

Impact Vaporization: Experimental and Numerical Insights

by

Megan Bruck Syal

B.A., Astrophysics, Williams College, 2007

B.A., Mathematics & Statistics, Williams College, 2007

Sc.M., Geological Sciences, Brown University, 2011

Sc.M., Engineering, Brown University, 2013

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This dissertation by Megan Bruck Syal is accepted in its present form by the
Department of Geological Sciences as satisfying the dissertation requirement for the
degree of Doctor of Philosophy.

Date: _____

Peter H. Schultz, Advisor

Recommended to the Graduate Council

Date: _____

Amy C. Barr, Reader

Date: _____

Carle M. Pieters, Reader

Date: _____

John F. Mustard, Reader

Date: _____

Karen J. Meech, Reader

Approved by the Graduate Council

Date: _____

Peter M. Weber, Dean of the Graduate School

Curriculum Vitæ: Megan Bruck Syal

EDUCATION

Brown University, Providence, RI

M.Sc., Geological Sciences, 2011

M.Sc., Mechanical Engineering, 2013

Ph.D., Geological Sciences, 2014 (anticipated)

Advisor: Peter H. Schultz

Thesis title: “Impact Vaporization: Experimental and Numerical Insights”

Williams College, Williamstown, MA

B.A., Astrophysics *with honors*, 2007

B.A., Mathematics and Statistics, 2007

Advisor: Jay M. Pasachoff

Thesis title: “High-Spectral Resolution Observations of the Solar Corona and Chromosphere”

EMPLOYMENT

Graduate Student	Brown University Geological Sciences Providence, RI	2009-present
Data Specialist	Chandra X-ray Observatory Cambridge, MA	2007-2009
Teaching Assistant	Williams College Astronomy Dept. Williamstown, MA	2004-2007
Planetarium Show Host	Williams College Astronomy Dept. Williamstown, MA	2004-2007

TEACHING

Solid Earth Geophysics

Teaching Assistant	Prof. Donald Forsyth	Fall 2011
	Brown University	

Computational Approaches to Modeling & Quantitative Analysis

Teaching Assistant	Prof. Marc Parmentier	Fall 2011
	Brown University	

HONORS, AWARDS, SERVICE

NASA Earth and Space Science Fellowship (NESSF), 2012-2015

Graduate Student Representative to Geological Sciences Faculty, 2012-present

Rhode Island Space Grant Summer Research Fellowship, 2012

NASA Group Achievement Award, with the EPOXI Science Team, 2011

Brown University First-Year Graduate Fellowship, 2009-2010

Sigma Xi Society Membership, for undergraduate thesis, 2007

Massachusetts Space Grant Research Fellowship, 2006

Williams College Scholar Athlete, Cross Country and Track and Field, 2003-2007

MISSIONS AND COLLABORATIONS

JPL Planetary Science Summer School: Uranus Orbiter and Probe, 2013

EPOXI Mission to Comet Hartley 2: Science Team Member, 2010-2013

Lawrence Livermore National Laboratory: Asteroid mitigation studies, 2010-present

PEER-REVIEWED PUBLICATIONS

Bruck Syal, M., Dearborn, D. S. P., Schultz, P.H. Limits on the use of nuclear explosives for asteroid deflection. *Acta Astronautica*, 90, 103-111 (2013).

Pillitteri, I., Evans, N. R., Wolk, S. J. , **Bruck Syal, M.** XMM-Newton Observation of the α Persei Cluster, *Astronomical Journal*, 145, 143 (2013).

Bruck Syal, M., Schultz, P. H., Sunshine, J. M., A'Hearn, M. F., Farnham, T. L., and Dearborn, D. S. P. Geologic Control of Jet Formation on Comet 103P/Hartley 2. *Icarus*, 222, 610-624 (2013).

Noble, M. W., Rust, D. M., Bernasconi, P. N., Pasachoff, J. M., Babcock, B. A., **Bruck, M. A.** Observing the solar corona with a tunable Fabry-Perot filter. *Applied Optics* 47, 5744-5749 (2008).

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Introduction and Motivation

Upon hypervelocity impact, projectile and target materials are frequently shocked to pressures and temperatures sufficient to induce complete vaporization. Vaporization associated with impacting comets and asteroids controls much of the geophysical and compositional evolution of planetary systems. A sampling of planetary science problems traceable to impact vaporization phenomena include: generation and deposition of nanophase iron within lunar soil rims [Keller and McKay, 1993], the Moon-forming impact [Canup, 2012], accretionary histories of the icy satellites [Ahrens and O’Keefe, 1985; McKinnon, 1989; Canup and Ward, 2009], erosion of the early Martian atmosphere [Melosh and Vickery, 1989; Shuvalov, 2009], target degassing consequences for the terrestrial biosphere from catastrophic impacts [Schultz and Gault, 1990], and impact delivery of materials to planetary surfaces [Moses et al., 1999; Pierazzo and Chyba, 2002; Ong et al., 2010].

Numerical methods to constrain impact vaporization have made enormous strides over the past two decades, perhaps catalyzed (in part) by the fortuitous 1994 impact of Jupiter by Comet Shoemaker-Levy 9 [Crawford et al., 1994]. Such planetary-scale impact “experiments” provide rare glimpses of the remarkable, high-energy vapor plumes released during impact events. Accurate modeling of impact-generated vapor, however, relies heavily upon the availability of robust equations of state and constitutive relations for the target and projectile materials of interest. Continued work to constrain the behavior of geologic materials at high temperatures and pressures, coupled with accelerating parallel computing capabilities, has opened up a new frontier of planetary-scale impact vaporization problems, as evidenced by recent, vigorous work on the Moon-forming impact [Canup, 2012; Ćuk and Stewart, 2012].

While numerical methods are powerful and necessary for considering planetary-magnitude events, experimental strategies to characterize impact vaporization remain

critical for testing the roles of various impactor and target material models. Furthermore, modeling of oblique impacts remains relatively difficult and computationally expensive; certain experimentally resolved trends, including enhanced vaporization at lower incidence angles [Schultz, 1996], are not yet captured numerically. Nature favors oblique impacts: the impact angle probability distribution scales as $\sim \sin 2\theta$ [Gilbert, 1893; Shoemaker, 1962]. Hence, a comprehensive approach to impact vaporization problems will consider impacts beyond the frequently modeled 90° case.

Here we employ a hybrid approach: both experimental and numerical methods provide complementary insights to impact vaporization phenomena. A wide range of planetary science problems are considered, but all are intimately tied to the way in which planetary surfaces and impactors vaporize.

Chapter 1 applies impact vaporization to the problem of delivery; specifically, it investigates the delivery of carbon-enriched meteoritic material to the surface of Mercury. A longstanding question concerns the low reflectance of Mercury, relative to the Moon; immature materials at the lunar surface are 50% brighter than immature materials at Mercury [Denevi and Robinson, 2008]. Since iron (both ferrous and metallic) functions as an important darkening agent at the Moon, the extremely low elemental iron at Mercury, $\sim 1.5\%$ [Nittler et al., 2011; Evans et al., 2012] suggests the need for a previously unidentified darkening agent. Carbon, which is available in bulk from micrometeoritic material, may assist in darkening Mercury. However, in order to quantify the delivery of carbon, it is necessary calculate the fraction of vaporized projectile that remains below escape speed. This calculation is sensitive to impact angle and is carried out in fully three-dimensional (3-D) simulations.

Chapters 2 and 3 experimentally investigate impacts into water ice targets and present paired numerical calculations for benchmarking purposes. Chapter 2 focuses on deriving the vaporized mass of water ice targets under different impact angle and porosity conditions, compares these results to laboratory-scale simulations, and

then extrapolates the insight gained at laboratory scales to planetary scale problems. Chapter 3 provides cross-sectional views of the impact cratering process in porous ice, via “quarter-space” target geometries. These results, which produce time-resolved information on crater dimensions and target compaction, are then contrasted with other well-studied particulate targets (e.g., quartz sand). The manner in which porous snow undergoes irreversible compaction informs not only final crater morphologies but also the way in which vaporization is enhanced in porous media.

Chapter 4 examines the mechanical and thermal consequences of recent cometary impacts at the Moon. Unique properties of cometary impactors include: high velocities, large volatile abundances, high porosities, and the presence of gas and dust-enriched inner comae. These key characteristics drive specific types of lunar regolith processing, assessed here both analytically and numerically, which are consistent with observations of the lunar swirls. Swirl regions consist of elegantly looped bright and dark patterns and are usually associated with paleomagnetic anomalies; originally captured in Lunar Orbiter images, these meandering, ribbon-like patterns are among the most fascinating and least understood lunar surface features. Both the impact vaporization of the icy comet nucleus and the impingement of the gaseous inner coma upon the lunar surface contribute to mobilization, entrainment, and heating of grains over large regions of the Moon.

Finally, Chapter 5 provides the most fundamental look at impact vaporization, utilizing a recently upgraded high-speed spectroscopic system to track impact-generated vapor plumes in time and space. Ultra-fast spectroscopic methods non-intrusively characterize compositions and temperatures within transient, rapidly evolving impact plumes. Multiple fields of view (six total) within each impact-generated plume are targeted by the spectrometers, allowing vapor cooling sequences to be resolved. Transitions in time and space between dominant molecular emissions, vapor phase temperatures, and blackbody temperatures are visualized with concurrent operation of 1

million frames per second cameras. Both target-derived vapor plumes and projectile-derived plumes are analyzed. Ultimately, observations from this study will be used to refine numerical and analytical approaches to impact vaporization under different conditions.

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Chapter 1. Painting Mercury Black with Cometary Carbon

Megan Bruck Syal and Peter H. Schultz

Department of Geological Sciences, Brown University

Providence, RI 02912

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Abstract

Mercury is darker than the Moon, yet its iron abundance is lower. As ferrous iron-bearing minerals and space weather-generated submicroscopic metallic iron are the primary known darkening materials at airless bodies, an unknown darkening agent is inferred to be globally distributed at Mercury. Although Mercury's intense space weathering environment drives additional darkening from large (> 40 nm) submicroscopic metallic iron particles, the addition of a mystery darkening agent is still necessary to reconcile the planet's overall low reflectance with its low iron abundance. Here we show that cometary carbon, delivered primarily by micrometeorites, provides a mechanism to darken Mercury's surface without violating remotely determined constraints on iron content. We find that delivery of carbon per unit surface area at Mercury is enhanced by a factor of ~ 230 -460, relative to the Moon. Spectroscopic analysis of products generated during hypervelocity experiments demonstrates how the presence of carbon effectively darkens and weakens spectral features, consistent with remote sensing observations of Mercury. Additionally, a significant fraction of delivered carbon may be in the form of nanodiamonds, which are generated by hypervelocity impacts. Carbon delivery by cometary dust not only provides an explanation for Mercury's globally low reflectance but also represents a novel method for darkening planetary surfaces.

1.1 Introduction

Efforts to characterize the composition of Mercury’s surface over the past several decades, including both Earth-based telescopic observations and higher spatial resolution data collected by the Mariner 10 spacecraft in 1974-75, revealed that the planet’s surface is described as dark and featureless in visible to near-infrared reflectance spectra [McCord and Adams, 1972; Hapke, 1977; Vilas, 1988; Rava and Hapke, 1987; Denevi and Robinson, 2008]. Multispectral data acquired by the MErcury Surface, Space ENvironment, GEochemistry, and Ranging (MESSENGER) spacecraft recently reconfirmed the absence of a 1-micron absorption feature at Mercury [Robinson et al., 2008; McClintock et al., 2008; Blewett et al., 2009]. The 1-micron absorption band, which is indicative of ferrous iron content in silicates, is commonly observed in near-IR spectra of the Moon, the airless body to which Mercury is most often compared. The absence of this band implies an upper limit of 2-3% FeO abundance at the surface of Mercury [Vilas, 1988; Blewett et al., 2009; Riner et al., 2010]. Such low ferrous iron abundances, however, are difficult to reconcile with Mercury’s low reflectance (relative to the Moon), as FeO functions as an important darkening agent in crustal materials [Rava and Hapke, 1987; Denevi and Robinson, 2008; Riner et al., 2010]. Hence, an unknown, global darkening agent is inferred to be distributed throughout Mercury’s crust [Rava and Hapke, 1987; Denevi et al., 2009; Blewett et al., 2009].

Initially, Fe- and Ti-bearing oxides were suggested as candidate opaque phase(s) capable of sufficiently darkening Mercury [Denevi et al., 2009; Lawrence et al., 2010]. However, the low elemental iron content ($\sim 1.5\%$ Fe) measured by both the MESSENGER X-ray Spectrometer (XRS) [Nittler et al., 2011] and Gamma-Ray Spectrometer (GRS) [Evans et al., 2012] has ruled out this hypothesis [Riner et al., 2009].

Differing space weathering environments at Mercury and the Moon present an added layer of complexity in interpreting the reflectance spectra of these airless bodies.

“Space weathering” is comprised of at least two distinct processes: impact-generated vapor deposition from meteoritic bombardment and sputtering by charged particles from the solar wind. Early analyses of lunar soil particles revealed that thin (60 - 200 nm) amorphous rims surrounded many soil grains [Dran et al., 1970; Bibring et al., 1972], which are produced by some combination of impact vapor condensation and solar wind sputtering. There has been considerable debate in the literature on the extent of isotopic fractionation in these rims [Epstein and Taylor, 1972; Clayton et al., 1974; Switkowski et al., 1977; Russell et al., 1978; Esat and Taylor, 1992] and on the relative contributions of impact vapor deposition and solar wind sputtering to these thin coatings [Keller and McKay, 1993; Bernatowicz et al., 1994; Keller and McKay, 1994; Christoffersen et al., 1996; Keller and McKay, 1997]. However, the spectral effects of space weathering at the Moon, including darkening, reddening of spectra, and weakening of absorption band strengths, have been well constrained by a combination of laboratory, remote sensing, and theoretical work [McCord and Adams, 1973; Hapke et al., 1975; Fischer and Pieters, 1994]. The accumulation of submicroscopic metallic iron (SMFe) within the rims of vapor-coated grains and inside of agglutinates is now known to account for these major spectral changes [Pieters et al., 1993; Britt and Pieters, 1994; Pieters et al., 2000; Hapke, 2001; Noble et al., 2001].

