Hypervelocity Impacts and Magnetization of Small Bodies in the Solar System

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Received October 10, 1994; revised January 17, 1995

The observed magnetism of asteroids such as Gaspra and Ida (and other small bodies in the solar system including the Moon and meteorites) may have resulted from an impact-induced shock wave producing a thermodynamic state in which iron-nickel alloy, dispersed in a silicate matrix, is driven from the usual low-temperature, low-pressure, ω-melt, melt, phase to the paramagnetic, ε (hcp), phase. The magnetization was acquired upon rarefaction and remelt into the ferromagnetic, α, structure. The degree of remagnetization depends on the strength of the ambient field, which may have been associated with a Solar-System-wide magnetic field. A transient field induced by the impact event itself may have resulted in a significant, or possibly, even a dominant contribution, as well. The scaling law of Housen et al. (Housen, K. R., H. B. Schmidt, and R. A. Holmapple 1991. Icarus 94, 180–190) for catastrophic asteroid impact disaggregation imposes a constraint on the degree to which small planetary bodies may be magnetized and yet survive fragmentation by the same event. Our modeling results show it is possible that Ida was magnetized when a large impact fractured a 125 ± 22-km-radius protasteroid to form the Koronis family. Similarly, we calculate that Gaspra could be a magnetized fragment of a 45 ± 15 km-radius protasteroid. © 1995 Academic Press, Inc.

1. INTRODUCTION

Determinations of the fossilized direction and intensity of magnetic polarization of the magnetizable material in igneous and sedimentary rocks in the Earth's crust have been used to good-effect to trace the history of the main geomagnetic field (see, e.g., O'Reilly 1984). It is generally accepted that this field is generated by the self-exciting magnetohydrodynamic (MHD) dynamo mechanism operating in the Earth's liquid metallic outer core. More controversial have been attempts to interpret the remanent magnetization of the Moon's crustal rocks and of the material of other "small" objects (e.g., asteroids) in the Solar System, all of which have, at some stage, been involved in hypervelocity impacts. It is thus clearly a matter of importance to investigate whether they could have been magnetized by impacts (Hide 1972, 1993).

When the Galileo spacecraft flew past Asteroid 951, Gaspra, (mean radius 7 km) in October 1991, its magnetometers produced evidence of a surface magnetic field which is within an order of magnitude of that of the Earth's surface field (0.5 Gauss) and a magnetic moment per unit mass within the range for iron meteorites and highly magnetized chondrites (Kivelson et al. 1993a). This interpretation was strengthened by subsequent theoretical calculations on the interaction of Gaspra with the solar wind (Buamgärtel et al. 1994). This important finding and the subsequent discovery that the larger Asteroid 243, Ida, appeared also to be magnetized (Kivelson et al. 1993b) has provided a fresh stimulus to study of impact magnetization.

In any hypervelocity impact between two initially solid bodies, rapid compression of material occurs near the point of impact followed by rapid decompression (stress unloading). The stress wave which propagates into the colliding bodies with a velocity exceeding that of sound, c, in the undisturbed medium by an amount which depends on the local amplitude of the disturbance. Because of both the deposition of irreversible work and geometri c attenuation, stress wave amplitude decreases with increasing distance X from the point of impact. The transient and any permanent physical (including magnetic) and chemical changes experienced by a material element depends inter alia on X, and in hypervelocity impacts, where the impact speeds greatly exceed the speed of sound in
the undisturbed bodies, four general regimes can be distinguished, namely (a) \( X > X_c \), (b) \( X_c < X < X_o \), (c) \( X < X_o \), and (d) \( X < X \), (Hild 1972). Within the first of these regimes \( (X > X_c) \), at great distances from the point of impact, the pulse is so weak that it travels at the ordinary ambient sound speed \( c \) and produces no permanent changes in physical (and chemical) properties. At the other extreme, within the fourth regime \( (X < X) \), near the point of impact, the pulse is so strong that, owing to the very high temperatures produced in the compressional phase, much material is vaporized and ejected rapidly into the surrounding space, initially as a mixture of ionized gas (plasma), excited ions, neutral gas, and dust particles at high velocities. Research is needed to fully elucidate these complex processes occurring in the ejecta. Of particular importance will be detailed quantitative studies of the transient magnetic fields produced in the ejecta regime.

