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TITLE: REDUCTION OF PHOTOSYNTHETICALLY ACTIVE RADIATION UNDER EXTREME STRATOSPHERIC-AEROSOL LOADS

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REDUCTION OF PHOTOSYNTHETICALLY ACTIVE RADIATION UNDER EXTREME STRATOSPHERIC-AEROSOL LOADS

by

Siegfried A. W. Gerstl and Andrew Zardecki

ABSTRACT

The recently published hypothesis that the Cretaceous-Tertiary extinctions might be caused by an obstruction of sunlight is tested by model calculations. First we compute the total mass of stratospheric aerosols under normal atmospheric conditions for four different (measured) aerosol size distributions and vertical profiles. For comparison, the stratospheric dust masses after four volcanic eruptions are also evaluated. Detailed solar radiative transfer calculations are then performed for artificially increased aerosol amounts until the postulated darkness scenario is obtained. Thus we find that a total stratospheric aerosol mass between 1 and 4 times 10^{16} g is sufficient to reduce photosynthesis to 10^{-3} of normal. We also infer from this result that the impact of a 0.4- to 3-km-diameter asteroid or a close encounter with a Halley-size comet may deposit that amount of particulates into the stratosphere. The darkness scenario of Alvarez et al. is thus shown to be a possible extinction mechanism, even with smaller size asteroids or comets than previously estimated.

I. INTRODUCTION

Recently, Luis and Walter Alvarez, Frank Asaro, and Helen Michel (1980) put forth the hypothesis that the Cretaceous-Tertiary (C-T) extinctions (sudden disappearance of dinosaurs and other reptiles about 65 million years ago) might have been caused by the consequences of a large meteorite (asteroid) impacting on Earth. Such an impact would produce a global dust layer that would stay aloft in the stratosphere for several years and thus suppress photosynthesis to a large extent that would explain the major features of the extinctions. The physical basis for this hypothesis was provided by the discovery of an apparently worldwide marine sediment layer that shows an anomalously high abundance of iridium most likely originating from an extraterrestrial source. This iridium anomaly has only recently been located in a terrestrial deposit by a group of Los Alamos National Laboratory researchers (Orth and others, 1981) in a drill core taken in Northern New Mexico.

The Alvarez/Asaro/Michel hypothesis is being studied and tested by many researchers throughout the country. In particular, the impact mechanics and possible climate consequences have been modeled with some success. Several independent arguments have been applied to estimate the size of the asteroid, which Alvarez et al. place at about 10 ± 4 km in diameter. In this report we attempt to estimate a lower limit of the asteroid's mass from solar radiative transfer calculations, assuming the presence of various amounts of dust in the stratosphere that we scale up from measured stratospheric aerosol distributions of volcanic origin. We are also discussing a new hypothesis, which postulates that the required aerosol mass could have been deposited in the stratosphere by a comet, whose nucleus was breaking up in Earth's vicinity (a grazing comet).

Recently, we have established an advanced computational capability for atmospheric solar radiative transfer calculations in the context of an environmental research project (Gerstl and Zardecki, 1981; Zardecki and Gerstl, 1981), where the biologically effective solar irradiance at ground level was computed for clear and polluted atmospheres. Potential effects on plant life as a consequence of reduced biologically effective solar irradiance were also studied. We now apply this computational system and its data base to calculate the photosynthetically active radiation (PAR) that reaches the ground under different stratospheric pollution scenarios. We can thus model the assumed extreme obstruction of sunlight necessary to reduce photosynthesis by several orders of magnitude with resulting extinction. In addition, we can compute the total mass of the stratospheric dust required to achieve an assumed PAR reduction. Using published data on impact mechanics, we can then give a lower limit estimate for the mass and size of the impacting asteroid.

II. DESCRIPTION OF STRATOSPHERIC DUST

A quantitative description of atmospheric pollution and its optical properties is usually based on the assumption of a polydispersion of spherical aerosol particles (Deirmendjian, 1969), which are allowed to vary in number density (particles/cm^3) with varying altitude R (vertical profile). The particles are also allowed to vary in size (radius r, measured in μm) according to a size distribution function n(r) such that n(r)dr gives the number of particles with radii between r and r + dr. The complete description of such an atmospheric polydispersion of aerosol particles can therefore be given by an altitude-dependent distribution function n(r,R)dr, which gives the number of particles per cm^3 volume with radii between r and r + dr at altitude R. If each particle is assumed spherical in shape, then its mass is given by $\frac{4\pi}{3} r^3\rho_p$, where ρ_p is the specific gravity of the aerosol particle. The vertical mass distribution of such an atmospheric polydispersion is then given by

$$dM(r,R) = \frac{4\pi}{3} r^3\rho_p \cdot n(r,R)dr \quad , \tag{1}$$

so that the total mass of all aerosol particles per cm^3 at altitude R is then computed as an integral over all aerosol sizes:

$$M(R) = \frac{4\pi}{3} \rho_p \int_0^\infty r^3 n(r,R)dr \ . \tag{2}$$

Assuming a uniform horizontal distribution of these aerosols over the entire globe, we can then integrate Eq. (2) over the entire volume of the stratosphere to obtain the total mass of stratospheric aerosols.

$$M = \int_{R_1}^{R_2} M(R) \cdot 4\pi R^2 dR \ , \tag{3}$$

where the $R_{2,1}$ indicate the upper and lower boundaries of the stratosphere (assumed to be a spherical shell) measured from the Earth's center. Assuming the same distribution of aerosol sizes at all altitudes in the stratosphere, we can rewrite the altitude-dependent distribution function n(r,R) as the product of a vertical profile N(R) and a size distribution function n(r),

$$n(r,R) = N(R)\cdot n(r) \quad , \tag{4}$$

so that N(R) gives the total number of aerosol particles (of all sizes) per cm^3 at altitude R, whereas n(r) is the size distribution function that is normalized to one,

$$\int_0^\infty n(r)dr = 1 \quad . \tag{5}$$

Both the vertical aerosol profile N(R) and the normalized size distribution function n(r) have been measured for several stratospheric pollution scenarios and many different aerosol characteristics, and representative data are readily available (Deirmendjian, 1969; Shettle and Fenn, 1979). Inserting Eqs. (2) and (4) into Eq. (3) gives the desired expression for the total mass of stratospheric aerosols:

$$M = \int_{R_1}^{R_2} \frac{4\pi}{3} \rho_p \int_0^\infty r^3 N(R)n(r)dr\cdot 4\pi R^2 dR = \rho_p\cdot N_o\cdot V_p \quad , \tag{6}$$

with

$$N_o = 4\pi \int_{R_1}^{R_2} N(R)R^2 dR \tag{7}$$

= Total number of aerosol particles in the entire stratosphere, and

$$V_p = \frac{4\pi}{3} \int_0^\infty r^3 n(r) dr \tag{8}$$

= Volume occupied by a median-size aerosol particle .

