Impact Erosion of Planetary Atmospheres

JAMES C. G. WALKER

Space Physics Research Laboratory, Department of Atmospheric and Oceanic Science,
The University of Michigan, Ann Arbor, Michigan 48109

The impact of a large extraterrestrial body onto a planet deposits considerable energy in the atmosphere. If the radius of the impactor is much larger than an atmospheric scale height and its velocity much larger than the planetary escape velocity, some of the planetary atmosphere may be driven off into space. The process is analyzed theoretically in this paper. The amount of gas that escapes is equal to the amount of gas intercepted by the impacting body multiplied by a factor not very different from unity. Escape occurs only if the velocity of the impacting body exceeds the planetary escape velocity. At large impact velocities the enhancement factor, which is the factor multiplying the amount of atmosphere intercepted by the impacting body, approaches a constant value approximately equal to $10^{12}/V_c^2$, where V_e is the escape velocity (in cm/sec). The enhancement factor is independent of atmospheric mass or surface pressure. Ablation of the impacting body and the planetary surface adds to the mass of gas that must be accelerated into space if escape is to occur. As a result, impact erosion of the atmosphere does not occur from a planet with an escape velocity in excess of 10 km/sec. © 1986 Academic Press, Inc.

INTRODUCTION

Very large amounts of energy are released by the impact of an extraterrestrial body with a rocky planet. The effect of this energy on the solid phase of the planet has been extensively studied, both experimentally and theoretically. Impact craters surrounded by fields of debris and underlain by fractured and melted rock are the result. Changes in atmospheric chemistry and climate resulting from extraterrestrial impact have also been extensively discussed (Fegley *et al.*, 1986; Matsui and Abe, 1986). But an aspect of the interaction with a substantial potential impact on the origin and evolution of planetary atmospheres has as yet been the subject of little quantitative research. This is the erosion of the atmosphere that might result if sufficient energy is imparted to atmospheric gas to blow it completely out of the gravitational field of the planet (Lewis *et al.*, 1982; Cameron, 1983).

Studies of the process of impact cratering of the solid phase of a planet are not useful guides to the fate of energy and momentum imparted by an impact to the gas phase (O'Keefe and Ahrens, 1982). Perhaps the most important difference is the much larger compressibility of the gas. High impact pressures compress an atmospheric gas by large amounts and yield very high temperatures. Much less compression and much lower temperatures are achieved in the relatively incompressible solid phase of a planet. While the solid part of a planet must, so to speak, get out of the way of the impactor immediately, the first response of the atmospheric gas to an encounter with an impactor is to compress. The subsequent flow of gas out of the way of the impactor is influenced by compressive increase of density by an order of magnitude or more and also by possible loss of energy resulting from radiation at high temperature.

On the other hand, extrapolation of knowledge gained from the atmospheric effects of large nuclear explosions is also a poor guide to the effects of an impact on an atmosphere (Jones and Kodis, 1982). Perhaps the most important inadequacy of the explosion model is the very different tem-

87

poral and spatial scales. An explosion releases a large amount of energy in a small volume and in a short period of time. The result is very high initial temperatures in the fireball and a major role for radiation in the subsequent expansion of the disturbance. A large impactor, by way of contrast, releases energy over an area comparable to the cross section of the impactor and over a time comparable to its radius divided by its velocity. The duration of energy release might therefore be tens of seconds for an impactor with a radius of hundreds of kilometers traveling at a speed of some tens of kilometers per second. This relatively gradual release of energy over a large area causes mechanical interactions to be relatively more important than radiative interactions in an impact as compared with an explosion.

It does appear, on the other hand, that knowledge concerning the interaction of meteorites with a planetary atmosphere can be applied to the problem of impact erosion of planetary atmospheres. The approach used in this paper is to draw on the extensive body of theoretical and experimental information concerning meteorite impact to extrapolate to the case of a very large impactor. The goal is to provide an estimate of the amount of gas that might be driven off into space by a given impact onto a given planet. These initial estimates must of necessity be approximate and involve many extreme simplifications. But the development of these estimates should show what processes and interactions are important and thereby point the way to more precise theoretical treatment in the future.

DESCRIPTION OF THE INTERACTION

I shall consider the case of an impactor with a dimension very much larger than the scale height of the atmosphere. This impactor collides with a rocky planet like Mars or Earth with a velocity that exceeds the escape velocity, possibly by a large amount. Typical impact velocities appear to lie between the escape velocity and 40 or 50 km/sec (Opik, 1958; Hawkins, 1964; Bronshten, 1983). I need to estimate how much energy the impactor imparts to the atmospheric gas.

