The Interaction of the Cretaceous/Tertiary Extinction Bolide with the atmosphere, ocean, and solid Earth

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ABSTRACT

The mechanics of large-scale (~10-km diameter) asteroidal, cometary, and meteoroid swarm impact onto a silicate Earth covered by water and a gaseous atmosphere demonstrate that only ~15% to ~25% of the energy of the 15 to 45 km/s bolide is taken up directly during the passage through the ocean and atmosphere, respectively. Upon impact with the Earth, ~10 to 10^6 times the bolide mass of water or rock can be ejected to the stratosphere; however, only ~8.1 bolide masses is in < 1 μm particles. The vaporized, melted, and (< 1 mm) solid ejecta transfer up to ~40% of their energy to the atmosphere and possibly oceanic surface water, giving rise to a short, possibly lethal (to large animals) heating pulse. The initial high-speed ejecta that looted to and above the stratosphere early in the cratering flow is enriched in bolide material and has concentrations of extraterrestrial material in the range of those measured (0.01 to 0.2) in the Cretaceous/Tertiary (C/T) boundary layer. We suggest that the depth of the C/T boundary layer is this ejecta, which is heavily shocked and in the ≤3 μm range and, hence, once entrained in the stratosphere may be spread worldwide. Penetration of the atmosphere by the bolide creates a temporary hole in the atmosphere surrounded by strongly shocked air. The resultant downward and upward flow of the shocked atmosphere backward along the bolide trajectory lofts the vapor, fine-melted and solid ejecta to heights greater than 10 km. The larger, millimeter-size to centimeter-size, melt droplets that are lofted by this mechanism reenter the atmosphere and may represent microtektites and tektites. Sufficient impact-induced vapor, melted and comminuted silicate is ejected to stratospheric heights to markedly reduce the light levels at the Earth's surface. The short-term effects of heating, followed by dust and possibly water-cloud deck induced worldwide cooling, provide several mechanisms to cause severe environmental stress to biota and possibly give rise to the varied and massive extinctions that occurred at the C/T boundary.

INTRODUCTION

Physical evidence for the impact of an extraterrestrial object that is contemporaneous with the Cretaceous/Tertiary (C/T) boundary (65 Ma) has been provided by Alvarez and others (1980), Ganapathy (1980), Hsu (1980), Kyte and others (1980), Ganapathy and others (1981), and Orth and others (1981). The evidence is contained in the 1- to 5-cm layer marking the C/T boundary. This layer is enriched relative to crustal abundances in noble metals such as iridium, osmium, gold, platinum, rhodium, ruthenium, and palladium, and also in nickel and cobalt by factors of 5 to 1,000. It has a siderophile element abundance pattern that is compatible with a concentration of extraterrestrial material in the range of 1% to 2% (Kyte and others, 1980). The layer, which marks this event, appears to be global in extent; it has been found in marine sediments in five localities near Guibbo, Italy; Stevns Klint and Jutland,
Denmark; Caravaca, San Sebastian, and Zumaya, Spain; Bari, Naples, Messina, Sicily, Italy, Tunis, Tunisia; Woodside Creek, New Zealand; Falls County, Texas; Haiti; two Deep Sea Drilling Project cores from the North-Central Pacific; and in lakes from the South Atlantic as well as in nonmarine sediments in the coal bearing strata in the Raton Basin, New Mexico, and recently in the Bighorn Basin, Montana (Alvarez, 1982, written commun.; Wänke, 1982, oral commun.). In addition, Smit and Klaiver (1981) found millimeter-size, chondrule-like spherules of nearly pure potassium feldspar that may have cooled rapidly from the melt, which also suggests possible impact origin of the layer.

The observed surface noble metal concentrations have been used to estimate the size of the impacting object or bolide (Alvarez and others, 1980). Using only the observed iridium abundances, which vary from 15 ng/g (Cribbio, Italy), 39 ng/g (Raton, New Mexico) to 520 ng/g (DSDP 46 5A), and the concentration of iridium in chondrites (0.5 × 10^-6), minimum bolide masses of 0.2 to 0.6 × 10^18 g or diameters from 5 to 16 km (density, 2.2 g/cm^3) have been calculated. The kinetic energy associated with such objects is ~10^{29} ergs (~10^4 Mton of TNT) comparable with that associated with the basin-forming impacts on planetary surfaces (O'Keefe and Ahrens, 1978).

It is highly probable that objects of the above size range have impacted the Earth. telescopic observations of the population of Earth-crossing asteroids and the post- mare record on both the Moon and the Earth suggest that the Earth could encounter a 10-km-size object on the order of every 10^5 years (Shoemaker and Wolfe, 1982). Moreover, collisions with objects of diameters of 10^3 km are inferred to have occurred on all the major planets since their formation because of the deviations of the angle of the spin axes of the terrestrial planets with respect to the ecliptic (Harriss, 1978).

The major physical evidence that is lacking at this time is the identification of the impact crater site. The diameter of the crater would be in the range of 60 to 100 km. Craters of this size are observed on the Earth; however, they are not contemporary with the C/T boundary. If the impact occurred in the ocean, the crater would be more difficult or impossible to locate since one-half of the ocean floor has been subducted in the past 65 Ma.

Because the layer is at the C/T boundary, it has been argued that the impact event is a source of C/T extinctions of biota. The most compelling evidence for the correlation of the impact event with extinctions is contained in the marine sediment record of the phytoplankton (Alvarez and others, 1980; Smit and Hertogen, 1980; Hsu, 1980). The correlation with the extinction of other biota is more controversial. According to the compilation of Emiliani and others (1982), the extinctions of species in the classes of Late Cretaceous genera were selective and limited to floating marine phytoplankton and zooplankton, swimming mollusks (ammonites and belemnites), swimming dinosaurs (ichthyosaurs and plesiosaurs), corals (shallow water), clams, and flying and walking dinosaurs, which include all land animals that had mass greater than 25 kg.