Of the two intermediate regimes, where \( X_o < X < X_c \) and \( X < X < X_o \), respectively, it is the former that is closer to the point of impact and within which material experiences pronounced metamorphism. During the compressional phase of the supersonic pulse propagating across that regime, material suffers shock heating to temperatures insufficient to cause vaporization, but high enough (in excess of \( 10^8 \) K) to cause melting. Because Curie point temperatures are substantially less than melting points, any ferromagnetic material will be demagnetized during the compressional phase and then remagnetized in subsequent cooling by various processes (decompression, radiation, convection, and conduction) in any ambient magnetic fields, at a rate determined by the dominant cooling process.

We are particularly interested here in the outer of the two intermediate regions (where \( X_o < X < X_c \)), defined as that region where shock heating during the compressional phase of the pulse is too weak to produce melting, but is sufficient to produce magnetic changes. As with the other intermediate region, any ferromagnetic material would suffer demagnetization as its temperature rises through the Curie point during the compressional phase. Subsequent cooling through the Curie point, including rapid cooling during the decompressional phase of the pulse, causes remagnetization of the material to an intensity dependent on the strength of any magnetizing field present at the time.

It is possible, but by no means certain, that thermal effects associated with the pulse generated by a hypervelocity impact provide the main process responsible for changing the magnetic state of the material through which the pulse passes. In this paper we suggest structural changes which occur at relatively low pressures for Fe, FeNi, and magnetite are the most common demagnetization mechanisms. Other mechanisms could also be at work, possibly involving direct pressure-induced magnetization changes in an orientation determined by the stress wave propagation direction. The time scales associated with magnetic state changes \( (\sim 10^{-11} \) sec) are very much shorter than the duration of the compressional and decompressional phases of the pulse.

Figure 1 is a sketch of an impact event resulting in target fragmentation. The principal aim of the present paper is to evaluate the extent to which shocked magnetizable material in region \( X_o < X < X_c \) could undergo magnetic changes without either disintegrating mechanically or escaping from the gravitational field of the parent body. This is clearly a necessary step toward an understanding of the magnetic properties of small bodies within the Solar System. As already noted, another necessary step will be further research on bursts of electric current generated in ejecta from hypervelocity impacts, but this lies beyond the scope of the present study.

II. MAGNETIZATION OF ASTEROIDS BY HYPERVELOCITY IMPACTS

A. Magnetic Minerals in Asteroids and Meteorites

Most meteorites originate from the main asteroid belt, and their chemical composition is directly related to that of asteroids. Meteorites are classified on the basis of their metal (iron-nickel) content, as "iron" (principally metallic), "stony iron" (containing abundant metal and silicates), and relatively metal-poor "stones," which are subdivided into "chondrites" and "achondrites," the latter being a small category with very diverse mineralogy. Chondrites are by far the most numerous of all types. The mean metal-element content is about 18–33 wt%, the oxidation state of Fe varies with different classes; the enstatite class is the most highly reduced, with about 70% of its Fe being in its metallic form; carbonaceous chondrites, on the other hand, are highly oxidized and contain little to no metallic iron, and magnetite is present to a significant amount in some of them. The third class comprises the most abundant ordinary chondrites, which
are intermediate between the two extremes, containing significant amounts of iron in both metallic and oxidized forms (e.g., Dodd 1981).

Asteroids are classified largely on the basis of their surface reflectance spectra at wavelengths between 0.3–1.1 μm, supplemented by reflectance in longer wavelengths and radar reflectance data (see e.g., Bell et al. 1989). Although reflectance in the infrared to visible range is an indication of the mineralogy of the asteroids, considerable diversity is still exhibited by each group and there is a considerable mismatch with the meteorite classification. Nonetheless, analogs between asteroid types and meteorite classes have been proposed (e.g. Dodd 1981; Bell et al. 1989). Of the major asteroid types, the S-type (of which both Gaspra and Ida are members) has reflectance spectra not unlike those of carbonaceous chondrites, ordinary chondrites, and iron; The C- and M-type asteroids are believed to correspond to carbonaceous chondrites (CI–CM) and irons, respectively. Therefore, we assume, for the purpose of the present paper, that the magnetic carrier in S-asteroids is a mixture of metallic FeNi and magnetite.

The history of shock-loading and unloading and the resultant shock-heating are sketched in Fig. 2. In the following sections, we first model impact-magnetization of planetary bodies (e.g., S- and M-type asteroids) by the magnetization of ferromagnetic iron metal in them, such as kamacite, before discussing the effects of the presence of iron alloys, which in meteorites contain up to ~20% Ni and minor amounts of other siderophiles (Co, Cr, Mn, W, Ta, V, etc.) and magnetite.