The quantities defined by Eqs. (6) through (8) will be evaluated numerically in the next section for four different stratospheric aerosol models that differ from each other by their specific aerosol characteristics (size distribution, absorption, and scattering coefficients) and vertical profiles.

III. OPTICAL DATA FOR FOUR STRATOSPHERICAL AEROSOL MODELS

We used for our radiative transfer calculations the complete set of optical data for atmospheric aerosols recently issued by Shettle and Fenn (1979) of the Air Force Geophysics Laboratory (AFGL). These data are based on a long history of measurements and have been verified against laboratory and satellite data. All necessary optical parameters (scattering and absorption coefficients, vertical profiles and size distributions of tropospheric and stratospheric aerosols, scattering phase functions, etc.) to perform solar radiative transfer calculations in realistic model atmospheres are taken from this AFGL data base in the spectral region of interest (0.35 to 0.75 μm). In addition to several tropospheric aerosol models, the AFGL data describe four specific stratospheric aerosol models that are our main interest:

BS = Background Stratosphere,
MV = Moderate Volcanic,
HV = High Volcanic,
EV = Extreme Volcanic.

Table I lists the vertical distribution of the aerosol number density N(R) (the vertical profile) for all four stratospheric aerosol models from 10- to 30-km altitude. These data from Table I were used to evaluate N_o from Eq. (7).

Three different aerosol size distributions n(r) are used in conjunction with the above stratospheric aerosol models:

BS = Background Stratosphere, 75% H_2SO_4,

FV = Fresh Volcanic,

AV = Aged Volcanic.

All three distributions can be described analytically by a modified gamma distribution function

$$n(r) = ar^{\alpha} \exp(-br^{\gamma}), \quad \text{for } 0 \leq r \leq \infty . \tag{9}$$

The four parameters a, α, γ, b, which fully describe the polydispersions, are listed in Table II. The maximum of the distribution function, Eq. (9), occurs at $r = r_c$ with

$$r_c = \left(\frac{\alpha}{b \cdot \gamma}\right)^{1/\gamma} , \tag{10}$$

and has a value of

$$n(r_c) = ar_c^{\alpha} \exp(-\alpha/\gamma) . \tag{11}$$

The mode radius is r_c, and r_c gives the most frequent radius encountered in the polydispersion, whereas $n(r_c)$ gives an indication of how narrow or broad peaked the distribution function is. Both parameters are also listed in Table II.

When Eq. (9) is inserted into Eq. (8), the integral over the size distribution function can be evaluated analytically so that

$$V_p = \frac{4\pi}{3} \frac{a}{\gamma} b^{-\frac{\alpha+4}{\gamma}} \Gamma\left(\frac{\alpha+4}{\gamma}\right) . \tag{12}$$

Values of V_p for the three specific size distributions under consideration are given in the last column of Table II. Comparing the last three columns in Table II, we may summarize that the aged volcanic distribution model represents the finest grain aerosol particles (r_c = 0.016 μm), whereas the fresh volcanic model describes much larger (r_c = 0.063 μm) and heavier particles ($V_p^{FV} \gtrsim 64\ V_p^{AV}$).

TABLE I

VERTICAL PROFILES OF AEROSOL NUMBER DENSITY FOR FOUR STRATOSPHERIC AEROSOL MODELS

Height (km)	Aerosol Number Density, N(R) (particles/cm³) Background Stratospheric	Moderate Volcanic	High Volcanic	Extreme Volcanic
10.0	1.04E+01*	2.05E+01	1.13E+00	1.13E+00
11.0	7.27E+00	2.34E+01	1.29E+00	1.29E+00
12.0	5.83E+00	2.72E+01	1.50E+00	1.50E+00
13.0	4.70E+00	3.11E+01	1.72E+00	1.72E+00
14.0	4.03E+00	3.21E+01	2.21E+00	2.21E+00
15.0	3.59E+00	3.24E+01	3.21E+00	3.21E+00
16.0	3.47E+00	3.04E+01	4.96E+00	4.96E+00
17.0	3.86E+00	2.73E+01	7.35E+00	7.78E+00
18.0	4.73E+00	2.33E+01	9.32E+00	1.42E+01
19.0	5.29E+00	1.90E+01	9.38E+00	2.97E+01
20.0	5.36E+00	1.50E+01	7.17E+00	6.13E+01
21.0	4.57E+00	1.21E+01	4.35E+00	3.37E+01
22.0	3.82E+00	9.54E+00	2.76E+00	3.74E+00
23.0	2.73E+00	7.32E+00	1.47E+00	1.47E+00
24.0	1.80E+00	5.71E+00	7.84E-01	7.84E-01
25.0	1.19E+00	4.55E+00	4.75E-01	4.75E-01
26.0	8.19E-01	3.55E+00	2.73E-01	2.73E-01
27.0	6.16E-01	2.78E+00	1.78E-01	1.78E-01
28.0	4.71E-01	2.33E+00	1.29E-01	1.29E-01
29.0	3.74E-01	1.38E+00	7.60E-02	7.60E-02
30.0	3.02E-01	8.43E-01	8.43E-01	8.43E-01

*read 1.04×10^1

TABLE II

PARAMETERS OF AEROSOL SIZE DISTRIBUTIONS CONSIDERED

Stratospheric Aerosol Model	a $(\mu m)^{-2}$	α (1)	γ (1)	b $(\mu m)^{-\gamma}$	r_c (μm)	$n(r_c)$ $(\mu m)^{-1}$	V_p $(10^{-14} cm^3)$
Backgr. Strat. (BS)	324.0	1	1	18	0.056	6.62	1.73
Aged Volcanic (AV)	5461.3	1	0.5	16	0.016	11.5	1.51
Fresh Volcanic (FV)	341.3	1	0.5	8	0.063	3.06	96.6

The total mass of all aerosols in the stratosphere, Eq. (6), can now be evaluated for all four models with the data from Tables I and II. In all cases, we chose the value of 3 g/cm³ as a typical specific gravity ρ_p for the aerosol particles. Our results are given in Table III. It should be noted that the fresh volcanic size distribution is assigned to both the high (HV) and extreme volcanic (EV) models, whereas the fine-grain aged size distribution is assigned only to the moderate volcanic (MV) model. These assignments are an integral part of the data base described by Shettle and Fenn (1979).