Because of its compressibility, the air in front of the impactor is hotter than either the solid planet or the impactor. Radiation, if it plays a role, must therefore remove energy from the air, not impart energy to it. I therefore concentrate on the mechanical work done by the impactor on the atmospheric gas. Mean thermal speeds of atmospheric molecules are obviously very much less than the escape velocity, so the impact speed is supersonic by a large amount. As the impactor travels down through the atmosphere it encounters increasing atmospheric density. The large size of the impactor implies that free molecular flow gives way to continuum flow somewhere above the atmospheric exobase. The most important interactions, of course, occur in the lowest one or two scale heights of the atmosphere where nearly all of the atmospheric mass is concentrated. The geometry of the interaction can therefore be thought of as a thin film of fluid being squeezed between two large balls.

Penetration of the impactor into the atmosphere generates an almost horizontal shock wave that travels almost vertically downward. The air has little opportunity to flow out and around the very large impactor, so it accumulates between the shock wave and the front face of the impactor. The shock wave travels slightly faster than the impactor in order to accommodate the accumulation of air, subject to the constraints of continuity. The difference in speeds is small because the air behind the shock has a much larger density than the air in front of the shock. For the same reason the layer of shock-heated air in front of the impactor has a thickness much less than an atmospheric scale height. Estimates of the density ratio and the thickness of the shock-heated layer will be presented below. Compression of the air in the shock wave imparts random thermal energy (enthalpy)

to the air. An approximately equal amount of energy is imparted to the gas in the form of directed kinetic energy. Behind the shock the air has been compressed and is flowing downward at approximately the impact speed.

The shock-heated air, which is stationary in the rest frame of the impactor, hits the ground shortly before the impactor itself does. The ground drives a shock wave back through this gas at essentially the impact speed. Some of the directed kinetic energy is converted into enthalpy in this interaction. I conclude that supersonic compression of air between the impactor and the solid surface of the planet generates very high pressures in the gas at the point where impact first occurs. These high pressures cause the air to flow radially outward horizontally from the point of impact. As the impactor is engulfed by the planet, the atmosphere is driven away from the point of impact at a speed comparable to the impact speed over a distance comparable to the radius of the more or less spherical impactor. I shall describe this interaction as a piston-driven shock wave in which the piston moves radially outward at the impact speed from the point of impact to a distance equal to the radius of the impactor (see Fig. 1).

Most of the impact energy is released in explosive form at some depth within the planet. The atmosphere receives further energy when this explosion reaches the surface, but this contribution to the atmospheric energy budget turns out to be minor because the velocity of the rebound decreases rapidly with distance from the point of impact (Melosh, 1984). As I show below, transfer of energy to the atmospheric gas falls off rapidly with decreasing speed of the solid driver, whether impactor or rebounding planetary surface.

The effect of the impact on the atmosphere is therefore to produce a region of air around the impact with a lot of thermal and kinetic energy. This energetic air expands both vertically into the vacuum of interplanetary space and horizontally, en-
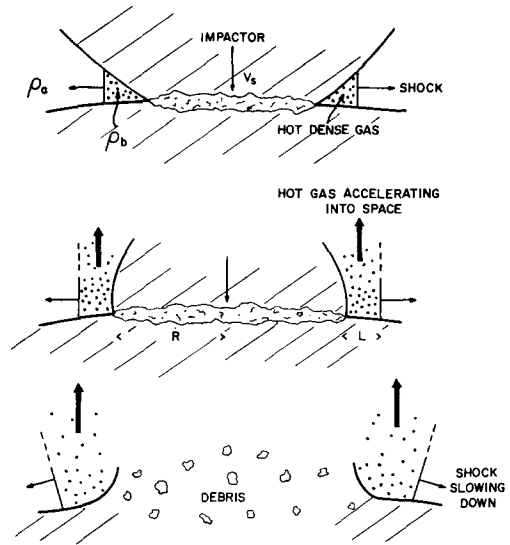

FIG. 1. Schematic illustration of the interaction between impactor, atmosphere, and planet. The impact initially drives a shock wave radially outward from the point of impact horizontally through the atmosphere. After the impactor has been engulfed by the planet the shock-heated gas accelerates off into space while the shock wave continues to move horizontally outward, engulfing ambient air and slowing down.

gulfing ambient unperturbed atmosphere. I seek to estimate how much air receives enough energy to escape from the gravitational field of the planet.

WORK DONE ON GAS

Change in the velocity of the impactor during the course of its interaction with the atmosphere, either as a result of atmospheric drag or of gravitational acceleration, can be neglected. In a reference frame fixed with respect to the impactor, the ambient gas, of density ρ_a, is flowing toward the impactor with velocity V_s. At the shock wave, a short distance in front of the impactor, there is a discontinuity in the gas properties. Pressure, density, and temperature jump to large values. The directed velocity falls to zero because the gas between the shock and the impactor is at rest with respect to the impactor. Subscript a refers to gas properties in front of the shock. Sub-

script b refers to properties between the shock and the impactor. The shock is moving slowing away from the impactor, but the difference in shock speed and impactor speed can be neglected because of the large value of the density ratio, ρ_b/ρ_a. This result will be demonstrated below after the properties of the shock-heated gas have been calculated.