B. Free (Bulk) Iron in Asteroids

When pure iron is the magnetizing medium, three distinct magnetization mechanisms are possible in different shock pressure-temperature regimes:

1. If the locus of shock states (Hugoniot) starting in the α field in the P–T plane crosses the Curie point at pressures between 0 and about 1.75 GPa (Fig. 3), natural

FIG. 2. (a) Shock-heating of rock. Temperature is raised from initial $T_i$ to $T_f$ as shock waves arrive. After the rocks are fully unloaded, the temperature remains at the post-shock, intermediate value $T_m$. (b) The Hugoniot (state 1 → II) and unloading path (state II → III) is the P–V plane. The isocounts on specific volume, $V_0$, is shown for reference. States I, II, and III correspond to those in (a). $V_i$ is the initial volume (including pores), $V_f$ is the volume of relaxed state, and $P_i$ and $P_f$ are Hugoniot and isocount pressures at compressed volume $V$.

FIG. 3. Shock temperatures vs shock pressure of gabbroic nor- thosite of different porosity values. "100% density" is 2.906 g/cm$^3$. Iron and magnetic phase diagrams (after Bandy 1965, Liu and Bassett 1986) are superimposed to demonstrate different transitions at different porosities. The dashed line with the label "CP" is the Curie temperature of iron. The dashed lines labeled (i) and (ii) are magnetic and structural phase transitions of magnetite (after Liu and Bassett 1986).
“Curie-point writing” (a term usually used in the field of magnetic memory) (Cullity 1972) occurs during or after being shock-heated to above the Curie temperature (1043 K for pure iron at 1 bar (Morrish 1965)). The phase change is second order. Owing to the weakness of its dependence on pressure (Osmol-chekov et al. 1969, Mirvad and Kennedy 1979) the Curie temperature can be taken as constant. As indicated in Fig. 3, this mechanism requires intensive shock-heating and only occurs upon shocking silicate iron-bearing rocks that are less than ~0% of crystal density.

2. If the P-T locus starting in the a field crosses the phase boundary between (7.75 GPa, 1043 K) and the a-α-γ triple point at (110 GPa, 750 K), iron undergoes a first-order phase transformation from ferromagnetic body-centered-cubic (bcc) structure (α phase) to paramagnetic face-centered-cubic (fcc) structure (γ phase) (Bundy 1965, Liu and Bassett 1986). When on the release of pressure the system returns through the phase boundary, the reverse transition occurs and the material becomes stably magnetized with a magnetization intensity dependent on the ambient field which is present upon pressure and temperature release. Silicate rocks with between 40 and 80% of crystal density containing kaemacite can be magnetized via this method (Fig. 3).

3. Shocked silicate rock with greater than ~80% crystal density may be magnetized upon the crossing of the locus of states with the a-α-γ phase boundary (between the a-α-γ triple point and (273 K, 14 GPa)) (Bancroft et al. 1956, Barker and Holttenhuise 1974). The transition pressure is slightly temperature dependent (from about 14 GPa at room temperature to about 11 GPa at the triple point), but can be taken to be approximately 13 GPa.

Which one of these three mechanisms is actually activated depends on the equation-of-state of the rocks in which iron is embedded and assumed to be in pressure and temperature equilibrium. Our calculations show that owing to the simultaneous requirement of low pressure and high temperature, the temperature-driven mechanism (1) (heating and cooling through Curie point) does not take place for likely loading paths within realistic materials. This point was previously emphasized in Wasilewski (1987a). In experimental studies of shock effects on magnetic properties of iron (Royce 1968, Wong 1969, Keeler and Mitchell 1969), magnetization changes were observed upon exposure of pure metallic samples to dynamic pressures of 8-32 GPa (with large disagreements between different measurements). The observed decrease in magnetization appear to be related to the α → s transformation.

Although shock-induced magnetic phase transitions produce remanent magnetization in, for example, metallic iron-bearing lunar rocks, other thermal and mechanical effects associated with shock waves may also alter the rock’s magnetic properties (Wasilewski 1987d). A shock-induced mechanism suggested for lunar rocks is piezomane- nance, originating from alignment of magnetic domains in minerals, associated with the deformation of the rock upon shock propagation (Brecher 1976). Cisowski et al. (1973, 1975, 1976, 1977) estimated from their experimental results that the normal remanent magnetization (NRM) of lunar samples (10−10−4 Gauss cm3 g−1) could be acquired by shocking to 5 GPa in external magnetic fields of 10−10−5 G (1 year = 10−12 Gauss). Cisowski and Fuller (1978) made similar observations using terrestrial rocks (with magnetite and hematite as the major magnetic carriers).