Although mercurian soil samples are not available to assist in differentiating space weathering effects at the Moon and Mercury, experimental study of synthesized analog soils demonstrates that the spectral character of Mercury’s surface is consistent with larger (> 40 nm) SMFe particles [Noble et al., 2007]. Unlike smaller (15 - 25 nm) SMFe particles, which dominate space-weathered lunar soils and produce both darkening and reddening effects in the near-IR, larger SMFe particles darken samples without reddening their spectra [Britt and Pieters, 1994]. These larger iron grains are often referred to as Britt-Pieters (BP) particles and typically reside in agglutinates [Noble

et al., 2007]. A more significant spectral contribution from BP particles at Mercury is consistent with enhanced vapor and melt production, relative to the Moon, due to greater impact velocities and a higher micrometeorite flux at Mercury [Cintala, 1992; Hapke, 2001; Noble and Pieters, 2003; Noble et al., 2007]. Another process that may contribute to BP particle formation at Mercury is Ostwald Ripening, whereby iron particles increase in size as a consequence of prolonged exposure to the planet’s extreme surface temperatures [Noble and Pieters, 2003].

The availability of high spectral- and spatial-resolution visible to near-IR data from both the MASCS and MDIS instruments on-board the MESSENGER spacecraft has motivated additional work to understand space weathering effects at Mercury. The radiative transfer model developed by Hapke [2001] was extended by Lucey and Riner [2011] to account for the effects of the larger BP particles believed to dominate space weathered materials at Mercury. This model successfully reproduced the darkening without reddening effects described by Noble et al. [2007] and, using spectral data from MESSENGER, estimated the abundance of SMFe at Mercury to be $\sim 3.5\%$ (compared to $\sim 0.5\%$ at the Moon). This analysis was then refined by Riner and Lucey [2012] to place bounds on allowable SMFe abundances as a function of an unknown, opaque mineral’s abundance. While Riner and Lucey [2012] demonstrated that differing amounts of BP particles can explain the range of albedos observed within various terrain types at Mercury, the addition of a mystery darkening agent is still necessary to reconcile the planet’s overall low albedo with MESSENGER’s XRS and GRS-derived constraints on total iron abundance.

Here we propose that carbon, delivered by comets and cometary dust, is the unknown darkening agent at Mercury. Carbon has been briefly considered as a candidate opaque phase at Mercury in a few previous studies [Denevi et al., 2009; Blewett et al., 2009]. However, the importance of carbon at the surface was ruled out, due its efficient sequestration to the core early in Mercury’s history [Hillgren et al., 2000] and

its propensity to be volatilized during volcanism [Denevi et al., 2009]. An exogenous origin for the carbon, in which it is delivered to Mercury’s surface on relatively short time scales by comets and cometary debris, overcomes these concerns.

1.2 Carbon Transport to Mercury

1.2 Cometary Flux

Mercury’s proximity to the sun (semi-major axis of 0.387 AU) places the planet in the path of many more comets than the Earth-Moon system, as the number of comets per unit area is inversely proportional to heliocentric distance [Oort, 1950; Hartmann et al., 1981]. In addition, the large thermal stresses experienced by comets in the near-sun region [Sekanina and Chodas, 2012] raises the probability of a near-Mercury fragmentation for these icy, organic-rich bodies, which contributes an additional flux of cometary material in the form of micrometeorite bombardment. Previous studies have investigated the delivery of cometary material at Mercury in order to quantify the delivery of water to the planet’s surface [Killen et al., 1997; Moses et al., 1999], but these estimates considered only short-period (or ecliptic) comets. Including the contributions of long-period comets is estimated to increase the cometary impact flux at Mercury by at least a factor of 20 [Hartmann et al., 1981]. This effect is especially apparent in light of the remarkable number of sun-grazing (and Mercury-crossing) comets discovered by the SOlar and Heliospheric Observatory (SOHO) since 1996 [Sekanina, 2002; Sekanina and Chodas, 2012; Schrijver et al., 2012]: roughly one new comet every three days.

A Monte Carlo approach applied to short-period comet impacts by Moses et al. [1999] estimated that a total of 1.4×10^{18} g of cometary material has been delivered to Mercury over the past 3.5 billion years. This estimate invoked a comet nucleus size distribution ranging from 0.5 to 50 km in diameter and used fairly conservative estimates for the percentage of impacting material retained by the planet: 1.5% for

Halley-type comets ($v_{\text{impact}} = 65$ km/s) and 7.5% for Jupiter Family Comets ($v_{\text{impact}} = 39$ km/s). While these retention rates were estimated using an analytical approach, experiments reveal that the internal energy of vapor plumes generated by oblique impacts may differ from model predictions [Schultz, 1996]. The significant abundance of organic material within comets, estimated to be $\sim 25\%$ [Chyba et al., 1990], implies that short-period comets have transported $\sim 1.0 \times 10^{14}$ g of carbon-rich matter to the planet’s surface over the past million years.

In order to determine the amount of carbon delivered by long-period comets, we scale the total short-period comet impact flux reported in Moses et al. [1999] by a factor of 20, based on the dynamical arguments made by Hartmann et al. [1981]. This yields $\sim 8.4 \times 10^{20}$ g of long-period cometary material impacting the surface over the last 3.5 billion years. Since the long-period comet population is described by a larger root mean square impact velocity ($v_{\text{impact}} = 87$ km/s) [Hartmann et al., 1981], a more modest retention rate (1.0%) is applied to this source. The end result is an additional contribution of $\sim 6.0 \times 10^{14}$ g from long-period comets to the carbon budget at Mercury over the past million years.

1.2 *Micrometeorite Flux*

Micrometeorites in the form of impacting interplanetary dust particles (IDPs) constitute a major source of exogenous carbon at Mercury. The spatial distribution of IDPs between 0.3 and 1 AU, initially constrained by Pioneer 10/11 and Helios 1/2 observations, varies as $n(r) \sim r^{-1.3}$ [Leinert et al., 1981]. This increase in spatial density of IDPs with decreasing heliocentric distance is associated with an increased micrometeorite flux at Mercury, relative to the Moon. This point was detailed by Cintala [1992], who estimated a factor of 5.5 difference in IDP flux between the two airless bodies. More recent numerical work to estimate the micrometeorite flux at Mercury incorporates Poynting-Robertson drag, solar wind drag, and the terrestrial

planets' gravity to determine the orbital evolution of dust grains (radius $< 100 \mu\text{m}$) in the inner solar system [Borin et al., 2009]. Calibrating this model with experimentally determined measurements of the IDP flux at the Earth [Love and Brownlee, 1993], Borin et al. [2009] find that the flux at Mercury is ~ 170 times greater than originally estimated by Cintala [1992]: $2.38 \times 10^{-14} \text{ g/cm}^2\text{s}$. The terrestrial flux has also been revised upward since the Cintala [1992] study and is used by Borin et al. [2009] to calibrate their Mercury flux values. Recent estimates for the lunar dust flux range from $5.17 \times 10^{-17} \text{ g/cm}^2\text{s}$ [Bruno et al., 2006] to $1.01 \times 10^{-16} \text{ g/cm}^2\text{s}$ [Thomas-Keprta et al., 2014], yielding a factor of $\sim 230 - 460$ enhancement in micrometeorite flux at Mercury, relative to the Moon.

Although micrometeorite origins are traceable to both asteroids and comets, multiple lines of evidence suggest that the IDP population inside of 1 AU is dominated by cometary material. In order to maintain the observed density distribution of particles, Grün et al. [1985] calculated that a source must replenish particles in the inner solar system on very short times scales, and the source strength must be radially dependent (most of the material is deposited $< 0.1 \text{ AU}$). Comets, which shed material at increasing rates with decreasing heliocentric distance, fit this source profile. Recently, two independent dynamical studies of cometary and asteroidal contributions to the near-Earth micrometeoroid complex have found that $> 90\%$ of IDPs are cometary in origin [Wiegert et al., 2009; Nesvorný et al., 2010]. As noted by Nesvorný et al. [2010], this finding is consistent with the observation that most antarctic micrometeorites are described by primitive, carbonaceous compositions typical of cometary dust [Dobrică et al., 2009; Duprat et al., 2010]. The carbon content of cometary dust is roughly an order of magnitude greater than carbonaceous chondrites [Jessberger et al., 1988]. Hence, the relative increase in IDP flux between the Moon and Mercury should result in the transport of $\sim 230-460$ times more carbon-rich material per unit surface area at Mercury.

Retention of impacting IDP material is substantially more efficient than short- and long-period comets, as micrometeoroids possess lower mean impact velocities. The dynamical evolution of these particles, after their release from cometary nuclei, places them in low-inclination, low-eccentricity orbits, from which their final impact velocities approach nominal asteroidal dust values [Liou and Zook, 1996]. The average impact velocity calculated from numerically integrating dust particle orbits in Borin et al. [2009] was 16.81 km/s, lower than the 20.50 km/s determined analytically by Cintala [1992]. At Mercury, velocities less than 20 km/s result in 100% impactor retention for 90 degree impacts, as reported in Moses et al. [1999]. As the probability distribution for bolide impact angles scales with $\sin 2\theta$ (peaking at 45°), near-vertical impacts are relatively rare [Gilbert, 1893; Shoemaker, 1962; Gault and Wedekind, 1978]. In order to more fully capture the actual delivery efficiency of micrometeorites at Mercury, we conduct numerical simulations of 20 km/s impacts at a range of impact angles and calculate the fraction of impactor remaining below Mercury’s escape velocity (4.25 km/s) for each case.

1.3 Retention of Carbon: Numerical Assessment

The three-dimensional shock physics code CTH [McGlaun et al., 1990] has been previously utilized to estimate impact delivery of organics at Europa [Pierazzo and Chyba, 2002] and Mars [Pierazzo and Chyba, 2003], with the latter work incorporating impact angle effects. These studies used tracer particles, distributed through the impactor, to determine the fraction of impacting material that remains below a planet’s escape velocity. Here we take advantage of a simpler method, afforded by newer versions of CTH: the total mass of impactor remaining below the escape velocity of Mercury (4.25 km/s) is output every $0.1 \mu\text{s}$ through the use of a data filter. For consistency with previous work and additional information on the evolving velocity distributions of the impactor components, we also include 555 tracer particles, distributed throughout

one hemisphere ($y > 0$) of the projectile (the problem possesses bilateral symmetry around the y-axis).

The impactor and target material were modeled using the ANEOS equation of state for dunite [Thompson and Lauson, 1972]. This choice was informed by past work to document impact-generated phase changes associated with the ANEOS relations for dunite [Pierazzo et al., 1997; Barr and Citron, 2011] and the fact that the reference density of dunite, $\rho = 3.32 \text{ g/cm}^3$, is comparable to the materials modeled in this study. Spherical impactors with radii of 0.25 cm and velocity magnitudes of 20 km/s were used in fully 3-D simulations of impacts at 15° , 30° , 45° , 60° , and 90° from the horizontal. In order to focus the calculations on the regions of greatest physical interest in the problem domain, Adaptive Mesh Refinement (AMR) was used in each simulation. The most highly-resolved portions of the problem corresponded to a resolution of 25 cells per projectile radius (cpr).

The percentage of impactor mass that remains below escape velocity, as a function of time, is shown in Figure 1.1a. As impact angle decreases, the duration of the penetration stage (time taken to fully couple the projectile’s energy to the target) increases, so that the final fraction of material retained by the planet is reached at progressively later times. For each case, the retained fraction converges to a final estimate within the first 4 μs . Final mass fractions delivered at each impact angle, along with the impact angle probability distribution function, are shown in Figure 1.1b. The total fraction of 555 tracer particles remaining below escape velocity is also plotted; our mass filtering algorithm is found to be in good agreement with the tracer particle method for angles 30° and greater. The highly asymmetric shock conditions within the projectile for the 15° case make the problem more sensitive to resolution discrepancies between the two methods. For 15° , the smaller fraction of delivered projectile calculated from the mass filtering method is attributed to the enhanced resolution at the edges of the projectile, which typically attain the highest

post-impact velocities.

The results illustrated in Figure 1.1b can be broken into three regimes: (1) a gradually sloped decrease in delivery efficiency between 90° and 60° and (2) a more steeply sloped decrease in efficiency from 60° to 30° , and (3) an increase in efficiency from 30° to 0° . These results, which are consistent with the findings reported in Pierazzo and Chyba [2003], can be integrated with the probability distribution function to calculate the overall efficiency of impact delivery at Mercury:

$$F = \int_0^{\pi/2} \sin(2\theta) f(\theta) d\theta \quad (1.1)$$

Where F is the total fraction of impacting material retained, $f(\theta)$ is the linear, least squares fit for fraction retained at a given angle interval, and θ is impact angle (in radians).

Impacts between 60° and 90° comprise 25% of all impacts; integration of this segment from Equation 1 yields a fraction of 0.2436 or a 97% retention efficiency in this segment. Half of all impacts occur in the 30° - 60° interval; integration of this component yields 0.3917 or 78% retention efficiency in this regime. The final quartile of impacts occurs between 0° and 30° ; applying the linear fit between 15° and 30° and integrating across this segment produces a value of 0.1964 or 79% retention over this range. The total efficiency, across the entire impact angle distribution, is then 83%. This should be viewed as a conservative estimate, as these calculations were for 20 km/s impactors and the mean impact velocity for IPD's at Mercury is 16.81 km/s [Borin et al., 2009].

Lower delivery efficiencies at increasingly oblique impact angles (down to 30°) result from enhanced partitioning of energy to the impactor, relative to the target. This result is contrary to the analytical approach commonly utilized to extrapolate 90° impacts to non-normal angles, wherein the energy of the vapor plume, ϵ_0 , is taken as proportional to $\sin^2(\theta)$ and used to calculate the total retained mass [Vickery and

Melosh, 1990]. At smaller impact angles, the analytical model predicts greater delivery efficiencies, rather than the lower efficiencies calculated here. Additionally, retention rates depend on impact angle in a complex manner. The increased partitioning of energy to the projectile at lower angles is, eventually (near 15°), counteracted by a combination of lower peak shock pressures (which vary as $\sin^2(\theta)$) and energy losses to ricochet [Schultz and Gault, 1990; Schultz, 1996], resulting in a reversal of the trend. Since fully 3-D hydrodynamical calculations are computationally expensive, it is often desirable to employ 2-D impact models (restricting the impact angle to 90°) and then generalize the results to other impact angle conditions. As such, this problem merits additional experimental and numerical work, which we will address in a future contribution.

The estimated delivery rates for carbon associated with IDPs, short-period comets, and long-period comets are summarized in Table 1.1. Globally, this is equivalent to the deposition of ~ 0.16 g/cm² of carbon over the past million years, significant enough to rapidly darken the upper centimeters of the regolith. IDPs clearly provide the bulk of exogenous carbon at Mercury. Therefore, any compositional differences arising from cometary carbon will be dominated by the factor of 230-460 difference in micrometeorite flux between the Moon and Mercury.