For a strong shock, with shock speed very much greater than the mean thermal speed of the ambient gas molecules, the thermal energy of the ambient gas incident on the shock is very much less than its directed kinetic energy. The energy content of the incident gas is therefore $(\rho_a/2)V_s^2$ per unit volume and the energy flux into the interaction region between shock and impactor is $(\rho_a/2)V_s^3$. Behind the shock the gas is at rest with respect to the impactor. All of the energy has been converted into random thermal energy of the gas and work done on compressing the gas. This energy content per unit weight of the gas is conveniently described by the enthalpy, h_b. For the time being, I shall neglect evaporation of the impactor. Then the kinetic energy carried by the gas flowing into the shock is used to increase the enthalpy of this same gas. So the flux of air into the shock that needs to be heated is $\rho_a V_s$. Energy balance is therefore expressed by the following equation

$$(\rho_a/2)V_s^3 = \rho_a V_s h_b \qquad (1)$$

or

$$h_b = V_s^2/2. \qquad (2)$$

This is a standard result for the enthalpy of the gas behind a strong shock (Vincenti and Kruger, 1975).

This is only one component of the energy budget, however. In the rest frame of the planet the gas, which had been at rest, has now acquired kinetic energy per unit mass equal to $V_s^2/2$ because the gas behind the shock is not only hot but also moving at the same speed as the impactor. In the rest frame of the planet, therefore, the shock-heated gas has total energy per unit mass

equal to V_s^2. This energy is divided equally between enthalpy (heat and compression) and the kinetic energy of mass motion.

The pressure of the gas in the interaction region can be derived by considering the balance of forces on a thin layer of gas right at the shock. Behind the shock the gas pressure is p_b, the gas is at rest, and the force of this gas on a unit area of the shock front is p_b. The pressure of the gas in front of the shock is negligible by comparison, but the gas is flowing into the shock with velocity V_s. The momentum of the gas relative to the shock is $\rho_a V_s$ and the momentum flux into the shock is $\rho_a V_s^2$, which is the force per unit area exerted by the gas outside the shock on a thin layer of gas at the shock. When these two forces are equated an expression for the pressure in the interaction region results:

$$p_b = \rho_a V_s^2 \qquad (3)$$

(Vincenti and Kruger, 1975).

To complete the energy budget, I consider now the work done by the gas on the impactor. Per unit area and time this work is p_b multiplied by the velocity, or $\rho_a V_s^3$. This is equal to the work done by the impactor on the atmosphere, per unit area and time an energy per unit mass of V_s^2 multiplied by a mass flux of $\rho_a V_s$.

It is because the energy imparted to the atmospheric gas varies as the square of the impact velocity that the contribution of planetary material ejected from the impact crater is minor. The interaction of pieces of planetary material flying through the atmosphere is presumably much the same as the interaction of the original impactor, but the pieces of planetary material have lower speeds. These speeds have been analyzed by Melosh (1984). Melosh shows that the impact of solid bodies of equal density imparts to the material of the target planet an initial speed of $V_s/2$. Much of the energy of the interaction is released as if by an explosion originating at a depth of approximately twice the average radius of the impactor. The particle speed in the spheri-

cally propagating compressional wave falls off with distance from this center approximately inversely as the square of the distance. At the surface close to the point of impact, therefore, the particle speed is approximately $(V_s/2)(1/4)$. Reflection of the compressional wave from the free surface doubles the particle speed, so rebounding planetary material travels up through the atmosphere with a velocity of about $V_s/4$ close to the impact. This velocity varies inversely as the square of the distance from the effective center of the explosion. Because the energy imparted to the atmosphere varies as the square of the velocity of the interaction, rebounding planetary material is less effective per unit area than the original impact by a factor of 16. I shall therefore neglect the additional energy imparted to the atmosphere by rebounding planetary material.

EVAPORATION OF IMPACTOR

One of the characteristic features of the interaction of a meteor with an atmosphere is evaporation of the material of the meteor. This process will have no significant impact on the mass of the very large impactor that I am considering here, but meteor vapors can have a considerable impact on the properties of the shock-heated gas. In meteor theory (Bronshten, 1983) it is the practice to describe ablation or evaporation of the meteor in terms of the fraction, Λ, of the incident energy in the rest frame of the impactor that is absorbed by latent heat of evaporation. Thus, the rate of evaporation per unit area is

$$dm/dt = -\Lambda(\rho_a/2) V_s^3/Q \qquad (4)$$

where Q is the latent heat of evaporation of the meteor material. According to Bronshten (1983), a typical value for Q is 8×10^{10} erg/g and Λ can have values between 0.1 and 0.2 at low levels of the atmosphere where the ambient density is large. I now consider the effect of this additional vapor on the properties of the

TABLE I

Relative Contribution of Evaporation

V_s (km/sec)	5	10	20	40
Λ		E		
0.1	.156	.625	2.5	10
0.2	.312	1.25	5.0	20

gas in the interaction region between the shock and the impactor.