TABLE III

TOTAL NUMBER AND MASS OF STRATOSPHERIC AEROSOLS FOR FOUR COMPUTER MODELS AND FOUR VOLCANIC ERUPTIONS

Stratospheric Aerosols	Type of Aerosol	$N_o \times 10^{25}$ (Particles)	M for ρ_p = 3 g/cm³ (g)
BS Model	75% H_2SO_4 droplets	3.8	1.9×10^{12}
MV Model	Aged Volcanic	16.9	7.7×10^{12}
HV Model	Fresh Volcanic	3.1	9.0×10^{13}
EV Model	Fresh Volcanic	8.9	2.6×10^{14}
St. Helens (1980)	Vol. of 2.7/600 km³		1.3×10^{13}
Agung (1963)	Vol. of 9×10^{-3} km³		2.7×10^{13}
Katmai (1912)	Vol. of 1.34×10^{-2} km³		4.0×10^{13}
Krakatoa (1883)	Vol. of 3.0×10^{-2} km³		9.0×10^{13}

For completeness we reproduce (Figs. 1 and 2) the wavelength dependency of typical optical data as used from the AFCL data base. Figure 1 shows the spectral variation of the extinction coefficients for the three stratospheric aerosol types as described in Table II, whereas in Fig. 2 their single scattering albedo is plotted. Because we will limit our calculations to the visible wavelength region (0.35 to 0.75 µm), we may characterize (Fig. 2) the fresh volcanic aerosols as more absorbing than the moderate volcanic type, whereas the background stratospheric model is almost exclusively scattering. These characteristics are further underlined by the specific values of r_c and V_p in Table II and the fact that a 75% solution of sulfuric acid in water has been assumed for the BS model (Shettle and Fenn, 1979).

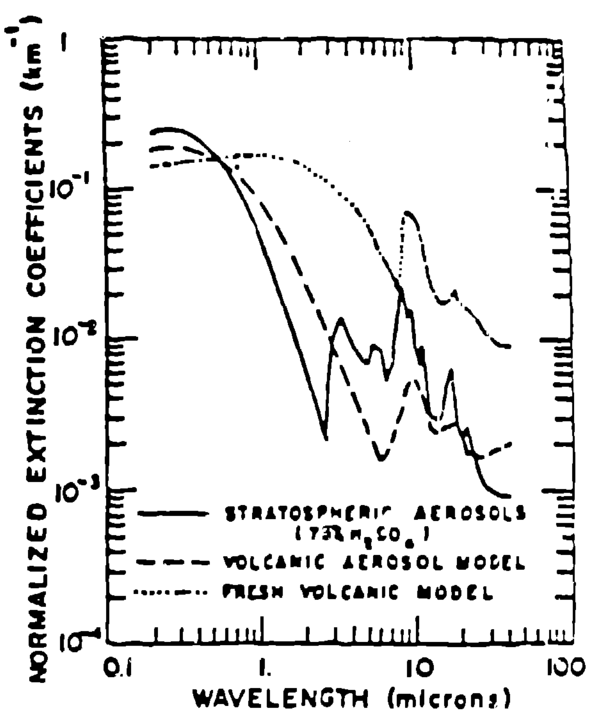

Fig. 1.
Wavelength dependence of the extinction coefficient for three different stratospheric aerosol types. From Shettle and Fenn (1979).

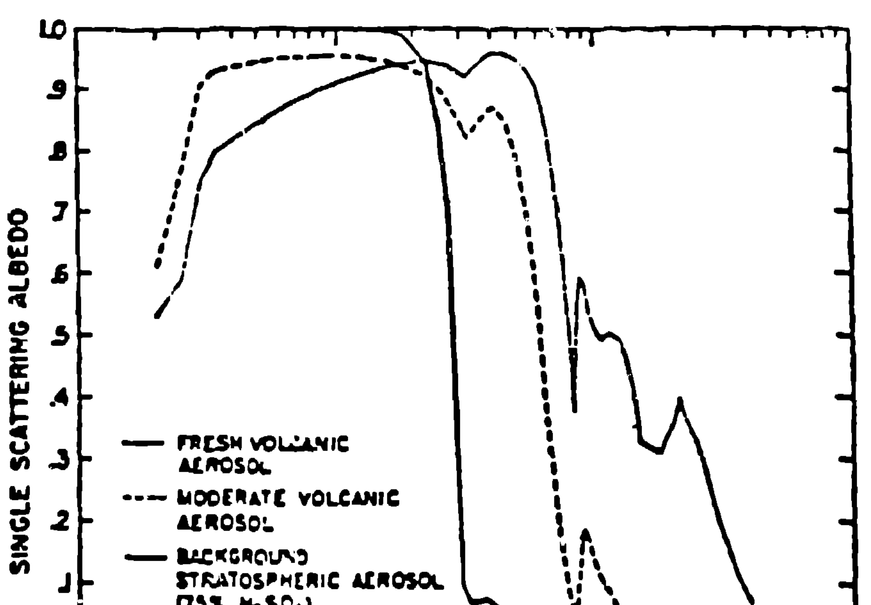

Fig. 2.
Wavelength dependence of the single scattering albedo for three different stratospheric aerosol types. From Shettle and Fenn (1979).