1.4 Previous Evidence for Carbon Darkening

Darkening effects from disassociated organics within impact-generated vapor plumes of cometary composition have been observed at both planetary and laboratory scales. In 1994, the natural impact of Comet Shoemaker-Levy 9 (SL-9) into Jupiter afforded a unique, planetary-scale opportunity to view the products associated with shocked and vaporized cometary material. After initial vapor plume formation and subsequent cooling, the impact sites of cometary fragments from SL-9 were associated with the prolonged presence of dark blemishes [Hammel et al., 1995]. The optical constants of

these blemishes were well-matched by organic aerosols, which were likely synthesized from disassociated cometary products during the high-temperature and high-pressure impact process [Wilson and Sagan, 1997].

In 2005, the Deep Impact Mission successfully observed the impact plume generated by a planetary-scale cratering experiment into the nucleus of Comet 9P/Tempel 1 [A’Hearn et al., 2005; Schultz et al., 2007]. Evidence for the condensation of dark organic material out of the vapor plume was provided by mid-infrared telescopic observations of the plume’s evolving continuum component [Sugita et al., 2005]. The addition of complex organics to target material in laboratory-scale impact experiments was also able to replicate a late stage increase in brightness or “blooming” seen at Deep Impact, attributable to the condensation of carbon-rich droplets [Schultz et al., 2007]. These observed organics are interpreted as products of vaporized and disassociated cometary materials. As comets are comprised of 20-25% organic matter [Chyba et al., 1990], there is ample material available for the formation of dark, carbon-bearing compounds.

1.5 Impact Experiments with Organics

1.5 *Experimental Method*

In order to fully assess whether carbonaceous cometary material transported to Mercury can function as an effective darkening agent, hypervelocity ($v = 5.68 \pm 0.01$ km/s) impact experiments were carried out at the NASA Ames Vertical Gun Range (AVGR) in Moffett Field, CA. Though the AVGR facility is capable of utilizing a range of impact angles, spanning 15° to 90° , all results reported here are for vertical incidence impacts. This geometry simplifies the collection of impact-generated agglutinates, which would be scattered downrange in the case of an oblique impact. Near-vacuum conditions, $P = 3.3 \times 10^{-4}$ atm, were maintained within the impact chamber during the experiments. Pyrex projectiles 0.635 cm in diameter were used,

as the melted glass quickly quenches and traps target materials into agglutinates for later analysis.

The lunar simulant JSC-1A, a granular mix of minerals developed to approximate the composition of lunar mare soil (see McKay et al. [1994] and Hill et al. [2007]), was chosen as a proxy material for the regolith of Mercury. Although the surface mineralogies of the Moon and Mercury likely differ, in the absence of available soil samples from Mercury, JSC-1A is a useful analog for exploring the effects of carbon exposure at airless bodies. Isolating the role of organics in impact processing of the soil was accomplished through the use of two different target conditions: (1) JSC-1A mixed in a 1:1 ratio with Ottawa quartz sand; (2) JSC-1A mixed in a 1:1 ratio with organics (i.e., sugar).

1.5 Impact Products

Impacts into each target type (with and without an organic component) generated a sufficient volume of agglutinates for collection and subsequent analysis. Optical and scanning electron microscopy reveals that the inclusion of organics produces agglutinates that are significantly visually darker (see Figures 1.2-1.4). The textures of each agglutinate type can be described as globular, though, at smaller scales, the sample darkened by organics exhibits a more fibrous texture. The texture, color, and opacity differences between these two very distinct varieties of impact products were consistent throughout the range of agglutinate particles retrieved from the experiments.

1.5 Visible/Near-infrared Reflectance Spectroscopy

Bidirectional visible/near-IR reflectance spectra of the impact products were obtained in RELAB at Brown University, using standard incidence and emission angles: $i = 30^\circ$; $e = 0^\circ$. Figure 1.5 depicts the resulting spectral measurements, which were performed on the agglutinate samples pictured in Figure 1.4. As expected from visual

inspection, the agglutinate sample that was generated in the presence of organics possessed a significantly lower albedo than the organics-free sample. While the brighter agglutinates produced by the JSC-1A and sand mixture possessed a relatively strong 1-micron absorption band, this ferrous iron signature was completely destroyed when organics were substituted for the sand, despite equal elemental abundances of iron in each experiment. In addition, the darker spectrum associated with the inclusion of organics was blue-sloped, in contrast with the organics-free case.

1.6 Discussion

Although evidence for the disassociation and condensation of organics into darker material has been observed during past impact events [Hammel et al., 1995; Wilson and Sagan, 1997; Schultz et al., 2007; Sugita et al., 2005], this study is the first to document the lasting effects of this process on the spectral properties of the target. The introduction of organics is found to dramatically darken impact-generated agglutinates, thereby implicating carbon as an effective darkening phase at planetary surfaces. Given current constraints on Mercury’s low elemental iron abundance ($\sim 1.5\%$ Fe), the overall low albedo of the planet cannot be explained by the presence of space weather-related SMFe alone. Carbon, which would successfully evade available remote sensing detection methods at Mercury, provides a new explanation for Mercury’s low albedo, as its delivery to Mercury via cometary materials is enhanced by a factor of ~ 230 -460 over the delivery rate at the Moon. The observation that fresh, immature units on Mercury are $>30\%$ darker than comparable lunar units [Denevi and Robinson, 2008] suggests that the opaque phase (here, proposed to be carbon) is well mixed within the crustal material and/or is delivered on very short time scales. As micrometeorite bombardment will darken the surface through a combination of carbon delivery and BP particle production in agglutinates, these processes can be viewed as working in concert to lower the albedo of Mercury, relative to the lunar

surface.

In addition to darkening effects, the presence of carbon in agglutinates produces relatively blue-sloped spectra. As seen in Figure 1.5, the red-sloped spectrum of the organics-free agglutinate contrasts with the flattened spectrum associated with the inclusion of organics. Direct comparison to reflectance spectra of Mercury obtained by MESSENGER illustrates how carbon-driven spectral darkening and flattening can account for variation between different units (Fig. 1.5b). On the basis of spectral data, the surface of Mercury is often segmented into three terrain types: high reflectance smooth plains (HRP), intermediate terrain (IT), and low reflectance material (LRM); much of the LRM is also described by bluer spectral slopes [Robinson et al., 2008; Blewett et al., 2009; Denevi et al., 2009; Ernst et al., 2010].

If carbon is the opaque phase responsible for the darker and bluer characteristics of the LRM, the distribution of LRM may provide insight into: (1) how the micrometeorite flux has varied over time, or (2) localized carbon deposition by recent, large cometary impacts. The interpretation that LRM, due to its association with impact ejecta, originates at depth [Denevi et al., 2009; Ernst et al., 2010], suggests that carbon delivery rates may have been greater in the past. While the time-averaged flux of carbon is greater for comet-derived IDPs than full-sized comets, a discrete comet impact event in the recent past could leave discernible, local traces of carbon deposition. LRM that appears to be more recently emplaced and/or exhibits cross-cutting relationships with nearby ejecta blankets may be attributed to carbon-rich materials condensing out of an impact vapor plume.

The experimental results also suggest that carbon may conceal ferrous iron content by obscuring, through the deposition of an opaque veneer, the 1-micron absorption band. Although the total abundance of iron is equal in both cases, the 1-micron band is absent from the spectrum of agglutinates produced with organics, while it remains clearly visible when no organics are included. In the case of most remote

sensing techniques, which sample only the uppermost layer of surface composition ($<100\ \mu\text{m}$), this carbon-induced obscuration of true iron content could be effective. However, MESSENGER’s GRS penetrates down to a depth of tens of centimeters and finds similarly low iron abundances [Evans et al., 2012]. Hence, the elemental iron content near to Mercury’s surface is truly low and not merely obscured by a veneer of carbon.

As the Moon is also exposed to impacting cometary debris (though to a lesser extent than Mercury), it is reasonable to expect geochemical signatures of carbon deposition on lunar soil grains. Indeed, soil samples from both the Apollo and Luna programs record contributions from exogenous carbon. Total carbon content in the Apollo soils was found to increase with prolonged exposure to micrometeorite bombardment, as evidenced by a correlation between surface carbon concentration and impact processing indicators (i.e., rare gas abundances and amorphous particle coatings) [Cadogan et al., 1972; DesMarais et al., 1973]. The highest concentrations of carbon reside in agglutinates, which have undergone the most prolonged soil cycling. An apparently high rate of comminution- and aggregation-driven cycling of carbon (relative to the rate of extralunar carbon accumulation) distributes the carbon throughout the particle interiors [DesMarais et al., 1973]. In addition, Luna 16 soil grains were observed to be coated by carbon- and zinc-rich films (up to 60 at% and 4 at%, respectively) [Dikov et al., 1998]. The layering structure of these deposited films is consistent with condensation out of a carbon-rich impact vapor plume; analysis of the carbon indicates that it is in the form of nanodiamonds or graphite.

Impact disassociation of cometary organics is likely to generate carbon in various forms at Mercury. Nanodiamonds, which are produced through impact-driven chemical vapor deposition and shock alteration [Gilmour et al., 1992; Daulton et al., 1996], are often associated with terrestrial impact craters [Masaitis, 1998]. At the surface of

Mercury, nanodiamonds may be particularly abundant, due to high rates of carbon delivery and high impact velocities. These conditions favor nanodiamond formation through: (1) efficient vaporization and subsequent deposition of carbon, and (2) rapid thermal quenching of shocked carbon.

1.7 Conclusions

Accounting for the low albedo of Mercury, given spectroscopic constraints on ferrous iron abundance at the planet, has challenged researchers for many years. Various compositional explanations have been invoked, including an array of opaque minerals, enhanced agglutinate production, and space weather-generated SMFe. While the unique space weathering environment of Mercury undoubtedly contributes to the dark and featureless character of its spectra, the distribution of an unknown, iron-free opaque material throughout the crust is required to sufficiently darken the surface.

Carbon is an abundant opaque phase within our solar system and is transported in significant quantities to Mercury. In particular, due to fundamental differences in the dynamical micrometeoroid environments at Mercury and the Moon, the delivery rate of cometary-derived carbon should be at least two orders of magnitude greater at Mercury. Impact experiments carried out to constrain the surface darkening potential of cometary organics demonstrate the efficacy of carbon-rich material for darkening planetary surfaces. Spectral effects including lower albedo, weakening of absorption bands, and bluer spectral slopes are consistent with the effects of the previously unknown darkening agent inferred to be present in the crust of Mercury. While direct detection of carbon at Mercury is not believed to be feasible with the currently available suite of remote sensing tools, prioritizing an investigation of carbon within the science objectives of future missions to Mercury could shed light on the role of this element at our innermost planet.

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Tables

Table 1.1: Carbon Delivery Rates at Mercury

Source ^a	Retention	Delivery Rate (g cm ⁻² Myr ⁻¹)
Micrometeorites ^b	83%	0.156
Short Period Comets ^c	1.5-7.5%	1.34×10^{-4}
Long Period Comets ^d	1.0%	8.02×10^{-4}
Total		~ 0.16

^aCarbon abundance of 25% assumed for each

^bfrom Borin et al. [2009]

^cfrom Moses et al. [1999]

^dfrom Moses et al. [1999] and Hartmann et al. [1981]

Figure Captions

Figure 1.1: (a) Percentage of impactor mass remaining below Mercury’s escape velocity (4.25 km/s) plotted as a function of time, for five different impact angle conditions ($v_{\text{impact}} = 20$ km/s). Enhanced partitioning of energy to the impactor at lower angles results in less efficient delivery down to 30° : 100% for 90° , 96% for 60° , 80% for 45° , and 59% for 30° . At 15° , the combination of lower peak shock pressures and kinetic energy losses to ricochet results in a reversal of this trend: 88% of the impactor is retained. (b) Final fraction of retained impactor mass as a function of impact angle, along with the impact angle probability distribution function: $P(\theta) = \sin(2\theta)$, which peaks at 45° . The delivered fraction function is split into three regimes: 60° - 90° , 30° - 60° (linear fit indicated by dotted blue line), and 0° - 30° . Integrating the delivered fraction with the probability distribution over the entire range of impact angles produces a total delivery efficiency of 83%. Results from the 555 tracer particles distributed throughout the projectile are indicated by light blue diamonds, illustrating the consistencies and differences between this previously-employed method and the new mass filtering algorithm.

Figure 1.2: Agglutinates generated by hypervelocity impact experiments were collected from the floors of craters. Impacts into a 1:1 mix of lunar simulant JSC-1A and quartz sand (a) produced relatively higher albedo impact products. Including carbon in the form of complex organics (1:1 mix of JSC-1A and sugar) produced visually darker impact products (b). Both agglutinate types have somewhat similar globular textures at the scale pictured here, though the carbon-rich case also possesses more fibrous areas, which are apparent at smaller scales (see Fig. 1.3).

Figure 1.3: Scanning electron microscope (SEM) images reveal some of the finer-scale

structure and texture variations within the agglutinates. Compared to the organics-free case (a), the agglutinates produced in the presence of carbon-enrichment (b) are described by a more fibrous texture.

Figure 1.4: Agglutinate samples measured by RELAB’s bidirectional, visible/near-IR reflectance spectrometer (see Fig. 1.5). a. JSC-1A and quartz sand agglutinate b. JSC-1A and complex organics agglutinate.

Figure 1.5: Visible/near-IR spectra of impact agglutinates and different terrain types at Mercury. (a) The presence of organics during agglutinate formation was associated with three major spectral effects: lower reflectance (< 0.05), weakening of the ferrous iron absorption band at 1000 nm, and bluer spectral slope. This process is driven by the disassociation of organics into dark, carbon-enriched compounds, which are entrained by impact agglutinates. (b) Spectra of representative terrain types at Mercury, obtained by MESSENGER’s MDIS instrument, are plotted for comparison with impact agglutinates: high reflectance smooth plains (HRP), intermediate terrain (IT), and low reflectance material (LRM). Trends associated with darker and bluer material, which require higher abundances of a mystery opaque phase, are consistent with the spectral effects of carbon.