Alvarez and others (1980), Smit and Hertogen (1980), Hsu (1980), and others have proposed that physical and possible chemical effects of the impact of the bolide would directly induce massive extinction of species. These extinc- tion mechanisms include short-term heating of the atmos- phere from impact energy (Emiliani and others, 1982), shielding of the sun by lofted fine impact ejecta (Alvarez and others, 1980) and a subsequent sharp cooling of the oceans and atmospheres (Toon and others, 1982), and a subsequent decline in photosynthesis rates (Gerst and Zar- decki, 1982), shock production of nitrogen oxide (NO) (Turco and others, 1981; Lewis and others, 1982), and loft- ing of several times the normal stratospheric complement of water into the stratosphere. The latter mechanism would give rise to either a terrestrial greenhouse or possible enhanced transmission of ultraviolet light due to decline in stratospheric O_3 levels (Turco and others, 1981).

We have examined the mechanisms of asteroidal, cometary, and meteoroid swarm impact on the Earth in order to address whether or not the impact process is comparable with the physical evidence and if so, what extinction mech- anisms are implied. Specifically, we have addressed the questions: (1) Is the extreme enrichment of projectile mate- rial (up to 0.2%) in the C/T boundary clay relative to the ordinary impact brecias and melts on the Earth and the Moon consistent with the impact process? What is the ef- fect of an ocean impact on the amount of the resulting projectile material in the ejecta? (2) Does the mass of the material in the C/T boundary clay represent impact ejecta lofted to stratospheric heights? (3) How is the kinetic energy of a large Earth impactor partitioned? How much energy is converted to heat in the solid Earth; how much is available to heat the ocean and the atmosphere. What mechanisms act to transfer ejecta energy to the atmos- phere? (4) What is the temporal sequence of impact phe- nomena that could give rise to the diverse set of extinctions that have been observed?

CALCULATIONS

In order to examine the cratering flows resulting from silicate and water-bearing projectiles impacting a silicate planetary surface (for example, Earth) and its atmosphere and ocean we have employed the Eulerian finite-difference calculation algorithms to solve the conservation equa- tions of mass, momentum, and energy (Dienes and Walsh, 1970; Hageman and Walsh, 1970) in axisymmetric geome- try. The thermodynamic equation of state used to describe the various materials has the form proposed by Tillotson
where \( p \) is pressure, \( E \) is internal energy density, \( v_0 \) and \( v \) are an initial and variable specific volume, \( \eta = v_0/v \), \( \mu = n - 1 \) and a \( 0.5 \) is the polytropic constant minus 1, at high temperature. The constant, \( b \), is defined such that (\( a + b \)) is the bulk modulus. The constants, \( b \), \( B \), and \( E_0 \) are obtained by fitting Hugoniot data up to \( 1 \) \( \text{Mbar} \) and Thomas-Fermi calculations in the \( 1 \) \( \text{Mbar} \) region (O’Keefe and Ahrens, 1975). The phase transitions that occur at \( 0.15 \) \( \text{Mbar} \) in silicates, or in ice below \( 0.2 \) \( \text{Mbar} \), are taken into account by defining high pressure phase regime such that the Rankine-Hugoniot energy is given by

\[
P \propto \frac{p(E_0)}{1 - \alpha} \cdot \frac{\gamma}{\gamma - 1} \cdot \left( \frac{E_0}{1 - \alpha} \right)^{\frac{1}{\gamma - 1}}
\]

where \( \gamma \) is the specific volume of the starting material, or low pressure phase, (1 \( \text{pp} \)) and \( E_0 \) is the increase in internal energy going from the 1 \( \text{pp} \) to the high pressure phase at STP (O’Keefe and Ahrens, 1977). The constants in the equation of state are listed in Table 1.

At high temperatures and low pressures, where \( v/v_0 \) \( \approx 1 \) for \( E > E_0 \), where \( E_0 \) is the energy at standard pressure for complete vaporization, the equation of state is given by form which approaches that of the ideal gas (Hageman and Walsh, 1970; Ahrens and O’Keefe, 1977).

\[
P \propto \frac{E}{(E_0)^{\frac{1}{\gamma - 1}}} \cdot \frac{1}{\gamma - 1} \cdot \left( \frac{E_0}{E} \right)^{\frac{1}{\gamma - 1}}
\]

The geometry assumed for the present calculations was that of a spherical impactor interacting with one of more constant density layers. These geometries were (1) a single layer of either silicate, water, or air and (2) double layers of water and silicate or air and silicate. Because the C/T bowl was at least several kilometers in diameter and because of the low strength expected for such objects and the Earth on this size scale (Braze, 1988; Passey and Melosh, 1980), we assume both the impactor and the silicate Earth to have zero strength.

**PENETRATION OF THE ATMOSPHERE AND OCEANS**

For bolides having diameters on the order of 10 km, which is comparable to the scale size of the atmosphere (7.1 km) and the mean ocean depth (3.6 km), the atmosphere and oceans are not expected to significantly impede the penetration of the bolide. To quantify this we have calculated the flow resulting from penetration of asteroids, comets, and object swarms into the atmosphere and oceans. Although the interaction of the atmosphere with very porous objects has often been discussed, for example, Liu (1978) and Petrov and Studov (1976), no observational evidence exists for very porous objects in the solar system.