I consider first the conservation of matter. Per unit area, material is added to the interaction region at the rate $\rho_a V_s$ across the shock and at the rate $-dm/dt$ by evaporation from the impactor. The thickness of the interaction region, L, must increase fast enough to accommodate this addition of material. Per unit area, the amount of matter in the interaction region is $\rho_b L$, so

$$\rho_b(dL/dt) = \rho_a V_s + \Lambda(\rho_a/2)(V_s^3/Q) \qquad (5)$$

or

$$dL/dt = V_s(\rho_a/\rho_b)(1 + E) \qquad (6)$$

where

$$E = \Lambda V_s^2/(2Q). \qquad (7)$$

The parameter E defined by Eq. (7) describes the relative contribution of evaporation to the mass of the gas in the interaction region. Table I presents illustrative values of this parameter for the values of Λ and Q cited above. My earlier assertion that the shock speed is very nearly the same as the impactor speed is based on Eq. (6) and the further result, not yet shown, that $\rho_b/\rho_a \gg 1 + E$.

I consider now the impact of evaporation on the energy balance. In the rest frame of the impactor the energy flux into the interaction region is still $(\rho_a/2)V_s^3$, but this energy now has to heat the vapors evaporated from the impactor as well as the ambient air carried into the interaction region. The energy balance becomes

$$(\rho_a/2)V_s^3 = h_b\rho_a V_s(1 + E) \qquad (8)$$

or

$$h_b = (V_s^2/2)/(1 + E). \qquad (9)$$

Evaporation of the impactor significantly decreases the enthalpy per gram of the gas in the interaction region.

The balance of forces at the shock is not influenced by the presence of meteor vapors in the interaction region, so the pressure of the gas behind the shock is not affected by evaporation. Equation (3) still applies. In addition, the overall energy budget is not changed because additional kinetic energy does not have to be imparted to the meteor vapor. Evaporation of the impactor therefore yields more gas with the kinetic energy appropriate to the impact velocity and the same total amount of enthalpy residing in a larger total mass of gas. Because the impact speed exceeds the escape speed, it is clear that this shock-heated gas has enough energy to escape from the gravitational field of the planet. Before considering escape, however, I shall estimate the properties of the shock-heated gas. These properties do not seem to bear directly on the amount of gas that escapes, but they do illustrate important features of the interaction.

PROPERTIES OF SHOCK-HEATED GAS

The pressure of the shock-heated gas is given by Eq. (3) and the enthalpy per unit mass is given by Eq. (9). In principle, the temperature of the gas can be calculated from the enthalpy and then the density from the equation of state. The problem is complicated by the fact that the range of impact speeds under consideration causes partial dissociation and ionization of the gas behind the shock. Calculation of the gas properties is therefore not straightforward and a detailed and precise calculation would be quite laborious. Because the escape process does not depend directly on the temperature and density of the shock-heated gas but only on its total energy, I shall not undertake a detailed calculation. In what follows I carry out a simple illustrative cal-culation that provides an indication of the properties of the shock-heated gas.

For purposes of illustration I assume that the shock-heated gas has the atomic and molecular properties of nitrogen. The relationships that follow are derived from the textbook of Vincenti and Kruger (1975). I further assume that it is only nitrogen atoms, not molecules, that undergo ionization. Let the fractional ionization be ϕ, equal to the number of ions per unit volume divided by the total number of ions, atoms, and molecules. Then the fractional ionization is given in terms of temperature and pressure by the following equation:

$$\phi^2/(1 - \phi^2) = K(T^{5/2}/p)\exp(-\Theta_i/T) \quad (10)$$

where

$$K = (2\pi m_e/h^2)^{3/2}k^{5/2}(2g^+/g) \qquad (11)$$

where m_e is the mass of the electron, h is Planck's constant, k is Boltzmann's constant, g^+ and g are the statistical weights of the ground terms of the ions and atoms, and $\Theta_i = 169,000°K$ is the characteristic temperature for ionization of atomic nitrogen.

The fractional dissociation is designated α and is the mass fraction of the gas in atomic rather than molecular form. It is the number of atoms divided by the number of atoms plus twice the number of molecules. To a close approximation α is given in terms of density and temperature by the following equation:

$$\alpha^2/(1 - \alpha) = (\rho_d/\rho)\exp(-\Theta_d/T) \quad (12)$$

where $\rho_d = 130$ g/cm³ is the characteristic density for dissociation of nitrogen molecules and $\Theta_d = 113,000°K$ is the characteristic temperature for dissociation.