IV. COMPARISON WITH FOUR VOLCANIC ERUPTIONS

To develop some feeling for the magnitude of the total masses of stratospheric aerosols as calculated in the last section and listed in Table III, we include in Table III the stratospheric aerosol masses as estimated for the four largest volcanic events recorded on Earth over the last 100 years. The volume of ejecta that reached the stratosphere in the Agung, Katmai, and Krakatoa eruptions has been derived by Deirmendjian (1973) from recorded measurements of optical phenomena. In particular, the increased optical thicknesses as derived from reduced transmissivities of sunlight were used to estimate the additional stratospheric turbidity created by these eruptions. Deirmendjian consistently derives a ratio of 1:600 for stratospheric dust to total mass of ejecta for the three eruptions, which leads him to the volumina listed in Table III. The volume of 2.7 km^3 of estimated total ejecta for the most recent Mt. St. Helens eruption is taken from R. and B. Decker (1981). Using the 1:600 ratio and $\rho_p = 3\ g/cm^3$ leads us to the estimate of 1.3×10^{13} g of aerosol mass injected into the stratosphere by the St. Helens eruption.

The largest volcanic event in recent history, the Krakatoa eruption in 1883, deserves some more attention because the Krakatoa Committee of the Royal Society estimated in its report (Symons, 1888) "the initial mass of material contained in the airborne dust to be equivalent to some 4 km^3 of solid matter expelled from the volcano," out of a total of 18 km^3 for all the ejecta. This ratio of 4:18 for dust to total ejecta is in conflict with Deirmendjian's ratio of 1:600. Recognizing this discrepancy, Deirmendjian (1973) writes: "We note in passing that if we accept the Krakatoa Committee's own estimate of 4 km^3 of material injected into the stratosphere as dust,...we arrive at...a dust veil with the fantastic optical thickness of $\tau_D = 80$ over the entire intertropical zone!" He concludes then: "Thus the total mass of either the Katmai or the Krakatoa atmospheric dust could hardly have exceeded 10^{-8} of that of the entire atmosphere (5.14×10^{21} g) or the large portions thereof that were affected, whereas the corresponding total optical thickness was little more than doubled with respect to that found under cloudless and very clear conditions away from urban pollution centers." Our detailed radiative transfer calculations, described in the next section, confirm Deirmendjian's conclusion. We accept therefore his estimates of airborne dust masses with long residence times for the Krakatoa, Katmai, and Agung events. The comparison of these stratospheric aerosol masses with the mass of the entire atmosphere appears quite instructive. We might add also, for comparison, that the total air mass in the stratosphere between 10- and 30-km height, where we assume our model aerosols to reside, is 1.7×10^{21} g of which the Krakatoa dust mass is only a very small fraction, namely 5.3×10^{-8}.

V. SOLAR RADIATIVE TRANSFER CALCULATIONS

Having defined our stratospheric aerosol models we intend to compute the spectral distribution of the solar irradiance reaching the ground (z = 0).

$$F(z=0,\lambda) = \int_0^1 \psi(z=0,\lambda,\mu)\cdot\mu d\mu \ , \tag{13}$$

where $\psi(z,\lambda,\mu)$ is the radiance at height z,* which we obtain as the solution of the radiative transfer equation. The discrete-ordinates computer code ONETRAN (Hill, 1975) is applied for the numerical solution, which includes all orders of scattering and allows highly anisotropic phase functions to be considered. An S_{40} approximation was employed that resolves the angular distribution of the radiance with 40 discrete directions. The entire visible region of the solar spectrum was divided into 42 wavelength groups between 0.35 and 0.75 μm, and the atmosphere was described with 31 horizontal zones up to 70-km height. In addition to the stratospheric aerosols already described in the previous section, we considered standard atmospheric conditions (midlatitude summer) above and below the stratosphere including molecular absorption (mainly ozone), Rayleigh scattering, and a constant distribution of rural-type aerosols in the tropospheric boundary layer corresponding to 5-km visual range at the surface. The addition of this constant aerosol layer in the troposphere should simulate the continuous settling of heavier stratospheric aerosol particles. We also assumed a constant 80% relative humidity in this tropospheric boundary layer up to 2-km height.

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*In Sec. II, we used the symbol R to denote the altitude measured from the center of the Earth. Here z is the height above sea level.

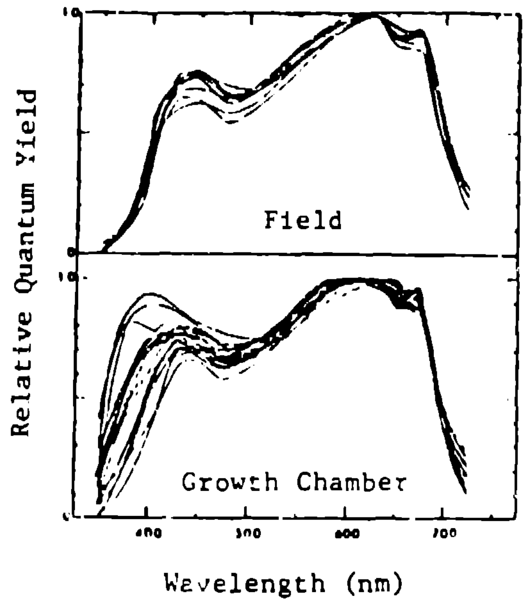
Because we are interested in effects of reduced solar irradiance on plants at ground level, we extracted from the spectral distribution of the solar irradiance the part that is active in the process of photosynthesis. This photosynthetically active radiation (PAR) is defined in photobiology (Larcher, 1980) as

$$PAR = \int_{\lambda_1}^{\lambda_2} E(\lambda)F(z=0,\lambda)d\lambda \ , \tag{14}$$

where $\lambda_{1,2}$ are the wavelengths for which the action spectrum $E(\lambda)$, for photosynthesis of a given plant, tends to zero. For example for the data by McCree (1971), $\lambda_1 = 0.35$ μm and $\lambda_2 = 0.75$ μm for all crop plants. PAR may also be described as the biologically effective irradiance (BEI) for that plant. Recently we performed computations of PAR for four different agriculturally important plants (corn, wheat, soybeans, and green algae) and evaluated quantitatively the effects of increased tropospheric aerosol loads on PAR (Gerstl and Zardecki, 1981; Zardecki and Gerstl, 1981). Considerable reductions of PAR have been found as aerosol concentrations in the 2-km-high tropospheric boundary layer are increased. Typically, the PAR for corn is reduced by 33% (53%) when the air pollution by rural (urban) aerosols increases so that the surface visual range of nominally 300 km for clear air is reduced to 5 km. The computed PAR reductions for the other three plants agree with those of wheat within 10%, which indicates that PAR according to Eq. (14) is insensitive to the choice of the four plants' action spectra $E(\lambda)$. In fact, the measured photosynthesis action spectra for 21 species are very similar to each other as shown in Fig. 3, which is reproduced from the work of McCree (1971) for completeness.