To examine the distance required to penetrate the penetration of various objects, we examined calculations of the impact mechanisms of these objects into a semi-infinite constant density atmosphere. Two atmospheric densities were assumed: 0.0012 and 0.00268 \( \text{g/cm}^3 \). The first represents the density at the base of the Earth’s atmosphere. Calculations of atmospheric penetration provide the basis for scaling to other values of the Earth’s and other planetary atmospheric densities. An example of these calculations is shown in Figure 1 for the impact of a low density (0.01 \( \text{g/cm}^3 \)) swarm of objects into a constant density atmosphere. The impactor punched a large hole in the atmosphere and for large objects (> 10 km in diameter) the hole would remain during the initial contact of the impactor with the planetary surface. We have calculated the penetration depth (normalized by projectile diameter) required for various bolides to lose their initial kinetic energy to the atmosphere (no stopped). In the case of porous projectiles (for example, density, 0.1 and 0.01 \( \text{g/cm}^3 \)), we are probably overestimating the stopping effectiveness of the atmosphere, since we do not allow the air to penetrate the swarm in our calculations. The effect of atmospheric interaction within a swarm has been briefly addressed by Passey and Melosh (1980). In Figure 2, we summarize the results of such a series of calculations for solid and porous water impactors interacting with a hypothetical 0.001 \( \text{g/cm}^3 \) atmosphere at 72 km/s. In this case, 0.01, 0.1, and 1 \( \text{g/cm}^3 \), 72 km/s projectiles will be arrested in 8.36, and 78 projects, respectively, for projectile diameters. The effect of atmospheric density on stopping distance is shown in Figure 3. Eulerian code calculations were made for the penetration of a 0.01 \( \text{g/cm}^3 \) bolide in atmospheres having constant densities of 0.01, 0.00268 \( \text{g/cm}^3 \). The penetration distances are very sim-
Figure 1. Flow field particle velocity (A), pressure (B), and density (C) at non-dimensional time $t = 2.3$ for impact of an initially spherical 0.1 g/cm$^3$ meteoroid swarming projectile into 0.003 g/cm$^3$ atmosphere at 72 km/s. Contours in B and C have units of (kbar) and (g/cm$^3$), respectively. Here $r$ is impact velocity times actual time divided by projectile diameter.

Figure 2. Log$_{10}$ fraction of bolide energy delivered to the atmosphere versus normalized penetration depth for 1.0, 0.1, and 0.01 g/cm$^3$ solid and porous water impacts on a uniform 0.003 g/cm$^3$ atmosphere at 72 km/s. Points are obtained using young's finite difference calculations, whereas arrows (SSF) indicate calculation carried out using supersonic flow approximation.

Figure 3. Log$_{10}$ (fraction of bolide energy delivered to the atmosphere) versus normalized penetration depth for impact of a 25 km/s, 0.01 g/cm$^3$ bolide into a 0.00123 and 0.00266 g/cm$^3$ uniform atmosphere. SSF has the same meaning as in Figure 2.
Figure 4. Log\(p_t\) (fraction of bolide energy delivered to atmosphere or ocean) versus normalized penetration depth for 0.01 g/cm\(^3\); 25 and 72 km/s projectile impacting air and 2.9 g/cm\(^3\) silicate bolide impacting water at 15 km/s. The distribution of energy in all three cases is very similar and for the air impact, insensitivity to initial velocity is demonstrated.

The equation (4) from an initial velocity \(V_0\) at \(t = 0\) to velocity, \(V\), at time, \(t\), yields
\[
V = V_t + \left(t - t_0\right)\frac{V_0 - V_t}{t_1 - t_0},
\]
where
\[
R_t = \frac{V_0}{\gamma + 1} M_0 (\gamma C_P/\gamma - 1)
\]
and
\[
\frac{V_t}{V_0} = R_{t1}/R_{t2}.
\]
Thus, the relative energy delivered to the atmosphere or ocean is given by
\[
\left(V_2^2 - V_1^2\right) = \left(\frac{V_0}{(\gamma + 1)} \frac{M_0 (\gamma C_P/\gamma - 1)}{R_{t1}}\right) R_t^{\gamma + 1}/(M_2^2).
\]

where \(p_t\) is the ambient pressure and \(\gamma\) is the polytropic exponent. Writing the stagnation pressure in terms of the drag coefficient as
\[
p_t = C_{dp} \rho V^2;
\]
the ambient pressure as
\[
p_t = p T \gamma p_t;
\]
and the Mach number (\(RT\) has the usual meaning) as
\[
M_t = V/(\gamma RT)^{\gamma + 1}/\gamma - 1,
\]
and substituting in equation (9) yields for the supersonic drag coefficient
\[
C_d = \left(\frac{\gamma + 1}{\gamma - 1}\right)^{\frac{\gamma}{\gamma + 1}} 2^\gamma.
\]

For an assumed polytropic exponent of \(\gamma = 1.4\), the supersonic drag coefficient is \(C_d = 0.92\). The values of fractional energy delivered to the atmosphere calculated from equations (6), (8), and (13) are indicated by the dashed curves labeled SSF in Figures 2 to 5. The effective polytropic exponent for water, \(\gamma = 1 + a = 1.7\) from Table 1, yields a good match to the finite difference calculations.

From these calculations, the maximum diameter object that would be stopped in the atmosphere or ocean can be determined. For example, a 1.0 g/cm\(^3\) density bolide impacting the atmosphere at 72 km/s would have a maximum diameter of 0.17 km (to just be stopped by the atmosphere), whereas a swarm of objects having an effective density of 0.01 g/cm\(^3\) would have an effective diameter of 1.5 km. In the case of an ocean impact at 25 km/s, a 15 bolide diameters (Fig. 5) are required to essentially stop an asteroid (2.93 g/cm\(^3\)). The depth of penetration of various density bolides as a function of density ratio is shown in Figure 6. The above results imply that a 10-km-size object would readily penetrate the atmosphere and ocean. The depth of penetration of various density bolides as a function of density ratio is shown in Figure 6.

The relative energy loss during the initial penetration of atmosphere and oceans by large bolides is also not significant. The fraction of energy transferred to the atmosphere as a function of penetration depth for various density projectiles is shown in Figure 2. For projectiles having diameters on the order of the scale height of the atmosphere, the energy transferred by shock heating is less than 1% for asteroids.