In terms of ϕ and α, then, the total number density of molecules, atoms, ions, and electrons in the shock-heated gas is $(\rho_b/m_m)(1 + \alpha)(1 + \phi)$, where m_m is the molecular weight of the undissociated gas. Obviously the mean molecular weight of the gas is

$$\overline{m} = m_m/[(1 + \alpha)(1 + \phi)]. \qquad (13)$$

TABLE II

PROPERTIES OF THE HOT GAS

V_s (km/sec)	p_b (dyn/cm^2)	h_b (erg/g)	E	$\overline{m}$ (amu)	T_b (°K)	α	ϕ	ρ_b (g/cm^3)	$(1 + E)\rho_a/\rho_b$	τ_v (sec)
15	2.78×10^9	3.62×10^{11}	2.11	24.0	15000	.665	.0139	.0534	.0719	13.8
20	4.94×10^9	4.21×10^{11}	3.75	23.5	17800	.789	.0315	.0783	.0750	12.3
30	1.11×10^{10}	4.77×10^{11}	8.44	23.7	21700	.850	.0616	.146	.0798	10.9
40	1.98×10^{10}	5.0×10^{11}	15.0	24.0	23900	.851	.0758	.238	.0829	10.2

Note. $\rho_a = 1.23 \times 10^{-3}$ g/cm^3.

The molecules are a mixture of ambient atmospheric molecules of molecular weight m_a and meteorite vapor molecules of molecular weight m_v. They enter the mixture in the proportions of $1 : E$, so

$$m_m = m_a m_v (1 + E)/(m_v + E m_a). \quad (14)$$

According to Öpik (1958) the mean molecular weight of meteor vapor is $m_v = 50$. For air I shall use $m_a = 29$.

The temperature of the shock-heated gas can be derived from an expression for the enthalpy per gram in terms of the temperature. In deriving this expression, I assume that each atom, ion, and electron contributes 5/2 kT_b to the enthalpy. Each molecule contributes 8/2 kT_b which allows for excitation of 1/2 of the possible vibrational degrees of freedom (Vincenti and Kruger, 1975), each ion pair contributes an enthalpy of $k\Theta_i$, and each atom pair contributes $k\Theta_d$. The enthalpy also includes a contribution from the latent heat of vaporization of meteor material. If this latent heat per gram is Q, then per molecule of meteorite vapor it is Qm_v. The fraction of all gas molecules that are meteorite vapor is $(E/m_v)/(1/m_a + E/m_v)$, so the latent heat contribution per average gas molecule is $QEm_a m_v/(m_v + E m_a)$. When all of these contributions to the enthalpy are combined, with due allowance for the relative concentrations of atoms, molecules, ions, and electrons, there results

$$h_b = R_m[5/2T_b\{(1 + \alpha)(1 + \phi)-(1 - \alpha)\}$$
$$+ 4T_b(1 - \alpha) + \phi(1 + \alpha)\Theta_i + \alpha\Theta_d$$
$$+ (Q/k)Em_a m_v/(m_v + E m_a)] \quad (15)$$

where $R_m = k/m_m$. The temperature of the shock-heated air, T_b, can be calculated from this expression for known enthalpy, h_b, in terms of the fractional dissociation and fractional ionization. The final equation needed to close the system is the equation of state

$$p_b = (\rho_b/\overline{m})kT_b. \quad (16)$$

In practice I solve the system of equations iteratively for a given value of impact velocity, starting with assumed values for fractional dissociation α and fractional ionization ϕ. Since p_b is related to ambient density ρ_a by Eq. (3) and h_b is given in terms of impact speed by Eq. (9), the equation of state (16) can be rearranged to yield an expression for the density ratio

$$(1 + E)(\rho_a/\rho_b) = kT_b/(2h_b\overline{m}). \quad (17)$$

Equation (15) for enthalpy, h_b, makes it fairly clear that this quantity is less than 1, indeed a lot less than 1, when there are significant amounts of dissociation and ionization. This is precisely the criterion derived above for there to be little difference between the shock speed and the speed of the impactor.

Table II presents illustrative values of the properties of the shock-heated gas for various values of the impact speed V_s and an ambient density $\rho_a = 1.23 \times 10^{-3}$ g/cm^3. In performing these calculations, I have assumed an intermediate value of 0.15 for the evaporation efficiency, Λ. The column for density ratio, $(1 + E)(\rho_a/\rho_b)$, gives not only the shock speed relative to the impactor

speed, $(1/V_s)(dL/Dt)$ in terms of Eq. (6), but also, from considerations of conservation of mass, the thickness of the shock-heated layer at the end of the impact event relative to the mean radius of the impactor, L/R. Because the gas density in the interaction region is so much larger than the ambient atmospheric density, the thickness of the layer is small compared with the radius of the impactor. Indeed, for the values presented in Table II, the thickness of the layer of shock-heated air is less than a typical atmospheric scale height of 10 km for impactors of radius less than 125 km.

The temperatures presented in Table II for the shock-heated gas are significantly lower than temperatures achieved in the fireballs of nuclear explosions (Brode, 1968). Although the amount of energy released in the impact is enormous, the rate of release is relatively slow and it involves a large volume whereas nuclear explosions start out in a very small volume. At these low temperatures the shock speed exceeds the speed of radiative expansion of the fireball. It is therefore appropriate to concentrate on the mechanical interactions of the shock rather than on radiative interactions.