We report here results of PAR calculations for a scenario where only the total amount of the stratospheric aerosols at heights between 10 and 30 km, as described in Sec. III, are varied. The composition of the rest of the atmosphere up to 70-km altitude and down to ground level remains unchanged. We chose this scenario because we are trying to identify situations that can lead to PAR reductions maintained over time periods of up to several years, but at least one growing season. Tropospheric rural-type aerosols are also expected to be present in such situations but not at an easily identifiable constant

Fig. 3.
Spectral quantum yields for photosynthesis of 21 species. From McCree (1971).



concentration due to frequent rainout and continuous settling-out of heavier stratospheric aerosol particles.

In Fig. 4 we give the calculated reductions of PAR versus increased total amounts of stratospheric aerosols for the four different aerosol models described in Sect. III. The PAR for a normal stratospheric aerosol load, $(PAR)_0$ as computed for the background stratospheric aerosol model with a total aerosol mass of 1.9×10^{12} g, is used as the reference value and is thus set to 1.0 in Fig. 4. As a representative photosynthesis action spectrum we chose that of corn (McCree, 1971), and all data in Fig. 4 are for an effective solar zenith angle of 52°, which is a reasonable global average used in many climatology calculations, e.g., Schneider and Dickinson (1974). Results for other solar zenith angles (40 to 60 degrees) deviate only insignificantly from the 52° results, especially for the high stratospheric aerosol masses of interest.

After computing PAR for the four aerosol models described in Sec. III, we multiplied their vertical number densities by a series of increasing numbers m (m > 1.0) and repeated all four calculations for each m. This scaling-up of the stratospheric aerosol amounts was continued up to $m = 10^5$. For a total stratospheric aerosol mass of 9×10^{13} g, which corresponds to the estimated airborne dust mass of the Krakatoa eruption, we obtain thus a PAR reduction between 10 and 25% for the three volcanic models. For comparison, Bullard (1976) cites a 20% reduction in solar radiation following the eruption of Katmai.

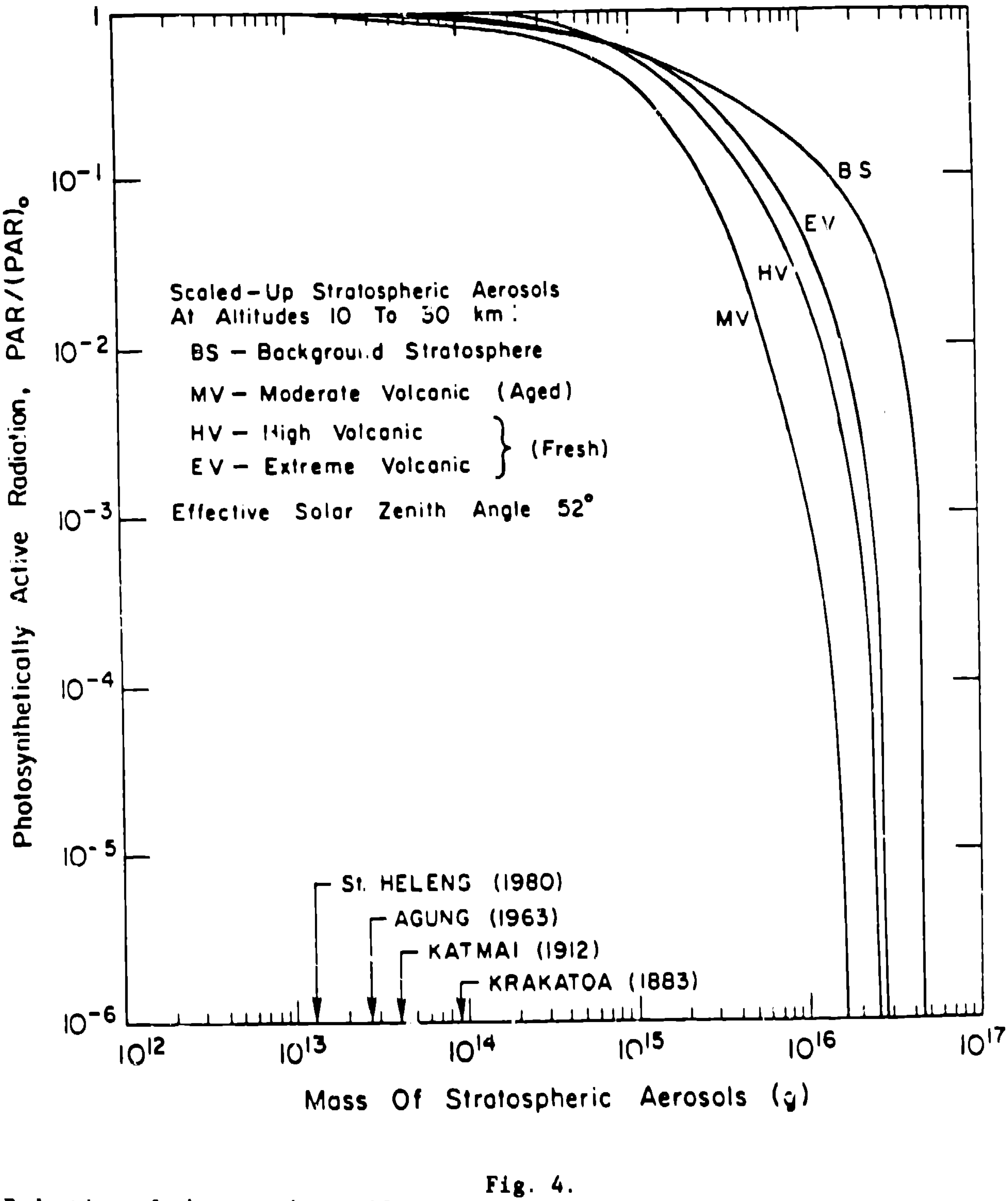


Fig. 4.
Reduction of photosynthetically active radiation resulting from increased stratospheric aerosols for four different aerosol models.