The amount of ablation compared to the mass of bolide is also small. The significance of ablation can be assessed by determining the relative time it would take to
penetrate the atmosphere as compared to the time it would take to ablate the bolide. The ratio of the atmospheric penetration time $t_p$ to the characteristic ablation time $t_a$ is given by

$$
\frac{t_p}{t_a} = \left( \frac{\rho_p}{\rho_a} \frac{C_p}{C_a} \frac{v^2}{g} \right)^{1/4} \frac{\sigma}{\rho_a \sigma_0.001 \cdot 10} \frac{\sigma}{\rho_a \sigma_0.001 \cdot 10}
$$

where $\rho_p$ ($\sim 2.5 g/cm^3$), $\rho_a$ ($\sim 0.001 g/cm^3$), $h_a$ (7 km), and $d$ (10 km) are the bolide and atmospheric densities, atmospheric scale height, and bolide diameter. Here $C_p$ ($\sim 0.01$) is the radiative ablation transfer coefficient and $Q$ ($\sim 4 \times 10^{10}$ ergs) is the energy of ablation (Liu, 1978). For an impact velocity of $V = 30 km/s$, $t_p/t_a \approx 0.003$. The ablation time $t_a$ is generally $10^3$ to $10^5$ times longer than the penetration time, implying insignificant mass loss and energy transfer due to penetration of the atmosphere.

The fraction of energy transferred to the ocean as a function of penetration depth is shown in Figure 5. This fraction is less than 10%, for a 3-km-deep asteroid impacting a 3-km-deep ocean.

**CRATERING OF THE EARTH—IMPACT MECHANICS**

The primary transfer of momentum and energy from the bolide occurs upon the impact with the Earth's surface. Examples of flow fields due to asteroidal and swarm impact at various velocities are shown in Figures 7 and 8. During the initial penetration, the bolide lines the wall of the growing crater cavity. With increasing impact velocity, the early-time ejecta flow field evolves from a conical to a hemispherical distribution because of an increasing amount of vaporization. The pressures produced during this time are sufficient to melt $10 (V = 15 km/s)$ to $90 (V = 45 km/s)$ times the bolide mass for asteroidal impact ($\rho = 2.93$ g/cm$^3$). Scaling relationships accounting for the amount of melt and vaporization for various density and velocity impacts (Fig. 9) demonstrate that much less melting and vaporization occurs upon Earth impact of either a comet or asteroid.

![Figure 5. Log$_{10}$ (energy delivered to ocean normalized to initial projectile energy) for 15 km/s silicate (2.9 g/cm$^3$) object impacting ocean versus normalized penetration depth. SSF has the same meaning as in Figure 2.](image)

![Figure 6. Normalized penetration depths versus log$_{10}$ (projectile into atmosphere, or ocean, stopping medium density ratio).](image)

![Figure 7. Particle velocity flow field from a silicate projectile impacting a strengthless silicone planetary surface at 15 and 45 km/s. Flow fields at (a) $v = 8$, (b) $v = 31$, where $r$ is normalized time.](image)
meteoroid swarm than a silicate or iron meteorite. The amount of melting or vaporization is proportional to the kinetic energy of impact at high velocities (>7.5 and 30 km/s), respectively, for impact of an iron meteorite.

The amount of mass ejected from the crater with at least a given ejection velocity is shown in Figure 10. These results can be used to compute the amount of mass that would be ejected in the absence of the atmosphere or ocean to at least the height indicated. At a given height, this is the maximum amount of ejecta that could be transported ballistically. However, this is not necessarily the maximum amount of ejecta lofted to a given altitude if other mechanisms such as buoyancy-driven flow effects (Jones and Kodis, 1982) are operative. From these results it can be concluded that in the absence of an atmosphere or ocean, 20 to several hundred times the bolide mass could be ejected to a height of at least 10 km, depending upon the impact scenario.

The initial high-velocity ejecta has an extraordinarily high concentration of extraterrestrial component (ETC).
SHOCK AND EJECTA MECHANICS

The energy partitioning resulting from a 30 km/s asteroidal impact with the Earth versus time is shown in Figure...
The initial kinetic energy of the bolide is transferred within a dimensionless time, \( \tau \), of less than 2 into kinetic and internal energy of the target via the initial shock interaction. The kinetic energy in the target material is subsequently partially transferred into target internal energy whose budget eventually contains some of the projectile energy. The effect of bolide density on energy partitioning is shown in Figure 14. While the equations of state differ for the asteroidal and cometary impactors, the primary effect on energy transfer to the Earth surface is due to its density difference. Most of the energy is transferred to the planetary material in the case of an asteroidal impact (~85%) and less energy (~40% to 50%) in the case of effective low density (0.1 to 0.01 g/cm\(^3\)) meteoroid swarm impacts. However, in any case, the flow field (Fig. 19) results in a significant fraction of the energy residing in the crater ejecta. This is also the case for an impact in the ocean for either an effectively porous meteorite swarm or a solid projectile (Figs. 15, 16, 17). Notably from 13% to 15% of the projectile energy resides in the water ejecta. The amount of energy ejected from the crater with at least a given velocity and the associated minimum ejection height in the absence of an atmosphere or ocean is shown in Figure 18. Referring to Figure 18, approximately 50% of the energy resides in the ejecta having ejection velocities greater than 10^4 cm/s for asteroidal impacts (V = 30 km/s). To further understand the partitioning of energy in the ejecta, we have compared the partitioning between the melt and vapor ejecta (Fig. 19). Most of the energy is contained in internal energy of the vapor and melt ejecta followed by that contained in the solid ejecta. The relative amount of kinetic energy in each of the solid, melt, and vapor energy budgets is much less than the internal energy for these phases.

From the above calculations we can conclude that the impact of large objects on the Earth can produce flow fields that eject up to several hundred times the object mass into the atmosphere with velocities greater than 10^4 cm/s and that the ejecta contains ~50% of the impact energy. However, this is not sufficient to establish whether or not the mass ejected from the crater can be distributed worldwide and that the energy contained in the ejecta can be transferred to the atmosphere. Both of these issues depend critically upon the particle size distribution of the solid, melt, and vapor ejecta and interaction with the atmosphere.