As the impactor is engulfed by the planet, then, all the air that was originally in its path is compressed into a thin layer of hot, high-pressure gas mixed with a substantial amount of meteor vapor. This gas has an enthalpy per gram given by Eq. (9) and it is moving outward from the point of impact at a speed essentially equal to the impact speed. This hot compressed gas can be expected to expand outward into the vacuum of space and also to expand horizontally and engulf the ambient unperturbed atmosphere. It might also expand backward into the void left by the passage of the impactor, except this void is quite probably filled with vapor and debris emerging from the impact crater.

EXPANSION OF SHOCK-HEATED GAS

The shock-heated gas is moving horizontally outward from the point of impact at

approximately the impact speed, entraining, accelerating, and heating more of the ambient atmosphere. At the same time the hot gas is expanding vertically into the vacuum of space. Consider the vertical expansion first.

The vertical component of the equation of motion of the gas is

$$\rho_b(dw/dt) = -dp_b/dz - \rho_b g \quad (18)$$

where w is the vertical velocity of the gas, initially zero, z is the altitude, and g is the acceleration due to gravity. At the start of the expansion the pressure is given by Eq. (3), $p_b = \rho_a V_s^2$. The pressure gradient is therefore

$$dp_b/dz = V_s^2 \, d\rho_a/dz = -V_s^2 \rho_a/H \quad (19)$$

where H is the ambient atmospheric scale height. This term is very much larger than the gravitational term in Eq. (18), so the initial vertical acceleration of the gas can be expressed as

$$dw/dt = (V_s^2/H)(\rho_a/\rho_b). \quad (20)$$

The pressure gradient decreases as the gas expands upward, so the acceleration decreases also, but an indication of the rapidity of this process can be derived by neglecting this effect. I derive a characteristic time for vertical acceleration by dividing this expression for the initial acceleration into the velocity that the gas would have if all of its initial enthalpy, $(V_s^2/2)(1 + E)$, were converted into mass motion. This characteristic time is

$$\tau_v = H/[V_s(1 + E)^{1/2}](\rho_b/\rho_a). \quad (21)$$

Illustrative values of τ_v appear in Table II. They were calculated for an ambient atmospheric scale height of 8.5 km. The characteristic times for vertical acceleration are about 10 sec.

This time can be compared with a characteristic time for radiative cooling derived as follows. The enthalpy content of a column of atmosphere of unit cross section is $h_b H \rho_b$. Assuming that the atmosphere is optically thick, this column loses energy by ra-

diation to space at a rate equal to σT_b^4, where σ is the Stefan-Boltzmann constant. The radiative cooling time, τ_R, is the ratio of these two quantities. Typical values of τ_R for the data of Table II are 5×10^3 sec. Evidently, at these relatively low temperatures, expansion of the gas into space is a more rapid process than radiative cooling.

Meanwhile, the shock is traveling horizontally with a speed of some tens of kilometers per second. In the characteristic time for vertical acceleration the shock will typically travel a few hundred kilometers. For impactors with average radii much greater than a few hundred kilometers, therefore, horizontal expansion of the shock is slow compared with its vertical expansion in the sense that most of the enthalpy is converted into vertical mass motion before most of the kinetic energy of horizontal mass motion is converted to enthalpy.

Once the shock is no longer being driven by the impactor, its velocity depends on the pressure jump, which depends on the energy content of the shock-heated gas. Vertical expansion of the shock-heated gas implies that the pressure behind the shock decreases with altitude less rapidly than the density outside the shock. The ratio p_b/ρ_a increases with altitude. But this ratio is just the square of the shock speed, by Eq. (3). Therefore, the speed of horizontal propagation of the shock increases with altitude and refraction introduces a downward component of the motion of the shock. This downward component will be reflected from the ground, probably causing evaporation of ground material, but the overall effect of refraction will be to keep the shock propagating horizontally over the surface of even a spherical planet. Mass motion in the shock will not acquire a vertical component, and the kinetic energy of this mass motion is not directly available to drive atmospheric gasses off into space. The process is indirect. As the shock propagates into the ambient atmosphere, its kinetic energy of mass motion is converted into heat and compression of the ambient atmosphere (into enthalpy). This enthalpy then causes the atmosphere to expand vertically into space in the manner outlined above.

This conversion of kinetic energy into enthalpy at the shock follows much the relations already derived, except that the impactor is no longer feeding energy into the gas. I assume that evaporation of the material of the impactor and the planet continues at the same relative rate as during the impact event itself, because the shock-heated gas is still in contact with the surface of the planet as well as with impact ejecta. Then, the shock traveling horizontally at speed V generates enthalpy per unit mass in the mixture of air and vapor behind the shock with a value $h_b = (V^2/2)/(1 + E)$, where E is given by Eq. (7) with V_s replaced by V. If this gas is to escape from the gravitational field of the planet, its enthalpy per unit mass must exceed $V_e^2/2$, where $V_e = (2gR_p)^{1/2}$ is the escape velocity and R_p is the radius of the planet. Therefore, the gas behind the shock has enough enthalpy to escape from the gravitational field of the planet only if the velocity of the shock is sufficiently large:

$$V > V_e(1 + E)^{1/2}. \qquad (22)$$

It follows at once that an impact event will not cause significant erosion of a planetary atmosphere unless the impact velocity also satisfies the condition in Eq. (22). Since E is very much larger than 1 for large V_s, as shown in Table I, this condition becomes $V_e < (2Q/\Lambda)^{1/2}$ for large V_s. With $Q = 8 \times 10^{10}$ erg/g and $\Lambda = .16$, I calculate that significant impact erosion does not occur from the atmosphere of a planet with escape velocity greater than 10 km/sec.