Experiments of Crawford and Schultz (1991, 1992a,b) suggest that strong magnetic fields can be generated for a short time period after a large impact due to charge separation in the ejecta. Compression of the ionized coma between impactor and target may also amplify the ambient (~10−5 G) interplanetary field (Gold and Soter 1976). Magnetic fields via these mechanisms can magnetize iron minerals in an impacted planetary object as they go through a shock-induced phase transition before the magnetizing field dissipates; e.g., in the Curie point (CP) transition case this requires an above-CV shock temperature and a below-CV postshock temperature.

An unamplified, constant interplanetary field present at the time of impact alone could be sufficient to cause significant magnetization, especially considering that the solar field had been stronger in the past. Therefore to induce Curie point magnetization all the volume shocked to temperatures higher than CP must be subsequently cooled either by rarefaction-wave isentropic pressure release quenching or heat loss via thermal conduction, convection, or radiation. Upon first-order (polymorphic) phase changes, the reverse transitions generally take place rapidly upon unloading; therefore little difference in magnetization is expected upon impact in the presence of a transient or steady magnetic field.

In the present study we assume metallic iron is dispersed in a largely silicate or hydrosilicate matrix containing a wide range of minerals. However, for the bulk rock we assume the properties of a generic silicate-bearing rock (e.g., gabbrro anorthosite).

The equation-of-state assumed for a generic rock material is the Tulloton equation-of-state relating pressure P, specific volume V, and specific internal energy E, where $\eta = V T$. For gabbrro anorthosite (Alvarez and O’Keefe 1977), $a = 0.5$, $b = 0.145$, $A = 0.765$ Mbar,

$$P = \left( a + \frac{b}{(E/K)^{1/3}} \right) \frac{E}{V} + A (\eta - 1) + B (\eta - 1)^2,$$   \hspace{1cm} (1)
\( B = \text{0.751 Mbar}, E_0 = 4.89 \times 10^{12} \text{ erg/g}. \) For the Hugoniot (Fig. 2b),

\[
E - E_0 = P_d(V_{is} - V)/2,
\]

(2)

where \( V_{is} \) is the initial specific volume. For the isentropic (Fig. 2b),

\[
E = E_0 - \int_{V_i}^{V_f} P_d(V') \, dV'.
\]

(3)

Shock temperature \( T_s \) and postshock temperature \( T_{ps} \) (Fig. 2a) are calculated according to Ahrens (1987)

\[
E_0 - E_i = \int_{T_i}^{T_s} C_i \, dT,
\]

(4)

where the isentropic temperature at volume \( V \) is

\[
T_s = T_i \exp \left[ - \int_{V_i}^{V} \left( \frac{C_i}{V} \right) \, dV' \right],
\]

(5)

where \( T_i \) is the initial temperature (298 K is assumed). The Grüniesen parameter is calculated according to its definition

\[
\gamma = V \left( \frac{dp}{dE_t} \right).
\]

(6)

by differentiating Eq. 1, we get

\[
\gamma = 0 + \frac{b}{((E/E_0)^{2} + 1)^3}.
\]

(7)

Postshock temperature is calculated along isentropic release as formulated in Ahrens (1987),

\[
T_{ps} = T_i \exp \left[ - \int_{V_i}^{V_f} \left( \frac{C_i}{V} \right) \, dV' \right],
\]

(8)

where \( V_i \) is the specific volume in the released state. Shock temperature calculations are shown in Fig. 3 for several different porosities. Postshock temperature calculations are shown in Fig. 4. In determining whether iron can be magnetically rewritten we only require that the target rock be shocked into paramagnetic regime and relax to a postshock state (where iron reverts to ferromagnetic \( \alpha \) phase during unloading or after long-term cooling at zero pressure). The latter corresponds to the ambient magnetic field assumption.