Figures

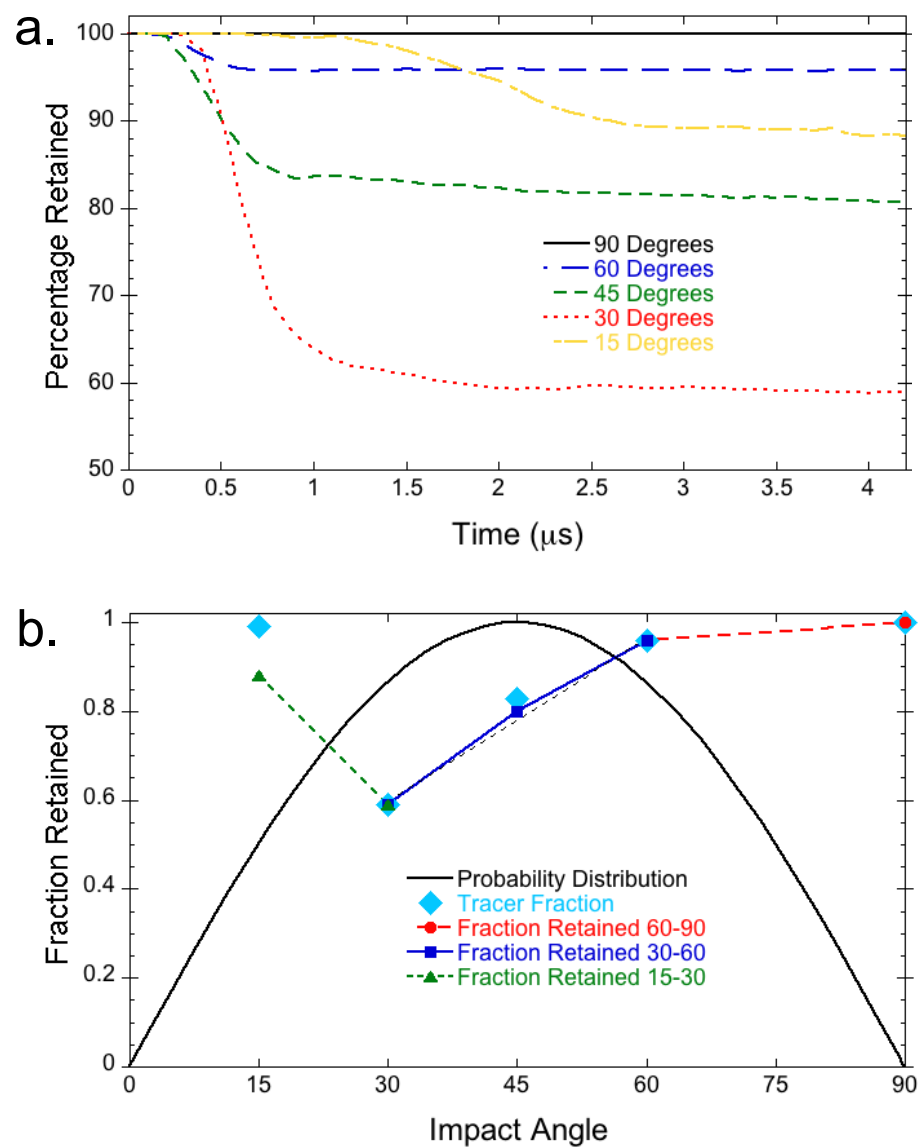


Figure 1.1

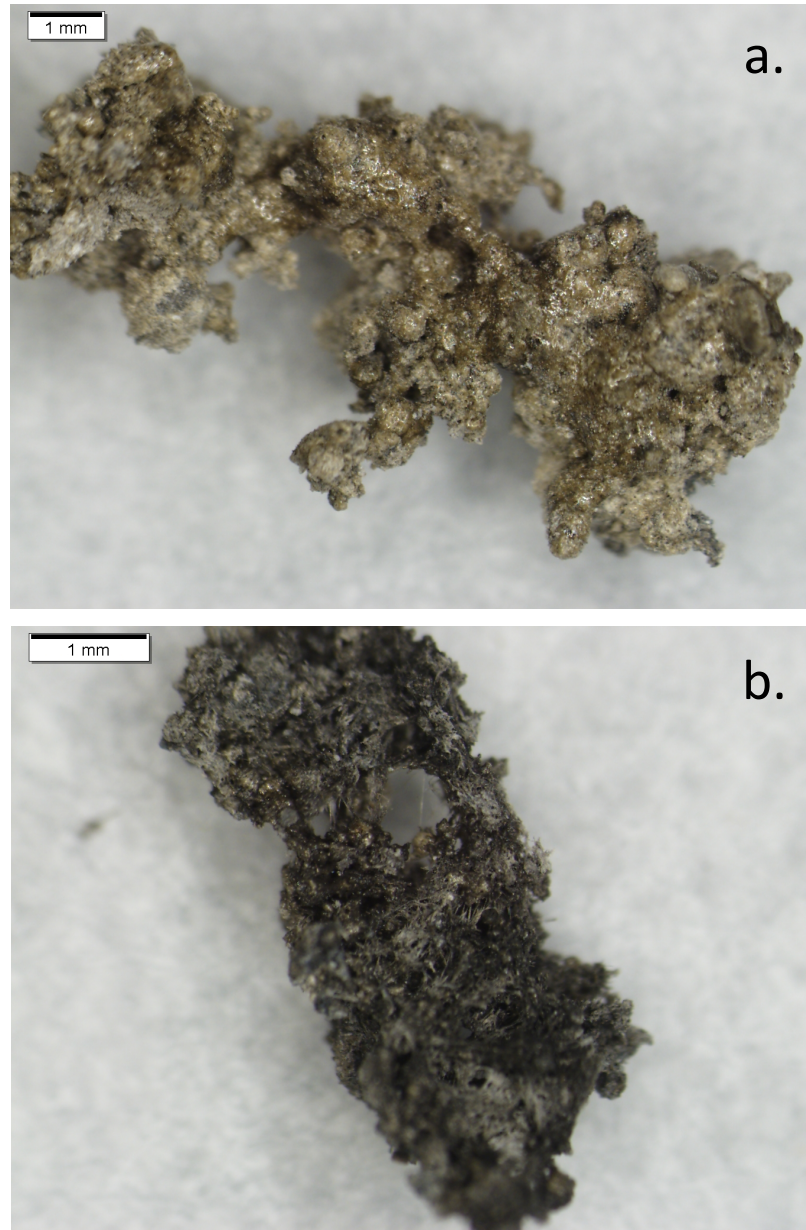


Figure 1.2

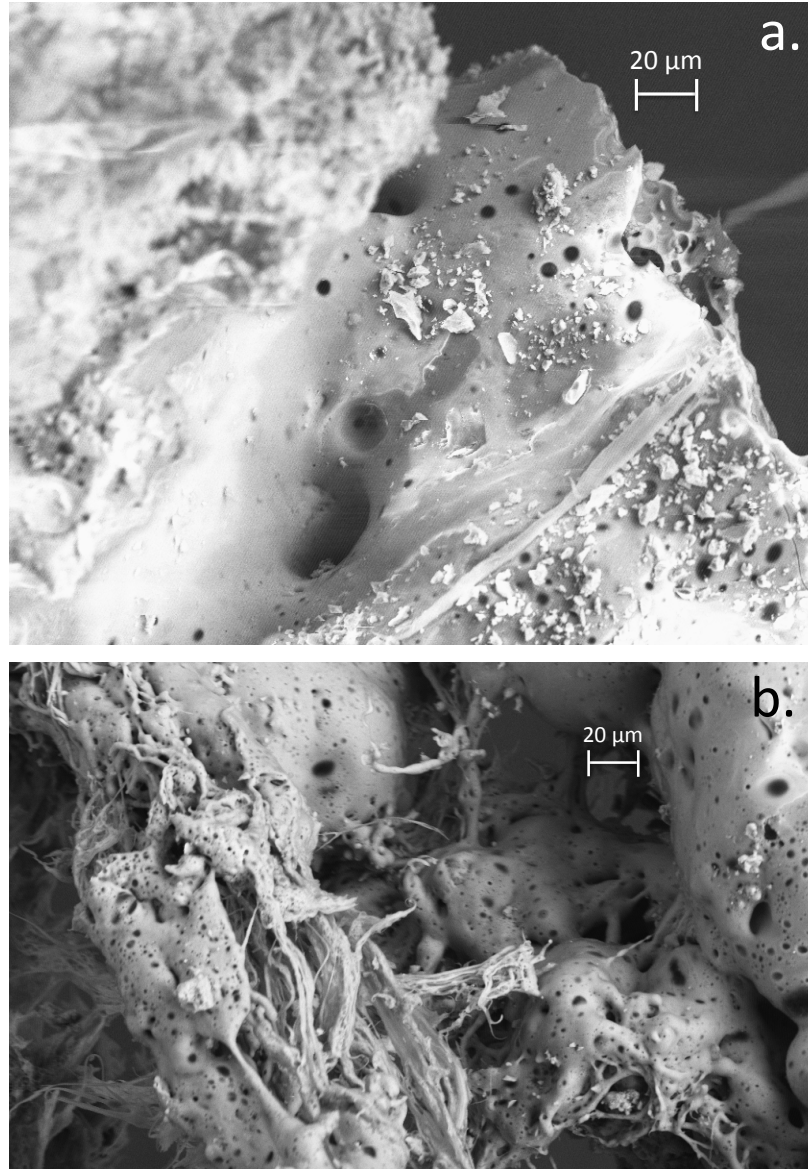


Figure 1.3

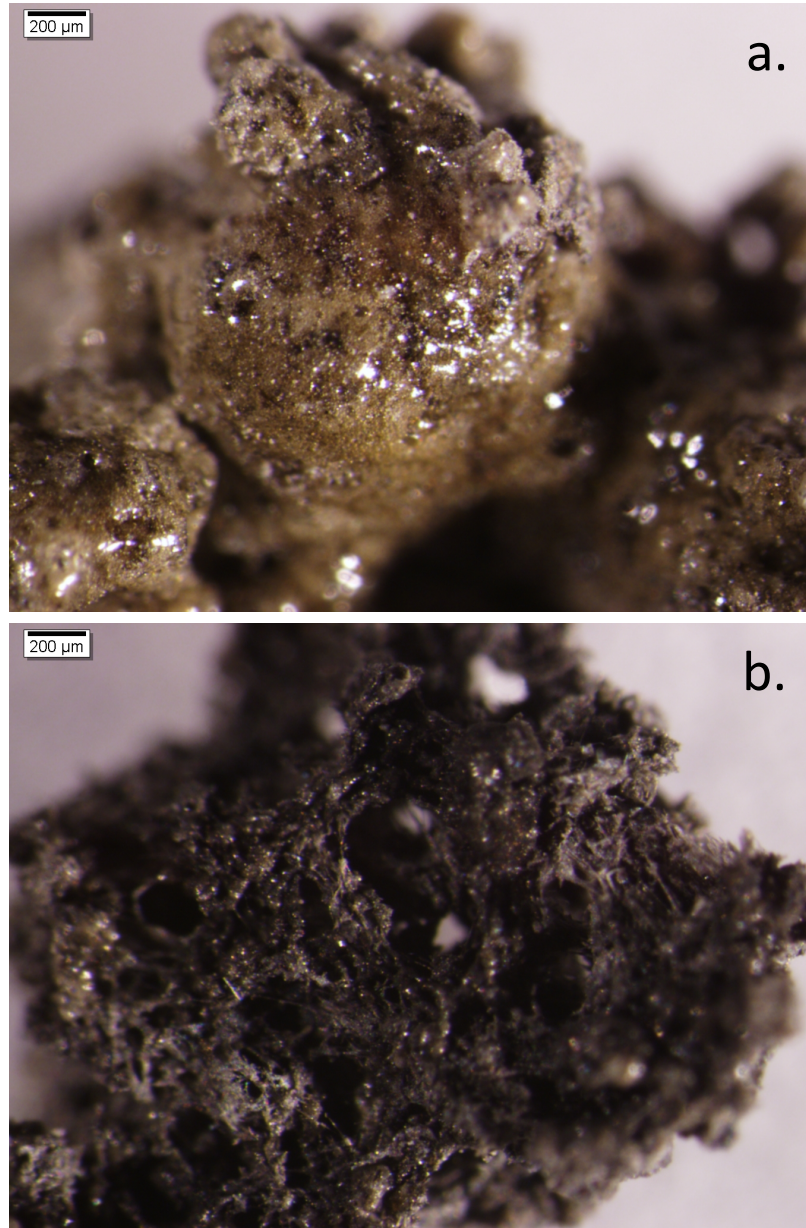


Figure 1.4

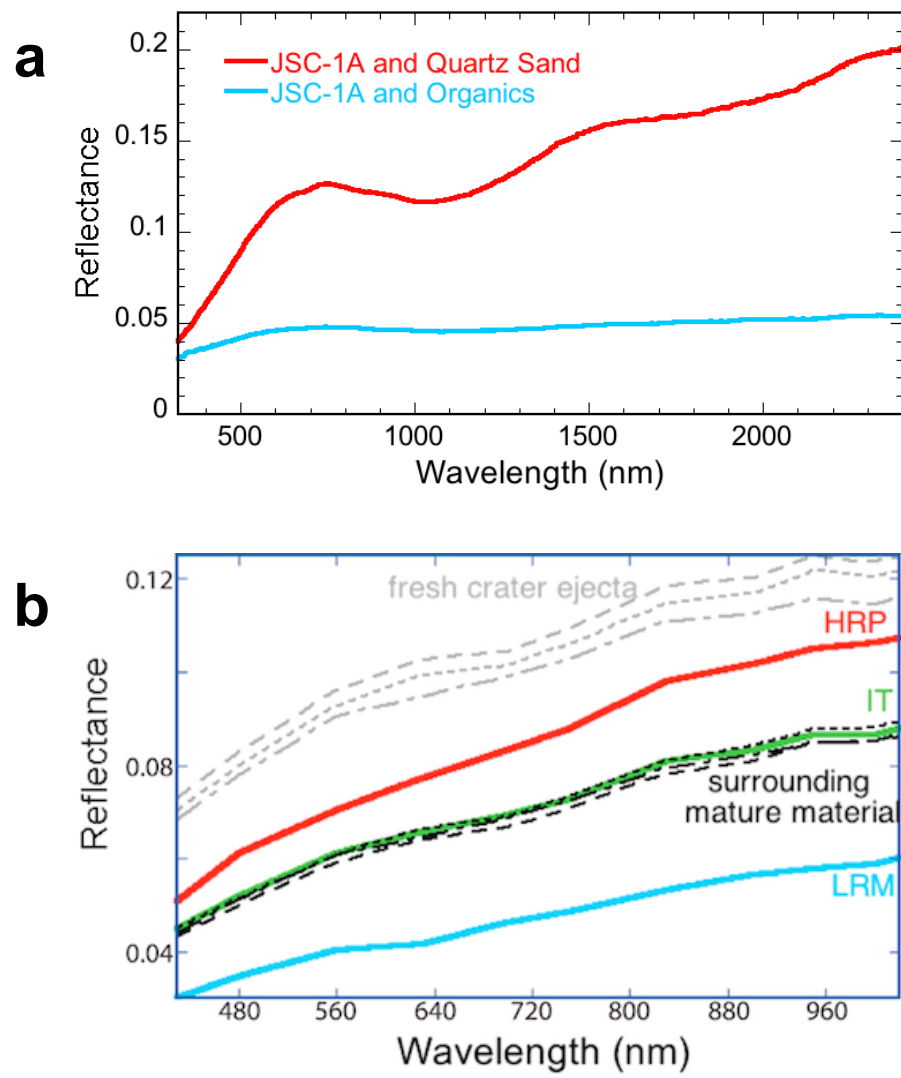


Figure 1.5

Chapter 2. Impact Vaporization of Water Ice: Experimental and Numerical Results

Megan Bruck Syal and Peter H. Schultz

Department of Geological Sciences, Brown University

Providence, RI 02912

Abstract

Experimental measurements of vapor plumes generated by hypervelocity impacts into porous and nonporous water ice serve as a benchmark for evaluating numerical approaches to impact vaporization. Using final release temperature as a phase change criterion, as opposed to critical shock pressures, yields numerical results for vapor generation that are more consistent with experimentally determined vapor masses. Applying the final release temperature metric to planetary-scale impacts at a range of impact angles, we find that porosity (ϕ) plays a greater role in enhancing vapor production than previously estimated: $\sim 40\%$ enhancement at $\phi=0.2$, and $\sim 80\%$ enhancement at $\phi=0.4$. In addition, increases in vapor generation are not independent of impact angle: porosity-driven increases in vaporization are exaggerated at oblique angles. These results suggest that material properties such as strength and porosity cannot always be decoupled from impact angle effects when assessing impact-generated phase changes. Furthermore, the significant enhancements in vaporization for porous materials, particularly from statistically dominant oblique impacts, provide a more efficient mechanism to increase rock fractions within porous bodies through collisional processing. Porosity, which is a key material property for many primordial bodies and some icy satellites, effectively lowers the critical velocities at which impact-generated phase changes will occur. This effect may be of particular consequence to the collisional histories of icy objects within the Kuiper belt, as impact velocities within that region of the solar system are relatively low (< 2 km/s).

2.1 Introduction

An accurate description of the impact-cratering process in ice or ice-enriched targets is essential for constraining the collisional evolution of icy bodies, including icy satellites [Chapman and McKinnon, 1986; Schenk, 2002; Barr and Canup, 2010], comets [Schultz et al., 2007; Schultz et al., 2013b], and Kuiper belt objects [Durda and Stern, 2000; Canup, 2005; Leinhardt et al., 2010]. Planetary surfaces with significant surficial ice, e.g., icy layered terrains on Mars [Wrobel et al., 2006; Senft and Stewart, 2008], also motivate experimental and numerical study of ice impacts. Many primitive bodies, including comets [Thomas et al., 2013a,b], Trojan asteroids [Marchis et al., 2006], and Kuiper belt objects [Brown, 2013], are known to contain porous ice, and geophysical modeling suggests that the regoliths of icy satellites such as Enceladus [Besserer et al., 2013] also may retain significant porosity. Porous materials introduce significant complications to impact cratering mechanics [Braslau, 1970; Schultz et al., 2005; Hermalyn and Schultz, 2011; Housen and Holsapple, 2003, 2012], as a large fraction of the impactor’s initial kinetic energy is consumed by the crushing of pore space. The additional PdV work done on the target during pore compaction works to enhance irreversible target heating, generally leading to increases in melt and vapor production [Zel’dovich and Raizer, 1966; Wünnemann et al., 2008], whereas attenuation of the shock wave decreases the maximal pressures experienced by target material.

The nature and volume of melt and vapor generated during impacts into ice is important to the understanding of many processes, including: the accretional histories of icy satellites [Ahrens and O’Keefe, 1985; McKinnon, 1989; Canup and Ward, 2009], impact-induced differentiation [Barr and Canup, 2010; Barr et al., 2010], and potential habitability conditions [Artemieva and Lunine, 2003; Abramov and Mojzsis, 2009]. Substantial progress on quantifying melt and vapor volumes for ices has been made

with the recent development and implementation of new water ice equations of state [Stewart and Ahrens, 2005; Senft and Stewart, 2008; Kraus et al., 2011]. Water ice phase changes for an extensive set of impact conditions were documented by Kraus et al. [2011] for planetary scale impacts, using the maximum pressure tracer method detailed in Pierazzo et al. [1997].

Experimental studies of impact vaporization, however, exhibit significant departures from numerical results. In particular, vaporization efficiency is enhanced with increasing obliquity in laboratory experiments [Schultz and Gault, 1990; Schultz, 1996], yet numerical simulations of oblique impacts [Pierazzo and Melosh, 2000; Kraus et al., 2011] yield lower vaporized masses than vertical incidence events. Because the peak shock pressure experienced during an impact scales with $\sin^2(\theta)$, it is unsurprising that the maximum pressure criterion for phase changes yields lower vapor volumes at shallower impact angles. The enhanced vaporization observed at oblique angles is postulated to be due to shear heating [Schultz, 1996], a process that is also necessary to explain the terrestrial occurrence of impact-generated pseudotachylites [Spray and Thompson, 1995]. Although difficult, non-intrusive measurements of vapor plume masses from impact experiments (through observations of the evolving plume radii) provide a critical benchmark for studying phase change criteria used in computational methods. Insights gleaned from vaporization phenomena may also have ramifications for impact-generated melt.

Numerical treatment of impact melt and vaporization generally neglects material strength, under the assumption of minimal heat dissipation during or after pressure unloading. While this assumption may be valid under certain conditions, e.g., very high-velocity events, it is highly desirable to reconcile experimental trends in vapor production with computational results. Accounting for the apparent discrepancy in impact angle effects is especially important, since oblique impacts are known to statistically dominate the bombardment history of planetary systems [Gilbert, 1893;

Shoemaker, 1962]. This study expands upon the work of Schultz [1996], which quantified the effects of incidence angle on impact-generated vapor for a range of target materials. Here we assess angle and porosity effects on water ice vaporization; these results are then directly compared to a range of hydrodynamical models in order to isolate the most important parameters and considerations for improved accounting of impact-generated phase changes. Using new knowledge from matching laboratory-scale experiments and numerical simulations, we then consider the implications for vaporization of porous materials at planetary scales.

2.2 Experimental Approach

Impact experiments were conducted at the NASA Ames Vertical Gun Range (AVGR), at incidence angles ranging from 30° to 90° and with $v \sim 5$ km/s. All experiments utilized spherical aluminum projectiles of $r = 0.3175$ cm; targets consisted of either porous ($\rho = 0.5$ g/cm³) or nonporous ($\rho = 0.92$ g/cm³) water-ice. Porous ice consisted of commercially-produced snow that was sifted into 59 cm-diameter target buckets (Fig. 2.1), while nonporous ice consisted of commercially-manufactured ice blocks (Fig. 2.2). See Table 2.1 for a complete listing of all experiments reported here.

The impact chamber at the AVGR allows for free expansion of impact-generated vapor plumes above the target surface. In order to obtain the necessary thermodynamic information from the plume’s expansion, a tenuous atmosphere ($P \sim 8$ mbar) was maintained within the chamber. The ice targets were stored at ~ 260 K until just prior to shooting. Two pairs of high frame-rate cameras (black and white Shimadzu, $8 \mu\text{s}$ time resolution, and color Phantoms, $24\text{--}67 \mu\text{s}$ time resolution) tracked vaporization and crater formation throughout all experiments.