However, a constant reduction of the photosynthetically active radiation of about 25% or less during a growing season affects most agricultural crop yields only marginally (if at all) because the photosynthesis rate of most plants (all C_3 plants) is already saturated at much less (40 to 60%) than the normal full solar irradiance (Lawlor, 1979). In contrast, if we were to take the Krakatoa Committee's airborne dust estimate of 4 km^3 or 1.2×10^{16} g (with our assumed density of 3 g/cm^3), PAR would be reduced to 3% of normal if the EV aerosol model is assumed, or even to 0.03% if the MV model aerosol characteristics are assumed. Such large reductions in the biologically effective solar irradiance would certainly have been reported even if they had happened only in a small latitude band around the equator. Our results for the scaled-up background stratosphere model give the lowest PAR reductions for a given aerosol mass because the BS aerosol composition is assumed to be a 70% solution of H_2SO_4 in water droplets (Shettle and Fenn, 1979), which has very little absorption in the visible as seen from Fig. 2.

With the objective of simulating a polydispersion of pure water droplets in the stratosphere (e.g., due to ejected water from an ocean-impacting asteroid), we repeated the above sequence of radiative transfer calculations for the BS and MV models setting the absorption coefficients equal to zero. For such scaled-up nonabsorbing stratospheric aerosols we obtained the PAR reductions plotted in Fig. 5 which are very similar to the results of Fig. 4. We conclude from this model calculation that it is mainly the multiple scattering effect of these airborne particulates or droplets that is responsible for the PAR reductions and, as far as photosynthesis is concerned, it is not critical whether an asteroid impact on land or in water is assumed. Figure 5 also gives the total normal optical thickness at 0.55 μm for these scaled-up atmosphere models and relates the total mass of all stratospheric water droplets to the mass of the entire atmosphere.

VI. CONNECTIONS WITH AN ASTEROID OR COMET IMPACT

The mechanism that may have produced the Cretaceous-Tertiary extinctions, as described by Alvarez and others (1980), postulates an asteroid impact that ejected enough dust into the atmosphere to reduce photosynthesis drastically on the entire globe and for an extended period of time. As mentioned by other authors, such a darkness scenario might have been generated also by a collision

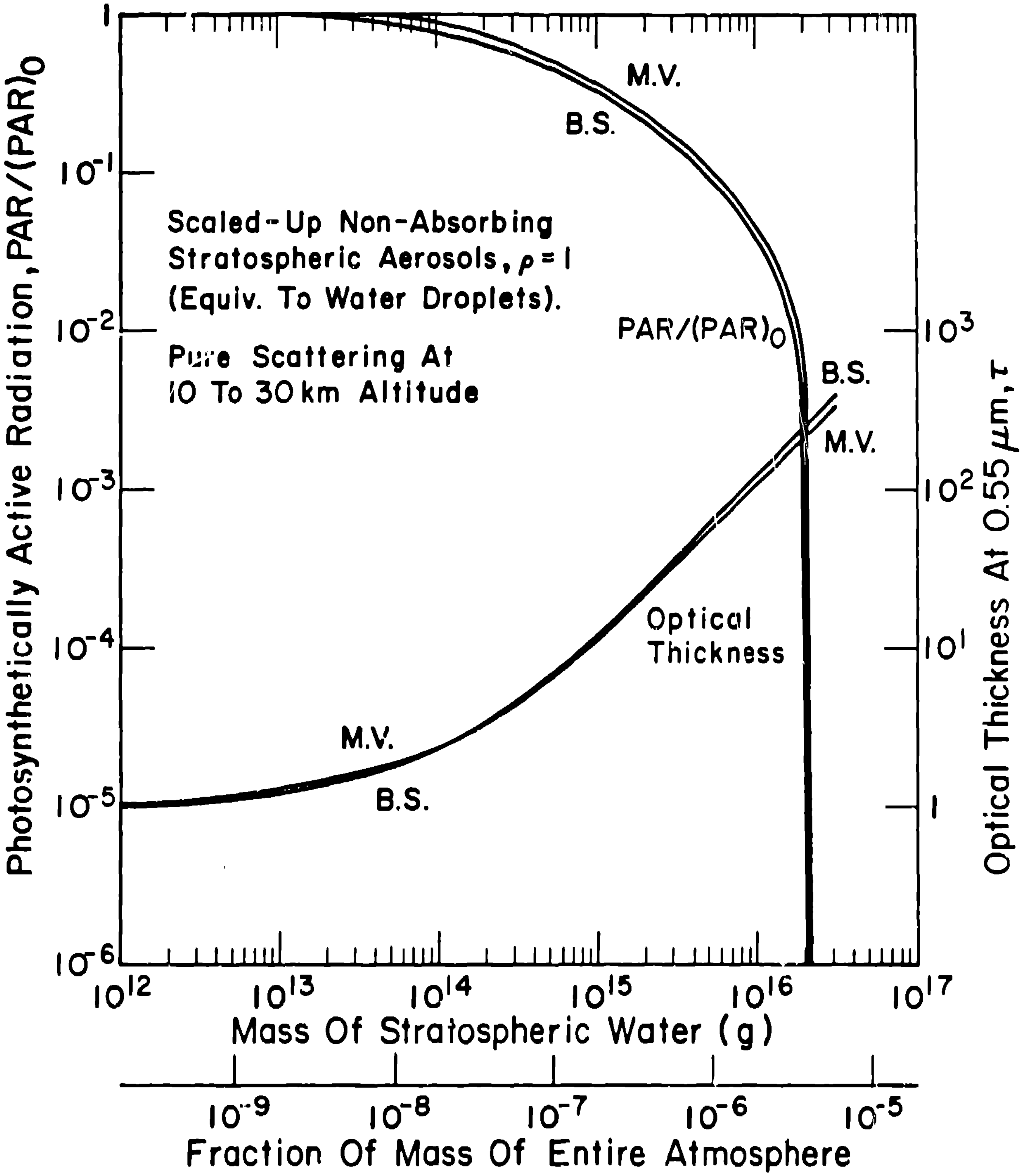


Fig. 5.
Reduction of photosynthetically active radiation (PAR) resulting from simulated pure water droplets in the stratosphere.

with a comet or by a super Tunguska event. We will comment on these hypotheses, which are based on the darkness scenario, in light of the results of our modeling.