This result is surprising, so it may be appropriate to review the assumptions that have led to it. First, I have assumed that the target planet gets in the way of escape so that the projectile can not simply sweep through the atmosphere and carry some gas away with it. The velocity of the impactor

is directed downward and the velocity it imparts to the atmospheric gasses is initially directed downward also. This downward velocity is converted into the upward velocity that can result in escape by compression of the gas and consequent increase in temperature, followed by expansion of hot gas into the vacuum of space. Gas cannot escape unless it has enough enthalpy at the end of the episode of compression. Quite different considerations apply to the escape of solid debris from the planet. Compression, expansion, enthalpy, and temperature are not important for the incompressible solids. Momentum of mass motion is. For a compressible fluid like the atmosphere, however, momentum of a part is rapidly shared with the surrounding medium. A portion of the solid planet can blast its way out into space, but a portion of the atmosphere can not. Instead, it must expand into space. For this it must have enough enthalpy. These arguments do not apply to the very small fraction of the atmosphere above the exobase, where the gas no longer behaves as a continuous fluid. Some molecules must therefore be knocked out into space for any impact at speeds in excess of the escape velocity, but the fraction of the atmosphere above the exobase is negligibly small.

The precise value for the maximum escape velocity for which erosion is possible depends, of course, on the evaporation efficiency, Λ. This parameter is not precisely determined although the agreement between different experimental evaluations and different theoretical evaluations is quite close (Bronshten, 1983). The value of this parameter depends on the nature of the atmospheric gasses, the speed of the interaction, the material of the impactor, and also the fraction of the energy of interaction that is absorbed by the impactor relative to the fraction dissipated into the ambient atmosphere. The opportunity for energy to escape into the ambient atmosphere must be less for a very large impactor than for the meteorites which have provided the

data leading to the value of Λ that I have used. Other aspects of the interaction are not changed by the size of the impactor. Therefore, the value I have used for Λ is, if anything, an underestimate and the limit on the escape velocity for which impact erosion is possible is therefore an overestimate.

Consider now the situation in which the impact velocity is large enough and the escape velocity small enough that the shock-heated gas does have enough enthalpy to escape. At the end of the impact event the mass of shock-heated gas and vapor is $\pi R^2 \rho_a (1 + E)$, and this gas has enough enthalpy to escape. It also has a kinetic energy of horizontal mass motion equal to this mass multiplied by $V_s^2/2$. The shock speed decreases as the shock front moves outward and ambient gas is compressed, heated, and accelerated. The kinetic energy of mass motion is converted into enthalpy in the shock-heated gas and this enthalpy drives the vertical expansion. The gas behind the shock front can escape as long as the shock speed satisfies Eq. (22). I calculate the mass of gas that receives enough enthalpy to escape by considering the conservation of energy. The condition is

$$MV^2/2 = \pi R^2 \rho_a (1 + E) V_s^2/2 \qquad (23)$$

where M is the mass of gas and vapor behind the shock traveling horizontally at speed V. The mass that acquires enough enthalpy to escape is derived by setting $V = V_e(1 + E)^{1/2}$. Thus,

$$M_e = \pi R^2 \rho_a V_s^2 / V_e^2. \qquad (24)$$

This mass is made up of atmospheric gas and the vapor of the impactor and the planet in the proportions $1 : E$. The mass of atmospheric gas that escapes is therefore $M_e/(1 + E)$. This is just the mass of atmospheric gas intercepted by the impactor, $\pi R^2 \rho_a$, multiplied by $V_s^2 / V_e^2 (1 + E)$. Impact erosion occurs only when this enhancement factor is greater than 1. For large impact speed the enhancement factor approaches $2Q/(V_e^2 \Lambda)$ or approximately $10^{12}/V_e^2$.

The shock is focused by spherical geometry onto the point on the planetary surface opposite to the impact, but by this time it has accelerated and heated the entire atmospheric mass as well as planetary and impact vapor. It is not likely that this focusing will generate enough enthalpy to cause escape of gas from the antipodal point, although, for very large impact events, the point might merit further study.