2.3 Results

2.3 Impact Experiments

The style and extent of vaporization depends strongly upon impact angle. Qualitatively, enhancements in vapor volume and radiant intensity with increasing obliquity are evident in Fig. 2.1, which depicts impacts into porous ice at 90° and 30° . While both events produce hemispherically-expanding clouds of H_2O vapor, the oblique case imparts significant downrange momentum to the vaporized target. The earliest and hottest component of vaporization, termed the jetting phase (see Sugita and Schultz [1999] for a complete description), is suppressed in the porous ice, due to longer projectile penetration times for under dense targets [Schultz et al., 2007]. Jetting from impacts into crystalline targets consists of high-temperature plasma traveling downrange at up to ~ 3 times the initial impact velocity [Vickery, 1993; Sugita et al., 1998; Sugita and Schultz, 1999]. However, in the case of low-density, porous targets, e.g., Comet 9P/Tempel 1 [Schultz et al., 2007], the formation of any jetting phase occurs at a greater target depth, thereby complicating its expression at the surface. The subsurface expression of jetting in porous ice targets is documented in a series of quarter space experiments, described in Bruck Syal and Schultz [2014] (Chapter 3), that complement the vaporization results presented here. Suppression of jetting is visualized in Fig. 2.2, which compares vaporization in nonporous and porous ice targets for $\theta = 30^\circ$. The delayed emergence of vapor from the porous target produces a slower-propagating and more hemispherical plume, while low-angle jetting defines the leading edge of the vapor components from the nonporous target.

2.3 Measurement of Vapor Plume Energies and Masses

We utilize the method outlined in Schultz [1996] to determine the internal energies and masses of vapor plumes produced during impacts. At early times, the plume

rapidly expands in a non-isentropic manner, a phenomenon also observed at planetary scale during the Deep Impact cratering experiment [Schultz et al., 2007]. Once the expansion becomes isentropic, but before the plume is fully coupled to the tenuous atmosphere, the solution for vapor plume expansion in a vacuum [Zel'dovich and Raizer, 1966] applies:

$$u_{\infty} = \left(\frac{2E_v}{m_v} \right)^{1/2} \quad (2.1)$$

$$u_{\infty} = \left(\frac{\gamma - 1}{2\gamma} \right)^{1/2} u_{max} \quad (2.2)$$

Where u_{max} is the measurable velocity at the leading edge of the hemispherically-expanding plume, γ is the ratio of specific heats within the vapor cloud ($\gamma = 1.33$ for H_2O), u_{∞} is the mean expansion velocity within the plume, E_v is the vapor plume's internal energy, and m_v is the mass of the plume.

At later times ($> 100 \mu s$), the plume clearly decelerates as its energy couples to the surrounding atmosphere. Here, the similarity solution for an atmospheric blast wave [Taylor, 1950a,b] applies:

$$R = S(\gamma) t^{2/5} E_v^{1/5} \rho_0^{-1/5} \quad (2.3)$$

Where R is the plume radius, $S(\gamma)$ is an empirically-determined constant, t is time, and ρ_0 is atmospheric density. Linearizing the relation between R and t by plotting on a log-log scale allows a line to be fit to the late-stage plume radius as a function of time. The intercept of this line can then be solved for E_v . This result is used in combination with the u_{max} value observed during earlier, isentropic expansion, to calculate the total vaporized mass, m_v . An example plume radius evolution with time is shown in Fig. 2.3; note the distinct atmospheric deceleration of vapor expansion at later times.

Results for porous and solid ice targets at a range of impact angles are listed in

Table 2.1 and plotted (along with computational results, described in next section) in Fig. 2.4. As expected, porosity significantly enhances both the fraction of initial kinetic energy partitioned to vaporization and the total mass of vapor produced. Vaporization also increases with decreasing impact angle, consistent with the results of Schultz [1996].

2.3 Computational Comparison

Laboratory-scale numerical simulations using the CTH shock physics code [McGlaun et al., 1990; Hertel et al., 1995] allow direct comparison between experimentally measured vaporization and computational results. A widely-used method within the planetary science community for numerically estimating impact melt and vapor volumes relies upon stationary tracer particles, distributed throughout the target, to record the maximum pressure experienced during shock wave passage (see Pierazzo et al. [1997]; Kraus et al. [2011]). Critical pressures to induce phase changes are calculated from the pressure-entropy Hugoniot, and, under the assumption that post-shock dissipative processes contribute minimal additional entropy (isentropic release), these critical pressures serve as phase change criteria [Ahrens and O’Keefe, 1972].

For consistency with previous work, tracer particles track time-resolved pressures at distinct target points; tracers were distributed radially from the impact point at 15° intervals (see Fig. 2.5). Moreover, a mass filtering algorithm dynamically tracks the cumulative mass of target material subjected to pressures exceeding the critical pressures for melt and vaporization. This method benefits from enhanced resolution, as the spacing between tracer particles is typically much larger than the computational mesh spacing. Additionally, such a method does not depend on the post-processing step of fitting a sphere or other shape (an approximation that may be less accurate for oblique impacts) to melted or vaporized regions.

Yet another method to track phase changes is to observe the final release tem-

perature of the ice. This approach, unlike the critical pressure method, allows for additional heating contributions from dissipative processes during and after shock unloading. While this method has not been widely used for planetary problems, it is often employed in the shock physics community, and its application to planetary impacts is currently being explored in great depth for a variety of geological materials [Quintana et al., 2013]. As melting and vaporization temperatures are pressure-dependent, the noncumulative, total target mass exceeding the critical temperatures is assessed after the maximum shock pressures have decayed well below the critical pressures.

Library coefficients from the ANEOS equation of state (EOS) for aluminum were used to model the 0.3175 cm projectiles used in the experiments. For the icy targets, a range of H₂O ice equations of state were tested, including ANEOS water-ice, ANEOS water-ice with molecular phase transitions, and 5-Phase Water Ice [Senft and Stewart, 2008]. Initial temperatures were set to 260 K, and the porous cases utilized a P- α model [Herrmann, 1969] with initial density $\rho = 0.5$ g/cm³ ($\phi = 0.4$) and crush strength of 0.1 GPa, appropriate for this temperature regime [Stewart and Ahrens, 2004]. As melt and vapor determination is known to be highly sensitive to mesh resolution [Pierazzo et al., 1997], models incorporated 20, 40, and 80 cells-per-projectile radius (cpr) cases in 2-D. The radial pressure decay was nearly identical between 20 and 40 cpr cases, and the total calculated vapor mass (using the final release temperature) differed by only $\sim 1.6\%$ (see Fig. 2.6). Refinement to 80 cpr did not appreciably change the results. Hence, we employed 40 cpr for all 2-D models, and used CTH’s Adaptive Mesh Refinement [Crawford, 1999] capabilities in order to achieve a maximum of 25 cpr in 3-D simulations, which are necessary for modeling non-vertical impacts but bear a higher computational burden. Critical pressures and temperatures for melting and vaporization are listed in Table 2.2. Due to higher thermal pressures upon compaction, critical pressures for porous materials are lower.