From Fig. 4 we can obtain directly the total mass of stratospheric aerosols that is required to reduce PAR to a given fraction of the normal amount. Although it is not clear how much reduction in PAR is required to produce the C-T extinctions, we will assume here that a reduction to 1/1000 of normal is sufficient to initiate the extinction mechanism postulated by the darkness hypothesis, if this darkness persists over at least one full growing season worldwide. With this assumption, Fig. 4 shows that a minimum mass of 10^{16} g in the stratosphere is sufficient to produce a PAR reduction to $10^{-3} \times (PAR)_0$ if the aged volcanic aerosol characteristics are assumed. This minimum mass is increased to 4×10^{16} g if the background stratospheric aerosol model is used. Note that, because of the steep drop of all four curves in Fig. 4 for large aerosol loads, the assumption of PAR = $10^{-3}(PAR)_0$ to be sufficient for the C-T extinctions, is not critical. In fact, an order of magnitude over- or underestimation would change the minimum required total stratospheric aerosol mass insignificantly, even if an attenuation of sunlight of the order of 10^{-7} is assumed, as Alvarez and others (1980) do. Therefore, in the following discussion, we use,

$$M^{STRAT.}_{min.} = 1 \times 10^{16} \text{ to } 4 \times 10^{16} \text{ g} . \tag{15}$$

A. The Minimum-Size Asteroid

To estimate the minimum size of a single solid asteroid capable of depositing on impact with Earth 1 to 4 times 10^{16} g of aerosols or dust into the stratosphere, we require additional information on the impact mechanics of such large objects. O'Keefe and Ahrens (1981) have recently concluded from studying such impact mechanics that a bolide of mass M striking Earth could deposit ejecta in the stratosphere of total mass between 1 and 100 times the bolide mass, that is,

$$\alpha = \frac{M^{STRAT.}}{M^{ASTER.}} = 1 \text{ to } 100 . \tag{16}$$

For completeness, the data from which this estimate is derived are reproduced in Figs. 6 and 7 (from O'Keefe and Ahrens, 1981).

Assuming a spherical asteroid with volume $V_A = \frac{4\pi}{3} R_A^3$ and an average density of $\rho_A = 3\ g/cm^3$, we can easily compute with Eqs. (15) and (16) the diameter $D_{min.}^{ASTER.}$ of a minimum-size asteroid capable of injecting the aerosol mass $M_{min.}^{STRAT.}$ into the stratosphere.

(a) We obtain a lower limit for $D_{min.}^{ASTER.}$ of 0.4 km if the lowest value from Eq. (15) is taken together with the estimated maximum value of $\alpha = 100$ from Eq. (16).

(b) An upper limit for $D_{min.}^{ASTER.}$ of 2.9 km is obtained by using the other two extreme values from Eqs. (15) and (16).

Summarizing, we can write

$$D_{min.}^{ASTER.} = 0.4 \text{ to } 2.9 \text{ km}, \tag{17}$$

which is less than the 10- ± 4-km estimate of Alvarez et al. based on other arguments. However, this difference is no contradiction because $D_{min.}^{ASTER.}$ gives only the minimum size of the asteroid required to support the darkness scenario, whereas any larger asteroid would also initiate this extinction mechanism. Our relatively small minimum asteroid size may help to search for the as yet unidentified impact crater of an estimated minimum diameter between 8 and 60 km, using crater scaling as discussed by Wolf and others (1980). It would be highly desirable, of course, to reduce the large spread of α-values to be able to further narrow down the estimated minimum values for the size of the asteroid as well as its impact crater.

B. A Cometary Collision or Close Encounter

Because of the relatively small mass of stratospheric aerosols required to produce the postulated darkness, mechanisms other than the asteroid impact appear also possible or even likely to explain the deposition of $M_{min.}^{STRAT.}$ into the stratosphere. The most likely such mechanism appears to us to be a collision or close encounter with a comet. Because most comets are believed to consist of about equal masses of dust and icy constituents (Whipple, 1976; Hughes, 1977; Richter, 1954), it is most likely that most cometary particles

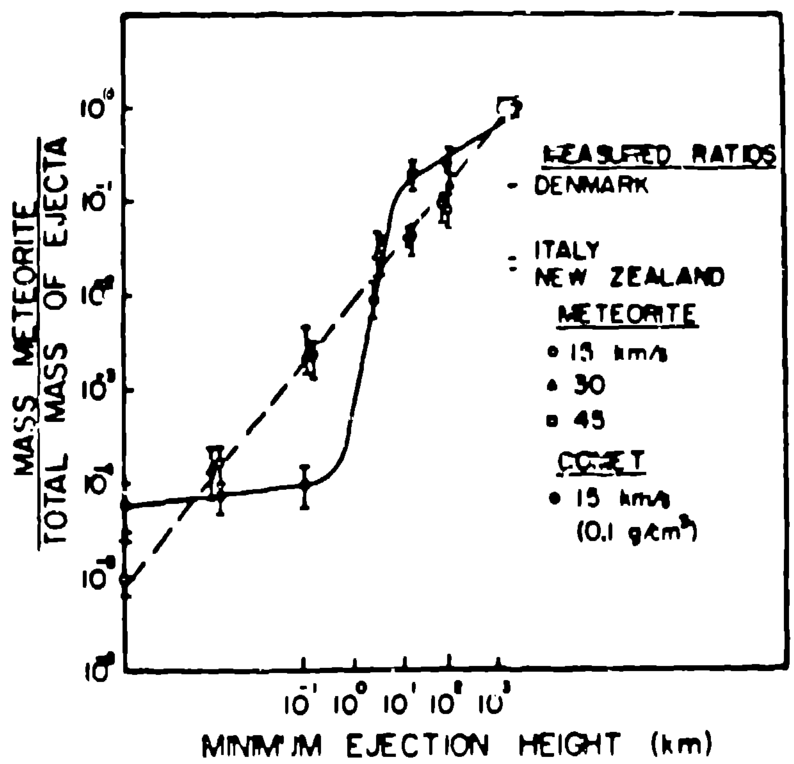


Fig. 6.
Meteorite mass (extraterrestrial material) divided by the total mass of ejecta lofted to at least a given height in the Earth's gravity field for silicate (2.9 g/cm³) and water (0.1 g/cm³) objects impacting a silicate earth. From O'Keefe and Ahrens (1981).