The particle size distribution of the solid ejecta is primarily dependent upon the details of the cratering flow field and the (unknown) strength properties of the rocks at the impact site. The comminution processes of the site material under hypervelocity impact conditions has not been theoretically formulated, and we take recourse to laboratory impact studies (Gault and others, 1965) and nuclear and chemical explosion field data (Schaubach, 1975). The particle size distributions for hypervelocity impact, nuclear, and explosion events, single and multiple fractioning processes have been found to have a cumulative mass distribution of the following form:

\[
M_e = C d^a
\]

where \( C \) and \( a \) are constants and \( d \) is the ejecta particle diameter. In the case of hypervelocity impact events, \( a \) ranges from 0.4 to 0.6 (Gault and others, 1965). This range of values for \( a \) is similar to that found for single-blown
Figure 15. Particle velocity flow field from 10-km diameter, 30 km/s silicate projectile impacting 5-km-deep ocean overlying a silicate Earth.

Figure 16. Fraction of impact energy partitioned into kinetic and internal energy of projectile and target (silicate and water) materials versus dimensionless time for a 3-km diameter, 0.5 g/cm³ density meteoroid swarm interacting with a 5-km-deep ocean overlying a silicate Earth at 30 km/s.

Figure 17. Fraction of impact energy partitioned into kinetic and internal energy of projectile and target (silicate and water) materials versus dimensionless time for a 10-km-diameter silicate bolide interacting with a 5-km/s ocean overlying a silicate Earth at 30 km/s.
fracturing of rocks (Hartmann, 1969). Shown in Figure 20 is the cumulative mass distributions from impact and nuclear events. Note that these data correspond to an $\alpha$ of 0.5. The particle size distributions measured from impact, nuclear, and chemical explosion events represents particles that have been ejected at various velocities and have been subjected to differing confinement histories. The high-speed ejecta that is lofted early in the impact event has been subjected to higher stresses and has experienced a greater degree of confinement than the low-speed ejecta. To account for this effect, we have assumed that there is a particle size distribution that is associated with ejection velocity. For a given ejection velocity, the particle size distribution is assumed to have the same form as the total distribution; however, the maximum particle size and the constant $C$ is assumed to be a function of ejection velocity ($V_e$).

$$M_e = CV_e^d$$  \hspace{1cm} (16)

The impact data of Gault and others (1963) provide approximate sizes of material that is ejected at high velocities ($>10^4$ cm/s), and the compilations of data by Seibergh (1975) supplies data for lower speed ejecta from a contained explosion. The latter yields an approximate relationship between maximum ejecta diameter and ejection velocity that is plotted in Figure 20.

In contrast to the solid ejecta, the melt ejecta particle size distribution is dependent upon the atmospheric (and possibly the ocean water vapor) flow field and surface tension values and not only on the impact site stress values. The general melt droplet distribution is difficult to determine theoretically; however, bounds on the minimum maximum particle size can be obtained. The minimum droplet size can be determined from the requirement that the kinetic energy of the liquid drop must be greater than the surface energy that is created. The minimum droplet size ($d_m$) as a function of ejection velocity is given by (Scarfari and Buxton, 1974)

$$d_m^2 = \frac{8 \sigma}{\rho_f V_e^2}$$  \hspace{1cm} (17)

where $\sigma$ is the surface tension (300 dynes/cm) and $\rho_f$ is the droplet density. For ejection velocities from $10^3$ to $10^5$
shown in Figure 21. For differential velocities from 10^3 to 10^5 cm/s, the maximum droplet size varies from 1 to 10^6 cm.

The particle size distribution of the condensed vapor in an expanding vapor cloud depends upon the amount of supercooling and the number of condensation centers, kinetics of droplet growth, and the hydrodynamics of expansion of the vapor. Raizer (1960) developed a theory for the condensation of a vapor cloud expanding into a vacuum, which can be applied to large-scale impacts in the atmosphere for the cases where the mass of the vapor is much larger than the mass of the atmosphere encompassed. Raizer (1960) carried out numerical calculations for a shocked vaporized iron sphere expanding into a vacuum and developed a scaling relation between the mean radii of droplets and the thermodynamic properties of the vapor.

The scaling relationship for the mean particle size is

$$R_p (cm) = \frac{\pi}{6} \left( \frac{\rho_s}{\rho_v} \right) \left( \frac{\epsilon}{\rho_v} \right)^{1/3} \left( \frac{a_0}{n_0^2} \right)^{1/3}$$

where $\rho_s$ is the initial diameter of the vapor sphere at condensed phase density, $\epsilon$ is the average initial internal energy of the vapor, and $n_0$ is the atomic number density at the instant of vapor saturation during expansion. The reference values are $\rho_v = 10^3$ cm, $\epsilon = 1.24 \times 10^{12}$ erg/g and $n_0 = 7.15 \times 10^{14}$ atoms/cm^3.

The particle number density at saturation (Raizer, 1960, Table 1) can be fitted to the following relationship:

$$n_0 (cm^{-3}) = 10^{15} \epsilon^{1/4}$$

Combining equations (19) and (20) yields

$$R_p (cm) = \frac{\pi}{6} \left( \frac{\rho_s}{\rho_v} \right) \left( \frac{\epsilon}{\rho_v} \right)^{1/3} \left( \frac{a_0}{n_0^2} \right)^{1/3} \epsilon^{1/4}$$

(21)

Now in order to account for the fact that the diameter of the vapor sphere varies with impact velocity, we use the scaling relationship from O'Keefe and Ahrens (1977) for the amount of vapor generated as a function of impact velocity, $V$. The relationship can be cast in the following form for impact on identical materials:

$$a_p = a_p + \frac{d_p}{\rho_p} \frac{\rho_v}{\rho_p} (V/V_0)^{1/4}$$

(22)

where $d_p$ is the diameter of the impacting projectile. Substituting equation (21) into (22) yields

$$R_p (cm) = \frac{\pi}{6} \left( \frac{\rho_s}{\rho_v} \right) \left( \frac{\epsilon}{\rho_v} \right)^{1/3} \left( \frac{a_0}{n_0^2} \right)^{1/3} \epsilon^{1/4} \left( \frac{V}{V_0} \right)^{1/4}$$

(23)

The reference velocity, $V_0$, is obtained from the condition for impacts on like materials, $V_0 = \sqrt{\epsilon}$, which gives a value of $2.23 \times 10^5$ cm/s.