Time-resolved pressures recorded by tracer particles closely match results from the mass filtering method for critical pressure. The time at which tracer particles cease to be shocked above critical pressures coincides with the convergence of the cumulative filtered mass to a maximum value (see Fig. 2.7). The melted or vaporized mass calculated from the final release temperature consistently peaks slightly after the pressure-filtered mass converges to a maximum (Fig. 2.8). With the assumption of equation-of-state accuracy for the temperatures at these conditions, dissipative processes behind the shock front must contribute to additional heating of the target.

Notably, the pressure method predicts no vaporization at these impact velocities (Fig. 2.8). As seen in Fig. 2.9, the maximum pressure near the impact point, for both the porous and nonporous cases, falls well beneath their corresponding critical pressures for vaporization, 28.36 and 66.9 GPa, respectively. The temperature method, however, yielded an improved match to experimental results. For all three equations of state (tested at 90° for comparison with lab-scale results) the temperature method produced non-zero vapor masses, which were comparable to experimentally measured masses (see Fig. 2.4).

The temperature method also successfully predicted enhancements in vapor volume with porosity for all cases. Each of the three water-ice equations of state slightly over-predicted vapor masses for the 90° case (Fig. 2.4). ANEOS provided the closest match and more accurately captured the significant increase in vapor mass between nonporous and porous ice at 90° , as found experimentally. As a result, the remainder of the simulations used the ANEOS water-ice equation of state. Vapor mass dependence on impact angle, calculated using the final release temperature metric, was more complex than previous studies [Pierazzo and Melosh, 2000; Kraus et al., 2011], which report $\sim \sin(\theta)^{0.7}$ - $\sin(\theta)^{0.8}$ dependencies. Interestingly, the temperature method predicted slightly greater vapor masses at 60° and 45° than for 90° (Fig. 2.10). The relative vapor mass calculated at 30° , however, conforms to the $\sin(\theta)^{0.7}$

relation from Kraus et al. [2011].

Comparisons between the radial shock pressure decay profile for porous and solid ice (Fig. 2.9) reveal the drastic pressure attenuation experienced by porous ice. The prominent kink in the porous ice profile, resolved by tracer particles near $2r_p$, may be related to additional thermal pressure generated during compaction of the underdense target. This feature remained at high (80 cppr) resolutions and may be unique to lab-scale conditions; it is not present in planetary-scale simulations (discussed in Section 3.4), which are conducted at higher velocities and lower target temperatures,. At sufficient radial distances (outside of the isobaric core region), shock pressure decay follows a power law [Croft, 1982]:

$$P(r) = P_0 \left(\frac{r}{r_p} \right)^{-n} \quad (2.4)$$

Importantly, pressure decay is non-axisymmetric for non-vertical impacts [Dahl and Schultz, 2001], so that the decay profile will vary with direction from the impact point. However, in the simplified 2-D case shown in Fig. 2.9, nonporous ice is fit by $n \sim 1.32$, whereas porous ice is fit by a much steeper $n \sim 3.36$. While the attenuation of shock waves in porous targets leads to lower volumes of target shocked to a given pressure, the correspondingly lower critical shock pressures needed to vaporize porous materials counteract this effect, generating greater volumes of vapor, as shown by Kraus et al. [2011]. Using a temperature-based criterion, however, simplifies calculation of vapor volumes for targets of various initial porosities, because it does not require *a priori* estimation of critical shock pressures.

2.3 Planetary Scale Models

The final release temperature criterion was shown to be more accurate than the critical pressure method for predicting vapor masses in laboratory conditions, which

are relatively warm and low-velocity, compared to planetary-scale problems of interest. With the insights gained from direct comparisons between experiments and laboratory-scale computational results, the implications for larger scale, higher velocity, and lower temperature impact events can be explored. Since power-law scaling relations for melt and vapor generation break down at low velocities (where target strength and other material properties become increasingly important) [O’Keefe and Ahrens, 1977], it is necessary to also test the efficacy of the temperature method at planetary scales. An example problem is the impact of a 1-km diameter, porous comet ($\phi = 0.4$) into Enceladus, a small saturnian satellite that has recently been estimated to possess moderate regolith porosity ($\phi = 0.2$) [Besserer et al., 2013]. Here, both target and impactor porosity may play a central role.

We modeled cometary impacts into Enceladus at $v = 25$ km/s, typical for its near-Saturn location [Zahnle et al., 2003], at 30° , 45° , 60° , and 90° incidence angles. All simulations, including the vertical impacts, were carried out in fully 3-D domains (in order to avoid artifacts arising from symmetry conditions) at 25 cppr resolution. Impactor density was maintained at a constant $\rho=0.5$ g/cm³, consistent with derived densities from recent flyby observations of comet nuclei [Thomas et al., 2013a,b]; the regolith density of Enceladus was modeled as both nonporous ($\rho=0.92$ g/cm³) and porous ($\rho=0.735$ g/cm³; $\phi = 0.2$). Both target and impactor were composed of pure water ice, using the same ANEOS water-ice equation of state utilized for the lab-scale simulations. Porosity was implemented with a P- α porosity model with a slightly higher crush strength (0.2 GPa) than the laboratory simulations, appropriate for the colder conditions in the outer solar system [Stewart and Ahrens, 2004]. A pressure-dependent yield surface with thermal softening and density degradation (referred to as “Geo-yield” in Table 2.3) incorporated a low (0.285 GPa) material strength after the porous regolith was fully compressed (see Table 2.3 for a tabulation of model parameters).

Both the critical pressure and final release temperature methods yielded estimates for vapor production. Additionally, newer versions of CTH permit direct user access of a material’s specific entropy for select equations of state. Hence, we output the cumulative target mass reaching the critical entropy to achieve vaporization, $S_v=12.67$ kJ/(K kg) [Kraus et al., 2011]. This technique, like the temperature method, does not require *a priori* information about critical pressures from the entropy-pressure Hugoniot, which greatly simplifies the incorporation of target porosity at different initial temperatures. The entropy criterion was also used for the laboratory-scale simulations but, like the critical pressure method, it did not predict any vaporization at those velocities (contrary to observations).

Results from the planetary-scale simulations are shown in Fig. 2.11. While porosity enhances vaporization under all conditions, the degree of enhancement is dependent upon impact angle and vaporization criterion. Fig. 2.12 shows the percentage enhancement in vapor mass associated with including $\phi = 0.2$ target porosity in each simulation. Porosity effects are generally more extreme in the temperature-filtered data than for the entropy- and pressure-filtered values. Notably, in nearly all cases (90° and 60° using critical pressure are the exception), the amount of vapor enhancement with porosity is substantially larger than the 20% reported in Kraus et al. [2011], and the effects of including target porosity are most extreme at oblique angles. However, this implies that key material property controls on melt and vapor generation, including porosity, cannot be cleanly decoupled from the effects of varying impact angle. Plots of the pressure distribution in the target at $t = 0.2$ s, shown in Fig. 2.13, provide context for impact angle effects on the distribution of early-time, high shock pressures. Subtle porosity effects can be seen at 30° , where the porous case attenuates the downrange-directed shock near the target surface.

Extrapolation of results from computationally efficient vertical impact simulations to non-vertical impacts is often necessary, in order to accommodate the actual impact

angle probability distribution, $P(\theta) \sim \sin(2\theta)$ [Gilbert, 1893; Shoemaker, 1962]. Our results suggest that such extrapolations should be done with care, as impact angle effects are not always independent of other model components. Of the three vapor calculation techniques, the temperature method appears to be most easily extrapolated to different impact angles when including moderate porosity effects.

While the temperature metric consistently predicts greater volumes of vapor than entropy or pressure for these simulations, trends with decreasing impact angle are well-fit by the same $\sin(\theta)^{0.7}$ scaling law found in Kraus et al. [2011]. However, this functional dependence still departs from the increased vaporization seen at oblique angles in impact experiments. Interestingly, the temperature criterion results for nonporous ice at 90° are consistent with the vapor masses calculated using the critical pressure method (for similar melt numbers) in Kraus et al. [2011], despite the use of a different equation of state.

As the effects of even moderate target porosity ($\phi = 0.2$) on vapor volumes were substantial ($\sim 40\%$), additional simulations with $\phi = 0.4$ further assessed the vaporization dependence in targets with higher initial void space, e.g., comets or Kuiper belt objects. Here only the temperature-filtered masses are used, for both simplicity and consistency with laboratory results. At higher porosities, impact angle effects on vaporization become even more important (Fig. 2.14): at 30° total vaporized mass is enhanced by over 100%, while at 90° there is a 66% enhancement. Consequently, the vapor mass dependence on impact angle is described by a smaller power law exponent for $\phi = 0.4$:

$$m_\theta/m_{90} \sim \sin(\theta)^{0.5} \quad (2.5)$$

At each impact angle, the vapor mass enhancement between $\phi = 0.0$ and $\phi = 0.4$ is roughly twice the enhancement calculated between $\phi = 0.0$ and $\phi = 0.2$, for an approximate average increase of 80%, relative to the nonporous case. The principal differences between the results yielded by our temperature-based approach and

previous results are: (1) greater enhancements with increasing porosity, which are consistent with experimental results; and (2) dependency of porosity effects on impact angle.

The contrasting shock pressure decay curves for ice targets of varying porosity are plotted in Fig. 2.15, where shock pressure profiles were taken along the 75° line beneath vertical impacts. Note that, unlike the $v=5$ km/s, laboratory-scale case (Fig. 2.9), porous planetary-scale targets subjected to $v=25$ km/s impacts transition smoothly from the isobaric core region to the power-law decay regime. Consequently, near-field porosity effects may become more complex in the case of lower-velocity impacts, which are typical in the Kuiper belt [Trujillo et al., 2001; Durda and Stern, 2000]. Additionally, the porous decay profile closely matches the nonporous decay profile for $d/r_p < 4$, which likely corresponds to the target region compressed to full density. Beyond this region, the porous profiles are described by the larger decay exponents expected for pressure-attenuating materials. Least-squares solutions for the decay exponents (Eq. 2.4) yield $n_{\phi=0} = 2.67$, $n_{\phi=0.2} = 4.62$ (for $d/r_p > 5$), and $n_{\phi=0.4} = 5.07$ (for $d/r_p > 5$). Notably, the pressure decay curves for $\phi = 0.2$ and $\phi = 0.4$ porosity diverge only slightly from each other, beginning at $d/r_p \sim 6$.

2.4 Discussion

Impact experiments into porous and nonporous water ice at varying incidence angles demonstrate significant vaporization, yet matching laboratory-scale hydrodynamical models do not record any impact-generated vapor when using the peak shock pressure criterion. The final release temperature metric provided a superior match to experimentally-derived vapor masses; however, large enhancements in vaporization efficiency recorded at oblique angles [Schultz, 1996], were not captured by the model results. More realistic porosity or strength models may be necessary to fully match the experimental results. The inclusion of a simple geological yield model for strength

(see Table 2.2) achieved modest increases ($\sim 3\%$) in vapor mass for the 30° case but did not affect the 90° case, thereby suggesting that shear strength effects on dissipative heating are more important for oblique impacts. However, these effects are not sufficient to explain the large vapor masses seen experimentally at oblique angles, which are related not only to shear heating but also to downrange scouring associated with projectile failure [Schultz, 1996]. More work is required to refine our computational approach so that it successfully reproduces enhancements in vaporization at lower incidence angles.

A more accurate approximation of experimentally measured vapor masses using final release temperature motivates this investigation of the temperature metric to planetary-scale impacts into ice. As many icy planetary surfaces contain significant void space, accurate accounting of extra vapor generation from porous targets is critically important. Calculating the total vaporized mass from final release temperatures demonstrates that porosity, implemented in the form of a P - α model, increases vaporization to a greater extent than previously calculated: $\sim 40\%$ enhancement at $\phi=0.2$, and $\sim 80\%$ enhancement at $\phi=0.4$. In particular, porosity is found to be increasingly important at oblique angles, whether using pressure, entropy, or temperature as a phase change criterion. Given the longer penetration times associated with under-dense, porous targets, and the larger volume of target material with which the projectile interacts during the penetration stage of oblique impacts [Schultz et al., 2007], a greater role for porosity at oblique angles is logical. Both of these findings suggest that impact vaporization of ice, particularly for primordial icy bodies, may be substantially more efficient than previously assumed. Additionally, impact angle effects are not easily separable from the effects of porosity, which implicates larger roles for other key material properties, including strength, at oblique angles. Further study is necessary to fully capture the combined effects of more realistic material models and non-vertical impact angles.

2.4 Implications

Enhanced vapor generation will work to accelerate the impact-driven loss of water ice from icy satellites [Nimmo and Korycansky, 2012], affect the accretionary and collisional histories of icy bodies [Ahrens and O’Keefe, 1985; McKinnon, 1989; Canup, 2005; Canup and Ward, 2009], and can drive additional atmospheric erosion at bodies such as Mars [Melosh and Vickery, 1989]. Ongoing characterization of Kuiper belt objects demonstrates that significant porosity can be retained over longer time scales, even for relatively large bodies; for instance, Kuiper belt object 2002 UX25, at a diameter of ~ 650 km, possesses a low density ($\rho = 0.82 \pm 0.11$ g cm $^{-3}$), which requires non-trivial interior void space [Brown, 2013]. The long-term processing of such icy and porous bodies, including the assemblage of dwarf planets with significantly higher rock fractions [Brown, 2013], is closely tied to the outcomes of giant impacts [Brown et al., 2007; Leinhardt et al., 2010]. Since impact velocities are relatively low within the Kuiper belt ($\Delta v = 1.16$ km/s) [Trujillo et al., 2001], the effective lowering of threshold velocities to achieve melting and vaporization, through the inclusion of porosity, is particularly important.

While this study focused on vapor generation, in order to incorporate experimental results, the large enhancements seen in vapor with increasing porosity may have additional implications for impact melt volumes. Previous work using the peak pressure method has found decreasing melt volumes with higher porosities [Kraus et al., 2011], due to the pressure attenuation beyond the isobaric core region (Fig. 2.15). This contrasts with results for silicate materials, for which increasing porosity is associated with increased melt production [Wünnemann et al., 2008]. However, whether porosity enhances or suppresses melting may be related to the radial distance at which phase changes no longer occur; Kraus et al. [2011] hypothesized that the decreased melt in porous ice was controlled by the relation between the size of the isobaric core and the distances at which melting takes place. For lower velocity

impacts into cold targets, porosity effects may not be sufficient to induce vaporization but could instead lead to significant increases in melt generation, as the zone of melting is more localized. Based upon the vaporization results reported here, we anticipate that porosity and strength will play greater roles in melt production at oblique angles. A thorough investigation of the final release temperature method’s application to melt and vapor generation in a variety of targets is ongoing [Quintana et al., 2013] and will be reported in depth in a future contribution.