Peak shock pressure distributions in planetary bodies for impacts are calculated as a function of impact velocity using the Holzapfl and Schmidt scaling relations (Holzapfl and Schmidt 1987). The peak shock pressure, \( P \), decays with the distance from impact center, \( r \), following different power laws in different regimes: For \( r < \text{impactor radius } a, P = \text{constant.} \) For \( r > a, P \propto r^{-3/2} \) (\( a \) is the exponent in the "coupling parameter," see Eq. (9) in Section III) until \( P \) drops to below \( \rho c^2 \), \( \rho \) are density and longitudinal sound speed of the target), then \( P\) follows \( r^{-3/2} \). More detailed pressure distributions for finite targets can be obtained via numerical simulations (e.g., Asphaug and Melosh 1993). The half-space scaling law gives a valid description for our purpose, when \( R \gg a \) (\( R \) is the shock radius and \( a \) is the projectile radius) and impact speeds of 2–50 km/sec. Together with the shock/postshock temperature to shock pressure relations, the radius of magnetization, \( R_{m} \), from center of impact at different velocities is obtained and plotted in Fig. 5. For a solid gabbroic norite host, the effect of variations in the \( a \sim c \) transition pressure is illustrated by the two dashed curves, corresponding to transition pressures of 11 and 14 GPa respectively, bracketing the solid. 13 GPa curve.

C. Iron–Nickel Alloy and Magnetite in Asteroids

Since iron meteorites are always composed of an Fe–Ni alloy, we assume the iron in asteroids is always alloyed.
with other transition metals, iron alloys (with small amount of V, Si, Co, Cr, Ni, or Mn) have similar phase diagrams as pure iron but with shifted phase boundaries (Devall and Graham 1977). We choose Fe–20 wt% Ni alloy as a representative composition for the effect of various solutes may have on the metal’s magnetization. At 1 bar, an alloy of this composition has a much lower CP (~693 K) than pure iron, also, the As transformation occurs at about 873 K (Metals Handbook, 8th ed., p. 304). The transition which corresponds to the "13-GPa" transition in pure iron occurs at about 10 GPa in this alloy (Devall and Graham 1977). The pressure dependence of the z = y boundary and the Curie temperature is unknown. If the a = x transformation governs the magnetic properties for an iron-nickel alloy in a nonporous silicate astrosphere, as it does for pure iron, the lowering of the transformation pressure increases R_A by about 14% for 5 km/sec impacts. This is a relatively small effect because shock pressure decays exponentially with distance far away from impact center.

Compared to iron metal or iron alloys, P–T phase diagram of magnetite (Fe_3O_4) is less well known (Liu and Bassett 1986). A magnetic transition and a structural phase transition (Mao et al. 1974) are plotted and labeled as (i) and (ii) in Figs. 3 and 4. Ambient ferromagnetic magnetite becomes paramagnetic at the Nd temperature of ~850 K at 1 bar, between T_C’s of iron and iron-nickel. Magnetic properties of the high-pressure phase (of unknown structure) were studied with Mössbauer spectroscopy (Mao et al. 1977) and found to be paramagnetic. This transition is reversible. The transition pressure is ~25 GPa at room temperature and has a slope of ~ ~22 MPa/K. Thus, when NMR results from magnetite, it is slightly more stable with respect to impact than that for iron-nickel alloy.

III. MASS OF THE LARGEST FRAGMENT FROM AN IMPACT: THE STRENGTH AND GRAVITY REGIMES

It was proposed by Holsapple and Schmidt (1987) that the stress wave and the cratering flow field induced by an impact of projectile of density ρ, radius a, and velocity U can be characterized by a coupling parameter (except for a region very near the impact point):

\[ C = a U^5 \delta^7. \]  

By fitting experimental data of crater volume and the growth of the crater with time as a function of parameters characterizing the projectile and target, Holsapple and Schmidt reported a varies from 0.4 for dry sand to 0.55 for competent rocks; \( \delta \) is about 1/6. The term \( \delta \) is relatively constant.

Based on the Holsapple-Schmidt scaling and dimensional analysis, Housen and Holsapple (1990) derived a relationship for the mass of the largest fragment after an impact (M_F) relative to the projectile mass (M_p):

\[ \frac{M_F}{M_p} = \rho [\rho_S \rho_I] \]  

where the strength and gravity scaling parameters \( \rho_S \), \( \rho_I \) are

\[ \rho_S = \frac{Q [G^3 a^2 \rho^{-1/2} R^{-1}]^{-1}}{U^{-1} [G^3 a^2 \rho^{-1/2} R^{-1}]} \]  

\[ \rho_I = \frac{Q [G^3 a^2 \rho^{-1/2} R^{-1}]}{U^{-1} [G^3 a^2 \rho^{-1/2} R^{-1}]} \]  

where \( \rho, R \) are the density and radius of the target, and \( G \) is the gravitational constant (6.67 \times 10^{-8} \text{ dyne cm}^2 \text{g}^{-2}).