Now substituting for the value $d_p = 10^5$ cm, and $\rho_v = 14.6 \mu$g/cm, where $\mu$g is the total energy of vaporization, and assuming the proportionality constant for the relative amount of vaporization for silicates from O'Keefe and Ahrens (1977) is similar for iron, we get the following rela-
tionship for the mean condensed droplet radius as a function of projectile diameter and impact velocity:

$$R_m = 1.15 \times 10^{-6} d_p \sqrt{v} \sqrt{a/(1+y^{10^{4}})}$$  \hspace{1cm} (24)$$

where \(d_p\) and \(v\) have units of centimeters and centimeters per second, respectively. This result is plotted in Figure 22 as a function of bolide diameter for various impactor velocities. The mean particle size increases linearly with bolide size and slowly with impact velocity. The strongest variation is with variations in average internal energy. In the case of impact into nearly semi-infinite targets such as the Earth, the average internal energy of the vapor does not vary significantly with impact velocity, since the mass vaporized varies as the square of the impact velocity and thus the average internal energy varies slowly with impact velocity. The average internal energy was assumed to be 14.6 times the energy of vaporization, which is the internal energy behind the shock for a 25-km/s impact. While the average internal energy of the vapor does not vary significantly with velocity, the deviation from the mean would be expected to be greater at higher velocities. This implies that the deviation from the mean particle size would increase with impact velocity. In addition, these droplets are also subject to stability requirements on velocity fluctuations (eq. 18), which may reduce their size. Referring to Figure 22, in the case of large-scale impacts (\(d_p \sim 10\) km), the mean condensed vapor droplet radius is less than 1 mm.

The relative amount of mass condensed was also computed by Raizer over a range of conditions of significance for impact and was found to be insensitive to impact conditions and was equal to 0.44.

In order to obtain worldwide distribution of ejecta, the residence time in the atmosphere has to be greater than several weeks. This restricts those particles in the C/T layer to the range of 1 \(\mu\)m or less (Toon and others, 1982).

The vapor, melt, and solid are all sources of particles in this size range. However, determining the amount of mass in that range is difficult, since it entails knowledge of the distribution functions in regimes that are not well known. The ejected mass of vapor ranges from 1 to 15 times the mass of the bolide. Although Figure 23 indicates that the amount of vapor ejected into the stratosphere is a factor of \(\sim 20\) less than the amount of melt, the mass of very fine condensed particles (< 1 \(\mu\)m) may be comparable to that resulting from the impact melt ejecta. The ejected mass of melt (\(V_m > 0.5\) km/s) is \(\sim 20\) times the mass of the bolide for a 30-km/s asteroid impact (Fig. 23). For all-droplet relative
velocity fluctuations of $10^6 \text{ cm/s}$, the maximum stable drop-
let size is $10^7 \text{ cm}$. We estimate that $\approx 10^2$ times the mass
of the bolide of ejected melt is in the $1 \mu\text{m}$ or less size range.
The ejected mass of solid ($Y_e > 10^6 \text{ cm}^2$) for a $30-\text{km}/\text{s}$
impact is estimated from Figure 10 to be $\approx 10^2$ bolide
masses. Referring to Figure 20, approximately $10^3$ of mass
ejected will have a diameter less than $1 \mu\text{m}$; thus, the ejected
amount of solid ejecta for a $30-\text{km}/\text{s}$ impact would be $\approx 10^2$
times the bolide mass. From the above calculations we
conclude that for a $30-\text{km}/\text{s}$ asteroidal impact $\approx 0.1$ times the
bolide mass of particles less than $1 \mu\text{m}$ can be produced by
impact. The mechanism for distributing these particles is
discussed in the following section on atmospheric effects.

**INTERACTION OF EJECTA WITH ATMOSPHERE**

As discussed above, the atmosphere has little effect on the
bolide during its initial penetration of atmosphere or
during the cratering flow in the Earth. However, the at-
mosphere has a dominating effect on the fine ejecta parti-
cles. To examine this effect, we have computed the $5-\text{km}/\text{s}$
impact of a $10-\text{km}$-diameter bolide on a uniform air layer
(scale height $\approx 10 \text{ km}$, $\rho_0 = 1.2 \times 10^3 \text{ g/cm}^3$) overlaying a
solid Earth ($\rho = 2.93 \text{ g/cm}^3$). Although only calculations
for a $5-\text{km}/\text{s}$ impact are now available, the phenomenology
was found to be similar to the extent investigated at $30$
$\text{km}/\text{s}$. The model for the atmosphere was a constant density
$10-\text{km}$ layer of air, having the same areal mass as the at-
mosphere. This model has a characteristic height that is the
same as the exponential atmosphere. The flow induced as a
result of impact onto a finite scale height atmosphere is
largely an upward expansion of the shocked air into regions
of lower density.

As depicted in Figure 1, the penetration of the bolide through
the uniform air layer produced a hole in the at-
mosphere approximately equal to the diameter of the bo-
lide ($10 \text{ km}$). The interaction of the bow shock wave in the
atmosphere in front of the bolide with the Earth resulted
in a series of shock reflections and Mach shock waves hav-
ing very high pressures ($\approx 10^6 \text{ dynes/cm}^2$) and internal en-
ergies ($\approx 10^{11} \text{ ergs/g}$). The penetration of bolide into the
Earth produced the common bowl-shaped-crater flow field with
ejecta emanating from the lip of the crater (Fig. 24).

During the penetration of the bolide into the Earth, the
hole in the atmosphere closed over the projectile. This oc-
curs on a relatively short time scale because the gas on the
periphery of the hole is highly shocked and has a high
maximum velocity of expansion into the vacuum of the
atmospheric hole. The gas on the periphery of the hole
flows radially inward filling the hole, and upward propell-
ing the air to high altitudes.