2.5 Conclusions

Vapor plumes from impacts into porous and nonporous ice demonstrated significant enhancements in vaporization with porosity and at shallower impact angles. Direct comparison with laboratory-scale numerical simulations established the final release temperature metric as the most accurate criterion for vapor mass estimates. Application of the temperature, critical pressure, and entropy metrics to compute vapor masses from planetary-scale problems revealed that: porosity-driven enhancements in vaporization are exaggerated at oblique angles, for all phase change metrics; integrated over the impact angle distribution, vaporized mass was enhanced by 40% for $\phi=0.2$ and by 80% for $\phi=0.4$, significantly greater than prior estimates; depending upon the impact conditions, using a combination of phase change metrics may provide the most complete information; and the temperature metric is a convenient way to account for heating enhancement, as it does not require a separate calculation to derive critical pressure at a given initial porosity

These results indicate that impact angle effects cannot always be cleanly decoupled from an event under a given set of conditions; the manner in which target vaporization occurs may depend upon the interplay between target and projectile material properties and impact angle. Porosity may play a larger role than previously assumed in the effective lowering of critical velocities to achieve melting and/or vaporization of

water ice. This effect could have particular consequences for the collisional histories of trans-Neptunian objects, which experience predominantly low velocity impacts. In light of the temperature method’s success in predicting vaporized target masses from impact experiments, this strategy to predict vapor and/or melt may prove very useful in future computational studies of impact-generated phase changes.

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Tables

Table 2.1: Experimental Results: Vapor Plume Mass and Energy

Target	θ	v (km/s)	m_p (g)	P (mbar)	u_{max} (km/s)	u_∞ (km/s)	m_v/m_p	E_v/KE_i
Porous	90°	5.36	0.376	7.93	0.901	0.317	1.7	0.006
Porous	60°	5.34	0.376	8.38	1.18	0.414	5.9	0.035
Porous	30°	5.47	0.376	7.98	1.46	0.515	18	0.159
Nonpor.	90°	5.52	0.376	8.10	0.682	0.240	0.11	0.0004
Nonpor.	30°	4.99	0.376	9.37	1.80	0.634	9.6	0.155

**Table 2.2: Computational Parameters:
Laboratory-scale Simulations**

Parameter	Value
T_{init}	260 K
v_{impact}	4.99-5.52 km/s
Projectile EOS	ANEOS Aluminum
Target EOS	ANEOS Water-Ice
$R_{projectile}$	0.3175 cm
ρ_{H_2O} (porous)	0.50 g/cm ³
ρ_{H_2O} (nonporous)	0.92 g/cm ³
T_{melt}	273.15 K
T_{vap}	373.15 K
P_{melt} (porous) ^a	1.23 GPa
P_{melt} (nonporous)	2.69 GPa
P_{vap} (porous)	28.36 GPa
P_{vap} (nonporous)	66.9 GPa
P- α Crush Strength ^b	0.1 GPa
Geo-yield Y	0.185 GPa
Geo-yield dY/dP	0.7
Geo-yield Poisson	0.33

^aCritical pressures from Kraus et al. [2011]

^bStewart and Ahrens [2004]

**Table 2.3: Computational Parameters:
Planetary-scale Simulations**

Parameter	Value
T_{init}	75 K
v_{impact}	25 km/s
Projectile EOS	ANEOS Water-ice
Target EOS	ANEOS Water-Ice
$R_{projectile}$	0.5 km
$\rho_{projectile}$	0.50 g/cm ³
ρ_{target}	0.50-0.92 g/cm ³
T_{melt}	273.15 K
T_{vap}	373.15 K
S_{melt}^a	3.51 kJ/(K kg)
S_{vap}	12.67 kJ/(K kg)
$P_{melt} (\phi=0.4)^b$	1.23 GPa
$P_{melt} (\phi=0.2)$	2.20 GPa
$P_{melt} (\phi=0.0)$	3.48 GPa
$P_{vap} (\phi=0.4)$	28.4 GPa
$P_{vap} (\phi=0.2)$	46.2 GPa
$P_{vap} (\phi=0.0)$	69.6 GPa
P- α Crush Strength ^c	0.2 GPa
Geo-yield Y	0.285 GPa
Geo-yield dY/dP	0.7
Geo-yield Poisson	0.33

^aCritical entropies from Kraus et al. [2011]

^bCritical pressures from Kraus et al. [2011]

^cStewart and Ahrens [2004]

Figure Captions

Figure 2.1: Impacts into porous ($\phi = 0.4$) ice targets at 90° (top) and 30° (bottom) generate distinctly different vapor plume morphologies. The 30° case produces an apparently more massive and luminous plume, which propagates downrange while expanding hemispherically, while the 90° case simply expands hemispherically above the impact point. The downrange-directed plume permits prolonged interaction with the target surface, subsequent to initial shock heating. This, coupled with enhanced roles for target strength and ricochet at lower incidence angles [Schultz, 1996] may account for the larger vapor masses at oblique angles. Images are 66, 134, and 200 μs after impact (from left to right).

Figure 2.2: Contrasting vapor plumes produced by 30° impacts into nonporous (left) and porous (right) ice, imaged at 48 μs . The longer projectile penetration times associated with impacts into porous targets result in suppression of the observable jetting phase at the target surface. Any initially-generated, high-speed jetting material will interact with target material within the penetration ‘funnel’ before being released at the surface, creating a more hemispherical and slowly-propagating plume. The subsurface details of this process are visible in quarter space experiments for under dense silicate [Schultz et al., 2007] and ice [Bruck Syal and Schultz, 2014] targets. In contrast, impacts into nonporous ice result in a more clearly defined jetting phase, where the leading edge of the vaporized mass, generated at first contact, travels rapidly downrange, elongating the observed vapor plume shape.

Figure 2.3: The evolving vapor plume radius, measured at early times with 8 μs resolution and at later times with 24-66 μs resolution, is plotted against time to derive, first, the energy of the plume upon full coupling to the atmosphere (denoted in

blue), using Eq. 2.3. At slightly earlier times, when expansion is approximately isentropic but the plume’s energy is not fully coupled to the tenuous atmosphere within the impact chamber ($P \sim 8$ mbar), a value for u_{max} is determined, allowing the mass of the vapor cloud, m_v , to be calculated from Eq. 2.1. Vapor results from impact into porous ice at 90° are shown here.

Figure 2.4: Experimentally measured vapor masses, normalized to the projectile mass $m_v = 0.376$ g (see also Table 2.1), plotted for comparison with results from laboratory-scale CTH simulations. As the critical pressure method predicted zero vaporization at these scales, masses were determined from the final release temperature method. A range of equations of state were tested at 90° , for both porous and nonporous cases. As ANEOS water-ice (blue squares) provided the closest match to experimental results (black circles), we utilized ANEOS for the remainder of the simulations in this work. However, numerical results did not reproduce the experimentally observed increases in vapor mass at oblique angles. Implementing a “Geo-yield” strength model (blue exes) increased the vapor mass at 30° only marginally (3%).

Figure 2.5: Tracer particle geometry for CTH calculations at laboratory scale. Stationary tracer particles are widely used in the planetary science community to infer melt and vapor volumes, using the maximal pressures experienced by each particle and then fitting the vaporized or melted region to a buried sphere. We use tracer particles here for consistency with past methods and to provide proof of concept for the dynamic mass filtering approach (see Fig. 2.7). As tracers are typically distributed at spatial intervals exceeding the grid resolution of the problem, dynamic filtering may be particularly important when melted or vaporized regions depart significantly from the nominal spherical shapes, e.g., oblique impacts, impacts into targets with complex material properties, or impacts into layered targets.

Figure 2.6: Total ice target mass exceeding vaporization temperature (red) and critical vaporization pressure (blue) for 90° laboratory-scale impact into porous ice, plotted as a function of time. Mass is normalized by projectile mass, $m_p = 0.376$ g. Increasing resolution from 20 to 40 cells-per-projectile-radius (cpr) yielded only 1.6% additional vapor mass. All 2-D simulations were carried out at 40 cpr; 3-D simulations used 25 cpr. Note that the critical pressure method does not predict any vaporized mass, contrary to observations.

Figure 2.7: As a test of the mass-filtering algorithm, the time-resolved pressure experienced by tracer particles distributed radially from $6.6R_p$ - $7.5R_p$ (top) is compared to the filtered total mass exceeding the critical shock pressure of 1.23 GPa. Near 6 μ s, tracer particles cease to be shocked above 1.23 GPa and, concurrently, the cumulative filtered melt mass converges. This demonstrates that direct filtering via critical pressure is consistent with the static tracer particle methods employed by previous melt and vaporization studies.

Figure 2.8: Example time-resolved plot of total target mass exceeding critical pressures (blue) and temperatures (red) for incipient melting, melting, and vaporization. Pressure values represent cumulative mass exposed to critical pressures, while temperature values are non-cumulative (total target mass exceeding 272 K or 372 K temperature at a given time). Note that temperature values peak slightly after pressure-filtered values converge to a maximal value, suggesting the importance of dissipative processes.

Figure 2.9: Modeled peak pressures as a function of radial distance from impact point for 90° laboratory-scale impacts into porous and nonporous ice. Critical pres-

tures to achieve vaporization are denoted by the red (porous case, 28.36 GPa) and blue (nonporous case, 66.7 GPa) lines; at these velocities ($v \sim 5$ km/s), no vaporization is predicted by the critical pressure method. Outside of the isobaric core region ($r/r_p > 2$), the pressure decay within the porous ice is much steeper ($n \sim 3.36$) than within the nonporous ice ($n \sim 1.32$). The ‘kink’ in the porous curve near $2R$ may be related to the relatively low impact velocity.

Figure 2.10: Vaporized mass as a function of impact angle for laboratory-scale impacts into porous ice ($v = 5.47$ km/s, $T = 260$ K). While vaporization was found to increase slightly at 60° and 45° , from the vertical case, the trends do not match the dramatic enhancements at lower angles, which were seen experimentally. Only the 15° case is well fit by the $\sin(\theta)^{0.7}$ relation from Kraus et al. [2011].

Figure 2.11: Vaporized mass as a function of impact angle for impacts into porous ($\phi = 0.2$) and nonporous Enceladus regoliths ($v = 25$ km/s, $T = 75$ K). Total vaporized mass was calculated using three different criteria: final release temperature (red), specific entropy (purple), and critical pressure (blue). While the temperature method yielded the highest vapor volumes, the temperature-filtered results were consistent with those calculated in Kraus et al. [2011] for similar melt numbers. Additionally, the decay in vapor mass with decreasing impact angle was well-fit by the $\sin(\theta)^{0.7}$ relation derived in Kraus et al. [2011] (red dotted lines).

Figure 2.12: Percentage enhancement in vapor mass associated with moderate porosity ($\phi=0.2$) in the Enceladus regolith. Porosity effects vaporization to greater degrees at lower impact angles, due to longer projectile penetration times. Integrated over the range of impact angles, $\phi=0.2$ increases the total vaporized mass by $\sim 40\%$.

Figure 2.13: Pressure in porous ($\phi=0.2$) and nonporous ($\phi=0.0$) ice targets at $t = 0.2$ s, for varying impact angles. At oblique angles, pressure attenuation by the porous target is more exaggerated, due to asymmetric shock conditions and longer coupling times for momentum and energy. These effects contribute to greater porosity-driven increases in vapor at oblique angles.

Figure 2.14: (a) Vaporized mass as a function of incidence angle for high ($\phi = 0.4$), moderate ($\phi = 0.2$), and no porosity, using final release temperature. (b) Percentage enhancement in vaporized mass, relative to nonporous ice, for high and moderate porosity targets. At higher porosity, impact angle effects become more important; hence, the decay in vapor volume with impact angle is more gradual ($m_\theta \sim \sin(\theta)^{0.5}$). On average, moderate and high porosity enhance vaporization by $\sim 40\%$ and $\sim 80\%$, respectively.

Figure 2.15: Peak pressure decay with radial distance for 90° impact of 1 km-diameter comet ($v = 25$ km/s) with nonporous, 0.2 porosity, and 0.4 porosity icy targets. In transition from isobaric core to power-law decay regime, porous target pressures are nearly coincident with nonporous pressures, due to compaction to full density, but then transition to steep decay slopes ($n = 4.62$ for $\phi = 0.2$, $n = 5.07$ for $\phi = 0.4$).