In the last two equations a is represented through the impact energy per unit mass of the target:

\[ Q = \frac{\delta U^5}{2 \rho R^2} \]
\[ Y_r = S \sigma_c R^3. \tag{14} \]

The property parameters for the impactor \((\alpha, U, b)\) as expressed in Eqs. (11) and (12) (and therefore in Eq. (10)) only appear in the form of the single variable \(C\) in Eq. (9).

A qualitative insight into the failure process can be gained by noting the scaling of the prevailing loading strain rate

\[ \dot{\varepsilon}_i = \frac{U(3b - 2\pi a) R^{3/2}}{R}, \tag{15} \]

and that of the tensile stress caused by the impactor is given by

\[ \sigma_r = \rho Q^{3/2} R^{3b - 2\pi a - 3/2} \tag{16} \]

(see Eq. (95) and Table 1 in Holzapfl and Schmidt (1987)). Forming the fraction \(Y \sigma_r^{3b-2}/c_1\) from the last two equations and comparing to the right-hand side of Eq. (11) yields

\[ \frac{Y_1}{\sigma_r} = \Pi_4 (3b - 2\pi a - 3/2). \tag{17} \]

Similarly, note that the compressive lithostatic stress at a radial distance \(r\) from the center of a homogeneous body of radius \(R\) is

\[ \sigma_0 = \frac{2}{3} \pi \left( 1 - \left( \frac{r}{R} \right)^2 \right) G_0^2 R^2 \approx k_r G_0^2 R^3, \tag{18} \]

where

\[ k_r(r) = \frac{2}{3} \pi \left( 1 - \left( \frac{r}{R} \right)^2 \right); \tag{19} \]

taking ratio of \(\sigma_0/\sigma_r\) and comparing to the right-hand side of Eq. (12) it follows that

\[ \frac{\sigma_0}{\sigma_r} = k_\Pi_4 3/2. \tag{20} \]

The \(\alpha\)'s are magnitudes and thus always positive by definition, but bear in mind \(\sigma_r\) is compressive while \(\sigma_0\) is tensile. Eqs. (17) and (20) show \(\Pi_4\) and \(\Pi_0\) are simple functions relating the condition required for fracture and the ambient lithostatic stresses, normalized by stock-induced tensile stress. Equation (10) is thus rewritten as

\[ \frac{M_i}{M} = F \left( \frac{Y_1}{\sigma_r^3} \right) \left( \frac{\sigma_0}{\sigma_r} \right) \tag{21} \]

Physically, the tensile wave must overcome the lithostatic compressional stress before fracturing occurs; therefore we can expect \(M_i/M\) to be a fraction of a single variable \((Y_1/\sigma_r^3)\alpha_1\). In particular, Hounsfie et al. (1991) chose the functional form

\[ \frac{M_i}{M} = 1 - \frac{3b - 2\pi a - 3/2}{3b - 2\pi a - 3/2} \left( \frac{Y_1}{\sigma_r} + \sigma_0 \right)^{3b - 2}/2 \tag{22} \]

(This equation is modified from Eq. (14) in Hounsfie et al. (1991) with the substitution \(q = U/2\)). This form also satisfies the asymptotic requirement that when \(M \to 0\) (cratering into half space), \(M \to M_i\) (ejecta mass) should be independent of \(M\). The proof is given in Hounsfie et al. (1991) following Eq. (14).

Denoting the constants of proportionality in Eqs. (15) and (16) by \(k_r\) and \(k_\Pi\) and substituting \(Y_1\), \(\sigma_0\), and \(\sigma_r\) with \(\Pi_4\) and \(\Pi_0\) (via Eqs. (17), (20), and (16)), we get the final equation used in our calculations:

\[ \frac{M_i}{M} = 1 - \left( k_r G_0^2 \Pi_4 (3b - 2\pi a - 3/2) + k_\Pi G_0^2 \Pi_0 \right)^{3b - 2}/2, \tag{23} \]

Hounsfie et al. (1991) obtained a good fit to their pressurized fragmentation experimental data on basaltics with \(K = 0.13\). Hounsfie and Holzapfl (1990) suggested \(k_r \approx 1\). We adopt these values in our calculations. Equating \(M_i/M\) to 0.5, the catastrophic fragmentation threshold (CF) is obtained:

\[ 2^{3b - 2} K(k_r G_0^2 \Pi_4 (3b - 2\pi a - 3/2) + k_\Pi G_0^2 \Pi_0) < 1; \tag{24} \]

Equations (23) and (24) are solved numerically in conjunction with Eq. (19), requiring that \(r\) be equal to the largest fragment radius. This is an attempt to extend theoretically the results of Hounsfie et al.'s (1991) experiments under uniform overpressure to an actual gravitational body in which lithostatic stress is spatially varying.

