dust. This index reaches maximum values of about 0.1 in the northern hemisphere and 0.3 in the southern hemisphere. Typical values of about 0.05 are shown for the northern hemisphere during 1964 and 1965.

Grams and Fiocco [1967] have published a large number of scattering profiles for the stratospheric aerosol taken during 1964 and 1965 at a latitude of 42°N using the laser radar technique. Profiles are also available from 18°N [*Clemesha et al.*, 1966]. After making some reasonable assumptions about the nature and size distribution of the particles, we can attempt to relate the extinction measurements used by Dyer to the backscattering coefficients obtained by laser radar.

To obtain the aerosol extinction coefficient Z_{ext} for a given particle population, we integrate the product of the extinction efficiency Q_{ext} with the geometrical cross section of the particle (assumed to be spherical) across the particle size distribution. Thus we have

$$Z_{ext} = \int Q_{ext}(a)\pi a^2 \, dN(a) \qquad (1)$$

where Q_{ext} (a) is the extinction efficiency for a particle of radius a and $dN(a)$ is the number of particles of radius a per unit volume per unit interval of radius.

Using a Junge type distribution of the form $dN(a)/da = Ca^{-4}$, where C is a constant, and converting to the size parameter $x = 2\pi a/\lambda$, where λ is the wavelength, we get

$$Z_{ext} = \frac{2\pi^2 C}{\lambda} \int Q_{ext}(x)x^{-2} \, dx \qquad (2)$$

If the values of Q_{ext} (x) for particles of refractive index 1.5 published by *Van de Hulst* [1957] are used, the integral in equation 2 taken from zero to infinity can be evaluated as approximately 1.8. For wavelengths in the optical region the main contribution comes from particles in the 0.1- to 1-μ range of radius. Thus we have $Z_{ext} = 3.6\pi^2 C/\lambda$.

If we now assume that the aerosols exist in a rectangular block layer of thickness Δh, we have the atmospheric transmission T given by $T = \exp -(Z_{ext}\Delta h)$. T in terms of the index used by Dyer is given by $T = \exp -\ln (I'/I)$, typically having a value of 0.95. By substituting Z_{ext} and h for T and using $\lambda = 0.55$ μ and $\Delta h = 10$ km, we get a value of 7.9×10^{-14} for C.

We shall now determine the volume backscattering coefficient presented by the particle population derived above.

The volume backscattering coefficient for Mie scattering σ_M is given by

$$\sigma_M = (\lambda/2\pi)^2 \int i_M(a) \, dN(a) \qquad (3)$$

where $i_M(a)$ is the intensity function for Mie scattering through 180° for a particle of radius a. Numerical integration between the limits of 0.01 and 10 μ (again the main contribution comes from the 0.1- to 1-μ region) and use of the methods given by *Van de Hulst* [1957] to compute the intensity function give a volume back scattering coefficient of $1.5 \times 10^{-7} m^{-1}$ $ster^{-1}$ for the 0.6943-μ wavelength used in the laser radar experiments.

The aerosol scattering profiles published by *Grams and Fiocco* [1967] and *Clemesha et al.* [1966] are given in terms of the ratio of the aerosol scattering coefficient to the molecular scattering coefficient. If we use a standard atmosphere to convert these to absolute values of the scattering coefficient, we find peak values of about 7×10^{-8} m^{-1} $ster^{-1}$, but reducing the profiles to a 10-km block layer gives an average scattering coefficient of only 2×10^{-8} to 3×10^{-8} m^{-1} $ster^{-1}$. We can see, then, that, if the values for the attenuation of solar radiation used by Dyer refer to extinction caused by stratospheric aerosols, the laser radar measurements should show an aerosol return almost an order of magnitude stronger than they actually do. Such intense returns are not even occasionally observed.

A possible explanation of this discrepancy might lie in the particle size distribution. If the volcanic dust contains a very large proportion of particles smaller than about 0.1-μ radius, it would be more effective for extinction than for backscattering; this would be particularly true if the particles were absorbent. Such a particle distribution has been suggested by *Pilipowskyj et al.* [1968] to explain the results of simultaneous infrared and laser radar measurements of the stratosphere. It should be pointed out, however, that the inverse fourth-power distribution, which has been used here, previously provided a weighting in favor of small particles, which would not be accepted by many workers. *Friend* [1966], for example, finds a size distribution with a virtual cutoff at 0.3-μ radius. The use of such a distribution would result in an even greater discrepancy between the extinction measurements and the laser radar results.

With regard to a stratospheric reservoir of dust in the tropical zone it is again difficult to reconcile this idea with the laser results. To explain the seasonal variation of the decrement in solar radiation Dyer suggests that the dust is carried preferentially toward the winter hemisphere by the tropical Hadley cell. For this process to continue over several years, a considerable reservoir of dust in the tropical stratosphere is required, but the backscattering measurements for 18°N do not show any greater particle concentration than those for 42°N.

In summary it seems unlikely that the atmospheric transmission measurements are indicative of stratospheric dust content, and it is suggested that the observed decrements in solar radiation are due to tropospheric turbidity. If this was so, the winter maximum would not be connected with the stratospheric circulation. A general increase in the tropospheric aerosol content during recent years might well be the result of increased man-made pollution.

References

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