IMPACT EROSION

According to this theory, therefore, the mass of atmospheric gas driven off into space by a large extraterrestrial impact is proportional to the mass of atmospheric gas encountered by the impactor, a result that is hardly surprising. The constant of proportionality is equal to the square of the ratio of impact velocity to escape velocity divided by $1 + E$, where E is the proportion of shock-heated gas contributed by vaporized impactor and planetary material. The parameter E describes the additional burden imposed on escape by the need to carry along vapor as well as atmospheric gas. Its effect is to make impact erosion negligibly small unless the impact speed is significantly larger than the escape velocity. Moreover, the mass of atmosphere removed becomes independent of impact speed at large impact speeds. In addition, impact erosion is not possible from a planet with escape velocity larger than about 10 km/sec, no matter how large the impactor or how large the impact velocity.

The theory that I have presented here involves many severe simplifications of a complicated set of physical interactions. The quantitative aspects of the theory can not be considered very reliable. Many elements of the interaction would merit much more detailed study than I have presented here. It appears that detailed predictions of the theory depend on the precise geometry of the interaction between the impactor and the planet. I have chosen to approximate the geometry of this interaction as a vertical, cylindrical piston expanding in radius at a steady rate. An alternative description might deal with the geometry of a sphere being engulfed by a plane, with multiple interactions between shock waves reflecting off the two solid surfaces. The treatment of geometry is not likely to impact the conclusions of this paper concerning the important physical processes and the general conditions for impact erosion. In spite of its simplicity, this theory is free of arbitrarily assumed parameters. Even the parameter Λ, which describes the fraction of shock energy used to evaporate the impactor, is known from theory and observations of meteoric phenomena.

The results presented here are quite different from those of Watkins (1983). In particular, a massive atmosphere does not, in this theory, inhibit the erosion process. This insensitivity to atmospheric density results from emphasis on the interaction of the impactor with the atmosphere. The energy deposited in the atmosphere is proportional to atmospheric density. The energy required to cause atmospheric escape is also proportional to atmospheric density. Watkins directed attention to energy released in the interaction between the impactor and the solid planet and considered the return of some of this energy to the atmosphere. As I have shown above, interaction with ejecta and rebounding planetary material is less important than the initial interaction with the impactor itself because of the significantly lower speeds of the rebounding material and the sensitivity of shock heating to speed. The results presented in this paper could usefully be applied to studies like that of Watkins (1983) of the effect on planetary atmospheric evolution of impact erosion caused by large numbers of impactors of diverse sizes and speeds.

REFERENCES

BRODE, H. L. (1968). Review of nuclear weapons effects. *Annu. Rev. Nucl. Sci.* **18**, 153–202.

BRONSHTEN, V. A. (1983). *Physics of Meteoric Phenomena.* Reidel, Dordrecht.

CAMERON, A. G. W. (1983). Origin of the atmospheres of the terrestrial planets. *Icarus* **56**, 195–201.

FEGLEY, B., R. G. PRINN, H. HARTMAN, AND G. H. WATKINS (1986). Chemical effects of large impacts on the Earth's primitive atmosphere. *Nature* **319**, 305–308.

HAWKINS, G. S. (1964). *Meteors, Comets, and Meteorites.* McGraw–Hill, New York.

JONES, E. M., AND J. W. KODIS (1982). Atmospheric effects of large body impacts: The first few minutes. In *Geological Implications of Impacts of Large Asteroids and Comets on the Earth* (L. T. Silver and P. H. Schultz, Eds.), pp 175–186. Geological Soc. America, Sp. Paper 190, Boulder, Colorado.

LEWIS, J. S., G. H. WATKINS, H. HARTMANN, AND R. G. PRINN (1982). Chemical consequences of major impact events on Earth. In *Geological Implications of Impacts of Large Asteroids and Comets on the Earth* (L. T. Silver and P. H. Schultz, Eds.), pp. 215–221. Geological Soc. America, Sp. Paper 190, Boulder, Colorado.

MATSUI, T., AND Y. ABE (1986). Evolution of an impact-induced atmosphere and magma ocean on the accreting Earth. *Nature* **319**, 303–305.

MELOSH, H. J. (1984). Impact ejection, spallation, and the origin of meteorites. *Icarus* **59**, 234–260.

O'KEEFE, J. D., AND T. J. AHRENS (1982). The interaction of the Cretaceous/Tertiary extinction bolide with the atmosphere, ocean, and solid Earth. In *Geological Implications of Impacts of Large Asteroids and Comets on the Earth* (L. T. Silver and P. H. Schultz, Eds.), pp. 103–120. Geological Soc. America, Sp. Paper 190, Boulder, Colorado.

OPIK, E. J. (1958). *Physics of Meteor Flight in the Atmosphere.* Interscience, New York.

VINCENTI, W. G., AND C. H. KRUGER (1975). *Introduction to Physical Gas Dynamics.* Huntington, New York.

WATKINS, G. H. (1983). *The Consequences of Cometary and Asteroidal Impacts on the Volatile Inventories of the Terrestrial Planets.* Ph.D. dissertation, Massachusetts Institute of Technology, Cambridge.