The above calculation does not consider the effect of postimpact reaccumulation of the fragments under mutual gravitational into a "rubble pile" which would increase \(M_i/M\). Hounsfie and Holzapfl (1990) proposed a reaccumulation scaling which introduced an additional parameter \(k_r\). The critical energy per unit target mass at CF was in-
ceased by nearly two orders of magnitude when reaccu-
   mulation was taken into account (ref. Fig. 5 in Housten and
   Holmesape 1990) and was in general agreement with
   the estimates by Pijz (1982) of the impact energy for
   the Hainayam families. However, subsequent exper-
   iments revealed a higher energy density at CF under exter-
   nal overpressure (Housten, et al. 1991). The modified
   scaling (Eq. 24) gives a energy density comparable to the
   previous scaling with reaccumulation taken into account.
   Therefore the agreement with the asteroid family data
   calls into question the role of accumulation, which has
   not been addressed in laboratory experiments to date.
   Uncertainties in Eq. (23) arise from values of K, μ, p, k,
   S, and φ. Of these quantities, K (0.13) and μ (0.55-0.66) for
   nonporous materials) are fairly well determined (Housten,
   et al. 1991, Holmesape and Schmidt 1987); we assumed p
   may vary between 1 to 8 g/cm³. The optimum value for k,
   is 1; in our calculations we assigned it extreme values of
   0.3 and 3. S and φ are material strength parameters; they
   are correlated with assumed fracture behavior of the mate-
   rial for rocks S is nominally 10⁶ in cgs units (Housten and
   Holmesape 1990) and may vary by an order of magnitude.
   Strain rate-dependent fracture models usually assume φ
   ~ S; φ ~ σ yields the strain rate-independent behavior.
   Calculations of fragment radius r (normalized to impactor
   radius a) for 1 km and 1000 km radius targets were carried
   out using optimum and extreme values (Fig. 6).

IV. CONCLUSIONS
A. Impact-Magnetization of Asteroids
Iron and Fe-Ni alloys commonly occurring in meteor-
ites have similar phase diagrams which feature:

1. Second-order ferromagnetic to paramagnetic phase transition (Curie point transition) in the normal a struc-
ture, requiring shock-heating to relatively high tempera-
tures at a low stress level (~1 GPa);
2. Transformation from α to γ phase at intermediate
   shock pressures and temperatures;
3. Transformation from α to ε phase at relatively high
   pressures (~10-20 GPa) and low to moderate tempera-

During hypervelocity collisions, our study indicates that for matrix rock of common densities, dispersed mag-
netic iron grains undergo the α-ε transition and, upon
unloading and cooling, can be impressed with the signa-
ture of magnetic fields present at the time of impact. For
porous target (of 90% or less of crystal density), the mag-
netization mechanism is the α-γ phase change. The Curie
point transition is of limited importance only for targets
of 60% or higher porosities (Figs. 3 and 4).

B. Magnetization without Fragmentation?

Solving Eq. (24), we obtained CF thresholds for plane-
tary bodies of different sizes and densities after impacts
of various velocities, R₉ at CF (i.e., maximum magnetiza-
tion for a given target) is then obtained from Fig. 5. The
results of R₉/R are plotted in Fig. 7a). It is clear that it is
very unlikely or impossible φ to magnetize a small (R < 10⁶
km) planetary body by hypervelocity impact without
severely fracturing it (meaning at least breaking it to
halves). On the other hand, taking into account the inher-
ent statistical uncertainty of the fragmentation process,
there is a narrow window for planetary bodies in the range
of ~10⁵ km in radius for which they may become entirely
magnetized without suffering severe fracture, given the
right-sized impactor. However, the “right” impactor size
has to be very large, and therefore these impacts must
be very rare and the homogenous magnetization in the
object would be subject to later demagnetization and dis-
rupting resulting from smaller and more frequent impact
events.