The ejecta emanating from the lip of the growing
crater intercepted the upward-moving atmosphere with a
velocity comparable to the air velocity ($V \approx 1 \text{ km/s}$). The

![Figure 24. Particle velocity flow field from a silicate projectile impacting a uniform $0.001 \text{ g/cm}^3$ atmosphere overtaking a silicate earth at $5 \text{ km/s}$ at $r = 2.8$ when projectile has delivered $\approx 90\%$ of its energy to the target(s).](image-url)

Fine ejecta particles ($< 1 \mu\text{m}$) are entrained in the air flow
and lofted to altitudes greater than $10 \text{ km}$. The very fine
particles ($< 1 \mu\text{m}$) would have long residence times and
could be distributed worldwide.

This upward-directed flow field is responsible for dis-
spersing shock melted particles and would also be responsi-
ble for dispersing the condensed vapor particles for the
higher velocity impact situations. Small spheres have been
found in the C/T boundary layer by Smit and Hertogen
(1981) and near the Eocene-Oligocene boundary by Gana-
pathy (1982). The latter author has correlated these with
tektites in the North American strewn field. Because of this
evidence, we conclude that the entrainment of melted or
condensed particles by an upward-directed flow field is the
source of tektites. This model for tektites provides (1) a
source of small and large melon particles: (2) a method for
ejection of molten particles out of the atmosphere and,
because of the limited extent of the upward flow area,
limits their strewn fields; and (3) a mechanism for cooling
droplets before atmospheric reentry. Moreover, if the im-
 pact is oblique, which of course is the usual case, then the
upward-flowing air is channelled backwards along the in-
tial bolide trajectory. This channelled flow of air would
distribute tektites and microtektites, preferentially in the
east-west directions, because the preferential influx of me-
teoroids onto the Earth from the plane of the ecliptic. The
asymmetry has been observed (Glass and others, 1973).
The primary transfer of energy to the atmosphere occurs through the interaction with the ejecta. Only a small fraction of the energy is transferred to the atmosphere by the primary shock wave. Shown in Figure 25 is the time history of the energy partitioning between the bolide, Earth, and atmosphere as a function of time. Note that less than 5% of impact energy is directly transferred to atmosphere. The transfer of internal and kinetic energy of the ejecta to the atmosphere depends upon the particle size distribution. This fine-scale, discrete particle-atmospheric energy transfer model has been modified in continuum calculations and has to be accounted for by separate analysis. The transfer of both the internal and kinetic energy of the ejecta to the atmosphere depends on the particle size distributions of the ejecta species.

The kinetic energy of particles less than $10^5$ cm in diameter is nearly totally transferred into thermal energy in the atmosphere by work done by drag forces. Particles having diameters greater than $10^5$ cm have essentially vacuum-like trajectories and transfer little of their kinetic energy to the atmosphere. This occurs not only because of the increased mass of the particles but also because the ejection velocities decrease with particle size.

The internal energy of particles can be transferred to the atmosphere when the characteristic cooling time is less than the residence time (Fig. 26). In the figure, we indicate the characteristic times for various phenomena as a function of ejecta diameter. The characteristic cooling time increases as the square of the particle size. This can be estimated from the time it takes for heat to penetrate the radius; for example, $t = R^2/\kappa$, where $\kappa$ is the silicate thermal diffusivity ($0.01$ cm$^2$/s). In the case of solid ejecta, the ballistic time-of-flight can be calculated for the maximum ejecta particle size, which varies with ejection velocity for the distribution in Figure 20. This characteristic time decreases with increasing particle size because the ejection velocity decreases. Shown also is the lofting and ballistic regimes. Small ejecta particles get entrained in the atmospheric flow field and have residence times much greater than ballistic times; the range of these particle sizes defines the lofting regime of Figure 26. The range of particle sizes that have residence times that are independent of the atmospheric flow field is designated as the ballistic regime. From these results, we conclude that the internal energy of particles can be transferred to the atmosphere when their diameters are less than $0.1$ cm.

Having determined the particle sizes that can transfer their energy to the atmosphere, the amount of energy can be calculated from the particle size distributions. As an example, consider a 30-km/s impact of an asteroid on the Earth. The fraction of internal and kinetic energy contained in the vapor, melt, and solid ejecta having ejection velocities greater than $0.5 \times 10^5$ cm/s is given in Table 2. The fraction of total impact energy in the kinetic energy of the solid ejecta ($V > 0.5 \times 10^5$ cm/s) is 5%. This fraction of the ejecta would have particle sizes less than 10 cm in diameter and thus would essentially transfer all the kinetic energy into thermal energy of the atmosphere.

The internal energy contained in the solid ejecta would not be completely transferred to the atmosphere because the relative amount of mass in particles less than 0.1 cm is small ($< 10\%$). The amount of internal energy in the solid ejecta transferred to the atmosphere is $< 0.1\%$. The kinetic energy contained in the ejected melt would be transferred to the atmosphere, since the mean particle size would be much less than 10 cm. The same argument holds for the
TABLE 2. FRACTION OF TOTAL ENERGY THAT EJECTA TRANSFER TO ATMOSPHERE FOR A 30 KM/S ASTEROIDAL IMPACT (AFTER O’KEEFE AND AHRENS, 1982b)