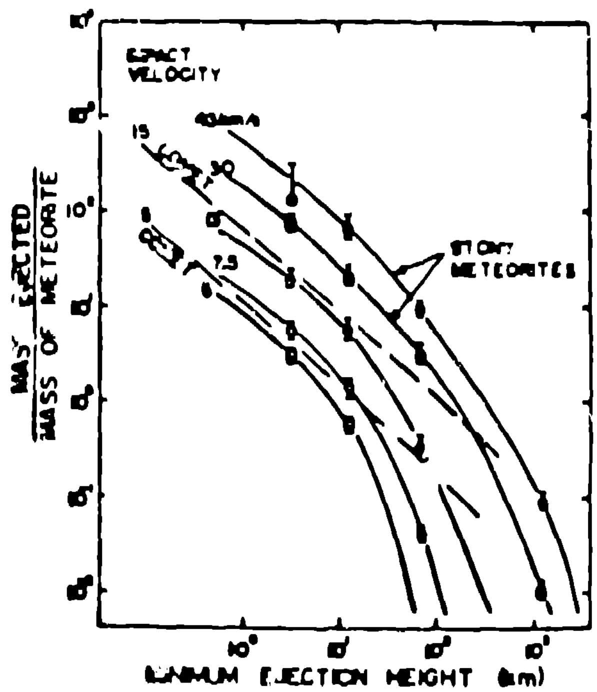


Fig. 7.
Total mass ejected (normalized to impactor mass) versus minimum ejection height in the Earth's gravity field. Impact of silicate (2.9 g/cm³) and cometary (0.1 g/cm³) objects onto silicate half-space is assumed. From O'Keefe and Ahrens (1981).

would never reach the Earth's surface in the case of a collision with Earth as is believed to have occurred on 30 June 1908 over Tunguska, Siberia (Brown and Hughes, 1977). Hence, no impact crater would be produced. In addition, the statistics of observed comets support the assumption of a much higher probability for a comet/Earth collision than for an asteroid impact (Hsü, 1980), especially if the newly estimated values for $M_{min.}^{STRAT.}$ are considered. We conjecture in the following that even a near miss allows the Earth's atmosphere to pick up enough particulate matter from the comet to produce the darkness scenario.

Up to 1950, a total of 43 periodically appearing comets were observed to complete at least 2 solar orbits (Richter, 1954), and Halley's comet alone came within the visual range of Earth on 29 orbits. The average mass of all observed comets has been estimated from photometric measurements to lie between 10^{16} and 10^{17} g. For Halley's comet, four additional independent mass determinations are cited by Richter (1954), which lead to 3×10^{19} g for this well-studied comet; more recent estimates arrive at about 5×10^{16} g ("International Halley Watch", 1980). For example, only a fraction of the total estimated mass of Halley's comet need be deposited in the stratosphere to create the postulated darkness scenario. If we assume that the nucleus of a Halley-size comet breaks up in the Earth's gravitational field outside the atmosphere, then the friction between the cometary fragments and the atmosphere can provide a dispersal mechanism for the cometary dust particles that would not require any direct impact of the comet on the Earth's surface. In fact, as the collision angle approaches 90 degrees (grazing incidence), the atmospheric dispersion mechanism reaches maximum efficiency. This deposition of fine particulates, gaseous, and plasma constituents in the upper atmosphere could provide a reasonable mechanism to create the postulated darkness scenario and produce the observed iridium anomaly at the C-T boundary without requiring the existence of an impact crater. The underlying assumption is, of course, that the dust-to-gas ratio in the comet's coma and nucleus is high enough to produce the observed C-T boundary layer after fallout. Both present-day views of comets as dirty snowballs or dust swarms (Hughes, 1977) do not exclude this possibility. The energetic effects of a 10^{18}-g comet colliding with Earth have been estimated by Urey (1973), who found that, if all energy were absorbed, it would be sufficient "to throw a mass of 3.24×10^{19} g in a circle about Earth." From such energy balance considerations alone Urey concluded that "it does seem possible and even probable that a comet collision with the Earth destroyed the dinosaurs and initiated the Tertiary division of geologic time." Kyte and others (1980) favor the super-Tunguska mechanism suggesting "that an object in a nonintersecting orbit broke up into a large number of objects as it passed inside the Roche limit, with a portion of the debris entering new orbits that intersect the Earth's surface." Obviously, detailed modeling calculations of how comets or other objects behave in close proximity to Earth could shed much light on the above hypotheses. A decisive ruling between the asteroid or comet hypotheses might be expected from measurements of the iridium mixing ratios at the C-T boundary by determining the relative amount of terrestrial material contained in that layer (which must be larger for the asteroid than for the comet impact).

VII. CONCLUSIONS

Our atmospheric radiative transfer calculations indicate that a total mass of aerosols between 1 and 4 times 10^{16} g distributed globally over the Earth's stratosphere is sufficient to reduce the photosynthetically active solar radiation at ground level to 1/1000 of normal. An equivalent amount of dust could be deposited into the stratosphere as ejecta from an impacting asteroid between 0.4 and 3 km in diameter, or as the consequence of an Earth/comet collision or close encounter. The resulting darkness on Earth's surface alone would be sufficient to initiate the Cretaceous-Tertiary extinction mechanism hypothesized by Alvarez and others (1980) if it persisted for at least one growing season. The superposition of other extinction mechanisms like a heat flash followed by substantial climate modifications, as discussed by Emiliani (1980), would add to the stress situation in the biosphere. Because our results determine only the minimum required stratospheric aerosol mass to produce the darkness scenario, we find no contradiction between the estimate of Alvarez and others (1980) of 10 ± 4 km for the asteroid diameter and our minimum-size estimate of 0.4 to 3 km. We extend, however, the number of impact craters on Earth that could possibly be related to the C-T event. Even a close encounter with a Halley's-size comet becomes a plausible source for the deposition of 1 to 4 times 10^{16} g of fine-grain particulates into the stratosphere without producing any sizable impact crater.

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