We note from our calculations that, for large bodies,
solid targets are more prone to fracturing than same-sized
porous counterparts; e.g., the crossover takes place at
R ~ 4 km for 5 km/sec impacts (Fig. 7b). This observa-
tion seems counterintuitive at first, considering the much
lower strength of the porous rock, but becomes obvious
when one realizes that the coupling parameter μ, which
also controls shock wave energy decay, becomes larger
for porous targets (0.55) than for solid targets (0.40). Thus
this effect tends to limit the shock damage in the vicinity
of impact. Another factor is that at ~4 km, gravity is
becoming increasingly important as the cohesive force
that prevents the body from breaking up, and the differ-
ence in gravitational stress between porous and solid targets is much less drastic than the difference in material strength. Also note the energy required to fragment a porous or nonporous target, as a function of target size, goes through a minimum when the transition from strength regime to gravity regime takes place.

C. Applications to Gaspra and Ida

Gaspra and Ida were found to be possibly magnetic during Galileo flybys. If these objects were magnetized by hypervelocity impacts and are remnant fragments from the same impacts that magnetized them, we can obtain a constraint on the minimum sizes of the impactors following the calculations in Section II. Then requiring the largest fragments to be larger than Gaspra or Ida, lower limits on the protocrateroids can be set following equations in Section III. Such derived lower limits of impactors and parent bodies which would leave 7-km (mean radius of Gaspra) and 26-km (half length of Ida)-radius impact fragments are shown in Fig. 8. The impact velocity (5 km/sec) is the most probable impact velocity in the asteroid belt. It is shown that the proto-Gaspra body was 45 ± 15 km and the impactor 7.6 ± 0.8 km in radius; for Ida, the minimum radii for parent body and impactor are 125 ± 22 and 57 ± 2 km respectively (ida = numbers are uncertainties on the lower limits).

By summing over family members, Gradie et al. (1979) estimated the minimum radius of the Koronis (to which Ida is a member) parent body should be 45 km. A larger radius of 55.7 km was obtained by Fujisawa (1982) using the same method but a different detection bias correction.

From the orbital data (velocity and semimajor axis) of the members, Fujisawa also estimated the ejection energy immediately after breakup of the parent body. He then inferred the impact energy and impactor radius assuming a fraction ~1% goes into the ejecta. Finally, using results from fragmentation experiments on basalt (Fujisawa et al. 1977) in the strength regime, $M_e/M$ and radius of the largest fragment are obtained. Fujisawa’s results are compared with the present calculation in Table I. The agreement is within a factor of about 3 for the radii. Fujisawa
TABLE I
Inferred Break Parameters of the Koronis Asteroid Family

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Note</th>
</tr>
</thead>
<tbody>
<tr>
<td>( h_2 = 0.1 )</td>
<td>55.7</td>
<td>This study</td>
</tr>
<tr>
<td>( h_2 = 0.2 )</td>
<td>55.7</td>
<td></td>
</tr>
<tr>
<td>( \sigma (\text{rad}) )</td>
<td>8.6</td>
<td></td>
</tr>
<tr>
<td>( \sigma (\text{deg}) )</td>
<td>6.8</td>
<td></td>
</tr>
<tr>
<td>( \gamma_1 (\text{rad}) )</td>
<td>7.1</td>
<td></td>
</tr>
<tr>
<td>( \gamma_1 (\text{deg}) )</td>
<td>9.6-9.8</td>
<td></td>
</tr>
<tr>
<td>( \beta_1 (\text{deg}) )</td>
<td>20°</td>
<td></td>
</tr>
<tr>
<td>( M_{\text{Lr}} )</td>
<td>3.85-3.01 \times 10^{-3}</td>
<td></td>
</tr>
<tr>
<td>( 1.30-1.78 \times 10^{-2} )</td>
<td>0.55-1.6 \times 10^{-2}</td>
<td></td>
</tr>
</tbody>
</table>

r: fixed value (mean radius of present body).

wara's results are consistently smaller than the present results, suggesting better agreement could be achieved if we assume the largest fragment size \( \gamma_1 \) is smaller than that of Ida. However, this would involve modeling of the recaccumulation processes, which remains an open question (see Section III). Considering the numerous uncertainties in asteroid composition, equation-of-state, fragmentation/recaccumulation processes, and possible variation in impact velocity, we conclude that the present analysis allows, but does not prove, that Ida could have been magnetized when a large impact fragmented a protosolar to form the Koronis family.

ACKNOWLEDGMENTS

Research supported by NASA. Support for R.H. Cabane was provided under Sherman Fairchild/Distinguished Scholar Program. Suggests for improving the manuscript were appreciated by K. Horner and an anonymous reviewer are appreciated. Contribution 5848, Division of Geophysical and Planetary Sciences.

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