<table>
<thead>
<tr>
<th>Ejecta Type</th>
<th>Fraction of total energy in the 0.5 X 10^5 cm/s ejecta</th>
<th>Efficiency of transferring energy to atmosphere</th>
<th>Fraction transferred to atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vapour-internal</td>
<td>0.12</td>
<td>1.0</td>
<td>0.12</td>
</tr>
<tr>
<td>Melt-internal</td>
<td>0.15</td>
<td>1.0</td>
<td>0.15</td>
</tr>
<tr>
<td>Solid-internal</td>
<td>0.03</td>
<td>0.1</td>
<td>0.00</td>
</tr>
<tr>
<td>Vapour-kinet</td>
<td>0.01</td>
<td>1.0</td>
<td>0.01</td>
</tr>
<tr>
<td>Melt-kinet</td>
<td>0.03</td>
<td>1.0</td>
<td>0.03</td>
</tr>
<tr>
<td>Solid-kinet</td>
<td>0.05</td>
<td>1.0</td>
<td>0.05</td>
</tr>
<tr>
<td>Air shock</td>
<td></td>
<td></td>
<td>0.03</td>
</tr>
<tr>
<td><strong>Totals</strong></td>
<td></td>
<td></td>
<td>0.39</td>
</tr>
</tbody>
</table>

Conclusions

We show that during the initial penetration of the atmosphere and the ocean (if any) relatively little mass or energy would be transferred from the C/T bolide. An incoming bolide would not be dispersed by the atmospheric drag forces (Passey and Melosh, 1980) because the object size is comparable to the scale height. This occurs even though the intrinsic dynamic tensile strength of ice is low in the case of comets (~10 MPa; M. Lange and T. J. Ahrens, 1982, unpub.) and the dynamic strength of previously impact brecciated asteroids is also low. We do not believe it is possible to reduce the effective bolide density as proposed by Kyte and others (1986) by tidal disruption. However, we have considered the impact of low-density objects, which to some degree represent swarms of meteoroids, while recognizing that there is no telescopic evidence for their existence. Over the range of velocities (5 to 73 km/s) and impact densities considered (0.01 ~ 2.9 g cm^-3), the dominant mass and energy transfer occurs upon impact with the solid Earth. The shock pressures and internal energy densities induced in cometary bolides upon impact would thermally decompose thermally less stable compounds such as cyano-gens, which if brought to the Earth in significant quantities might be considered possible agents for poisoning biota and causing extinctions (Hoel, 1980).

Upon impact on the Earth, the resulting flow fields can inject ~10^10 to ~10^11 times the bolide mass into the stratosphere. In the case of an ocean impact, water having a mass 10 to 10^2 times the bolide would also be injected into the atmosphere. The ejecta that is propelled early in the cratering flow is enriched in volatile material and has concentrations of extraterrestrial material in the range of those measured in C/T boundary layer materials. This volatilized material is highly shocked and consists of condensed vapor, melt droplets, and finely comminuted solid particles. Due to the higher ejecta velocities for the shocked extraterrestrial material achieved upon continental versus oceanic impact, the fraction of extra terrestrial material in the non-water ejecta is lower for oceanic impact than for continent impact.

For both terrestrial and oceanic impact, the resulting ejecta couples strongly with the atmosphere. Penetration of the atmosphere by the bolide creates a hole in the atmosphere that is subsequently filled by an air flow field that is inward and upward. The upward component of the air flow field lofts the vapor, fine melted and solid ejecta to heights greater than 10 km. The millimeter- to centimeter-size droplets that are lofted by this mechanism reenter the atmosphere and may represent microejectas and tektites. In contrast, the particles less than ~1 μm, which comprise ~0.1 times the bolide mass, can be lofted to heights greater than 10 km and be distributed globally in several months. Calculations of the solar transmission reduction due to such dust injection (Grossi and Zareňački, 1982) show that 10^18 g uniformly distributed over the Earth would reduce photosynthesis by a factor of 10^7. Notably, this mass is only 10^2 times the mass of the estimated C/T bolide from iridium concentrations (Alvarez and others, 1980). An
amount of dust injected of this magnitude would probably result in sooting of the stratosphere and heating of the stratosphere (Toon and others, 1982).

In the case of a cometary impact on land or an ocean impact, a significant fraction of the ejecta (off to high altitudes would be water. Undoubtedly much of this water would be rapidly condensed and rainout; however, the consequence could be a decrease in the concentration of ozone and possible triggering of an enhanced terrestrial greenhouse. The passage of the bolide through the atmosphere (Lewis and others, 1982; Turco and others, 1983) and, as we suggest, the interaction of the ejecta with the atmosphere would produce high-temperature shock waves and, thus, NO. This might interact with the ozone layer and deplete it for times on the order of a decade (Turco and others, 1981).

The atmosphere would be heated by both the shock waves produced by the bolide and ejecta and the transfer of internal energy from the first ejecta. The fraction of bolide energy transferred to the atmosphere is ~40%. This could, if not immediately reradiated to space, result in a global average temperature increase of at least 15°C for a 10^11 erg impact. However, we believe that a localized heat pulse would be more probable and that global heating via the ejecta interaction mechanism would be difficult. Although the characteristic time scale for the radiative decay of such a temperature excursion might be only a matter of days, Emiliani and others (1982) have argued that such an average global rise of only a few degrees would be intolerable to many biota (especially larger reptiles >25 kg).

The impact of the C/T bolide gives rise to a number of phenomena and processes that can result in massive extinction of biota by creating environmental stresses on time scales varying from hours to decades (Fig. 27). For both terrestrial and oceanic impact, the ejecta-atmospheric interaction is such that a significant portion of the atmosphere and the upper 1 km of the oceans could achieve temperature rises of ~15°C and ~5°C, respectively. Both the particulate and water ejecta could then give rise to a worldwide cooling period lasting from 10^2 to 10^3 days, which would be severe on land. At sea, the resulting lack of sunlight would halt photosynthesis in phytoplankton and lead to collapse of the marine food chain. Finally, the water from impact ejecta and/or NO production from the bolide-atmosphere interaction might give rise to major perturbations of the atmosphere's thermal radiation balance leading to an enhanced terrestrial greenhouse. In addition, the H_2O and NO could depose the stratospheric and mesospheric O_3 budgets and cause extensive and life-damaging UV irradiation onto the Earth's surface.

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Figure 27. Relevant time scales for extinction of environmental stress on biota resulting from the various effects of the impact of Cretaceous-Tertiary bolide in the ocean or on land.


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