Figures

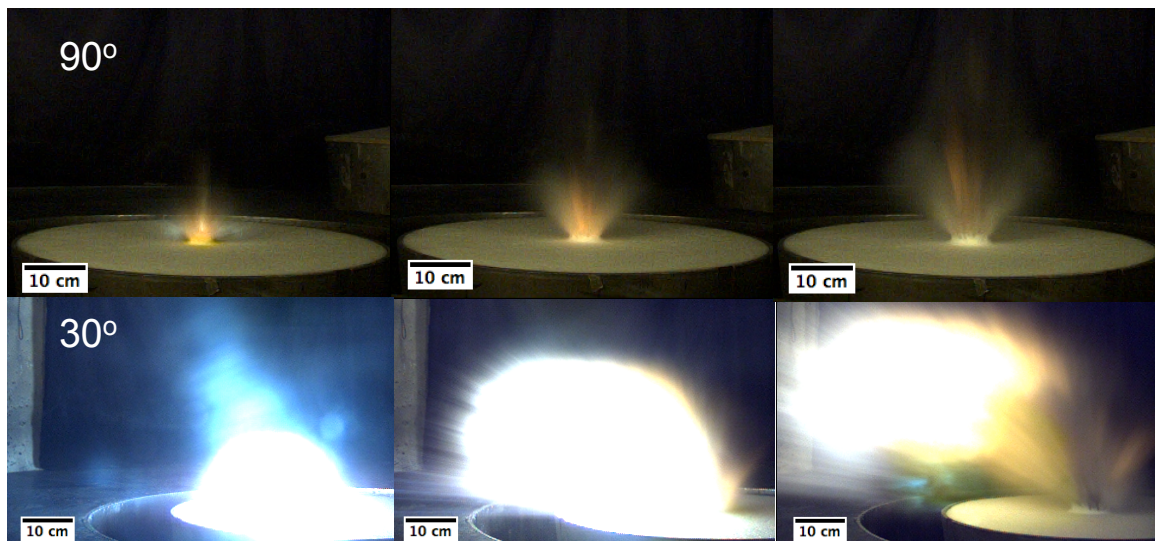


Figure 2.1

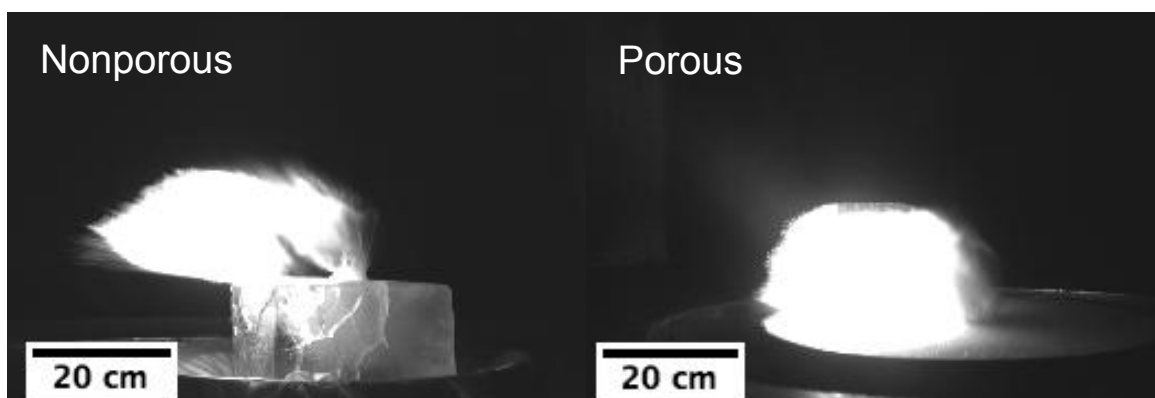


Figure 2.2

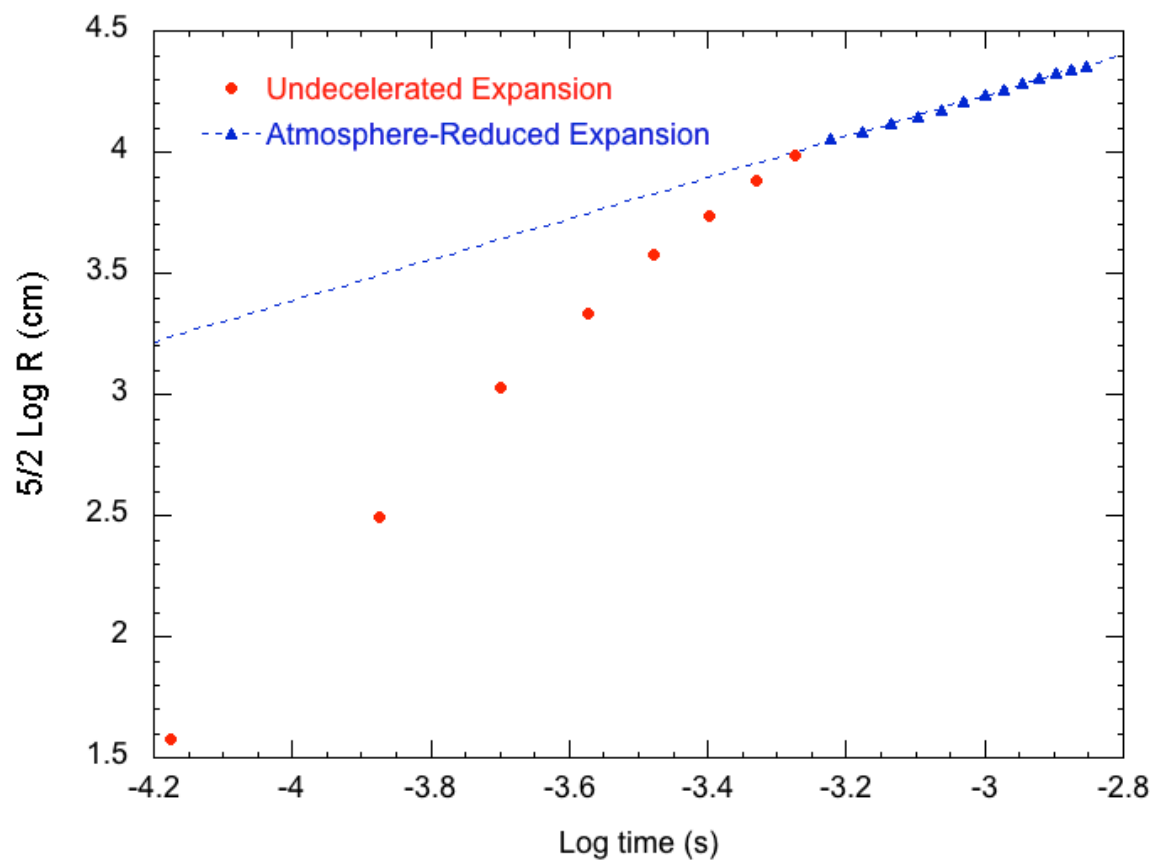


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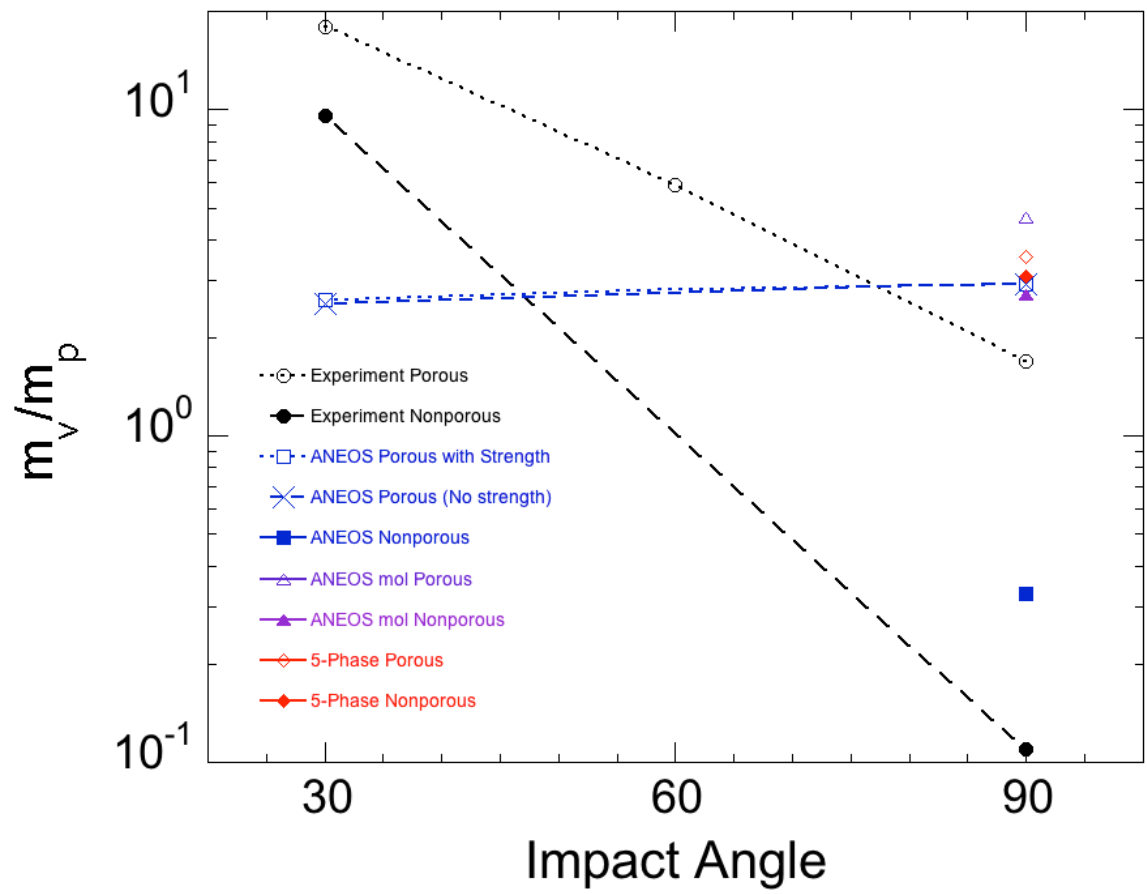


Figure 2.4

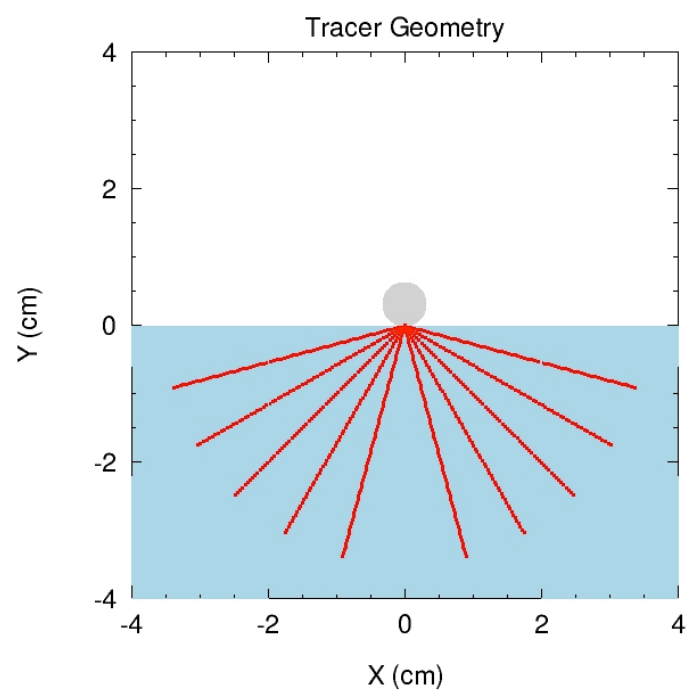


Figure 2.5

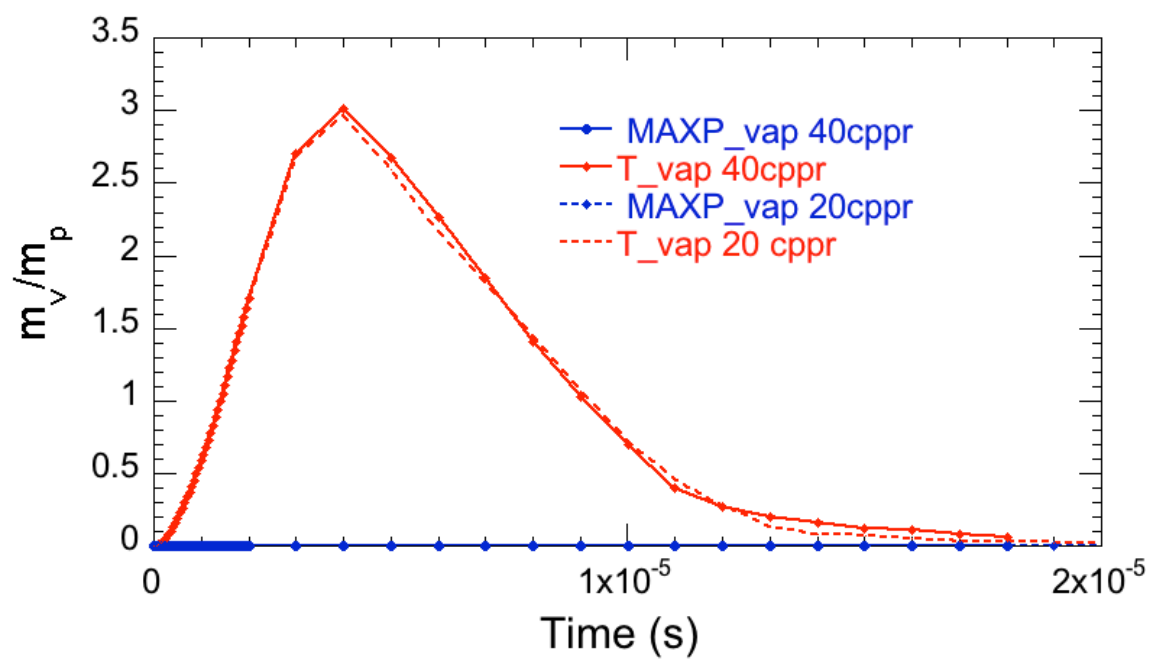


Figure 2.6

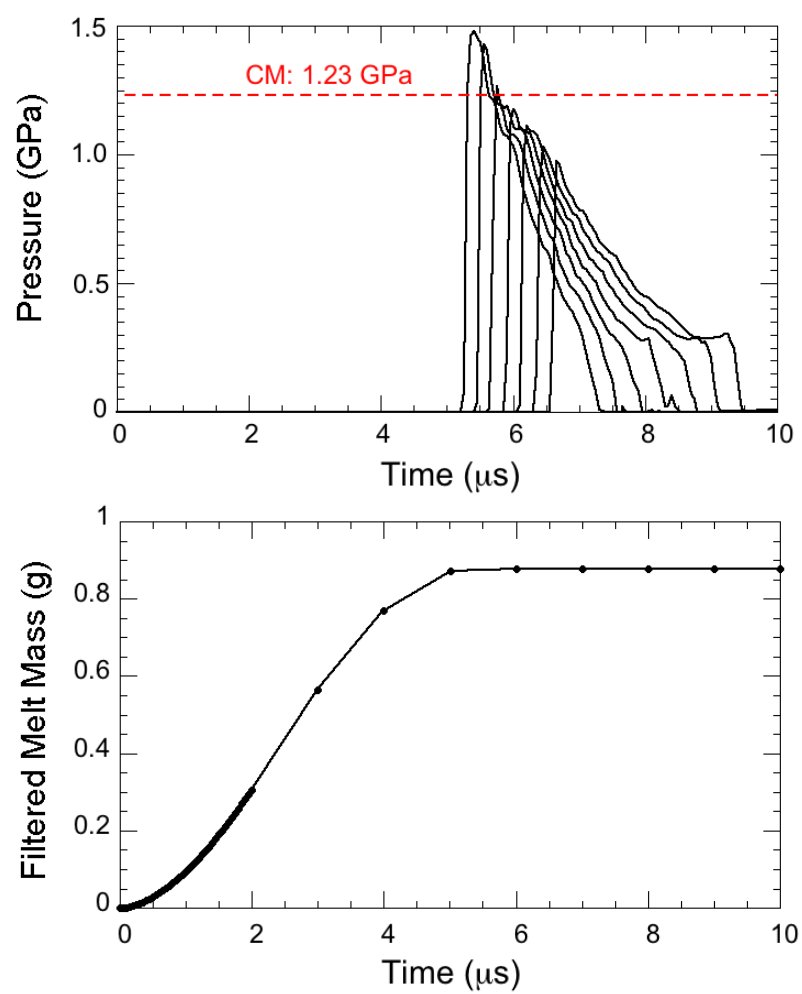


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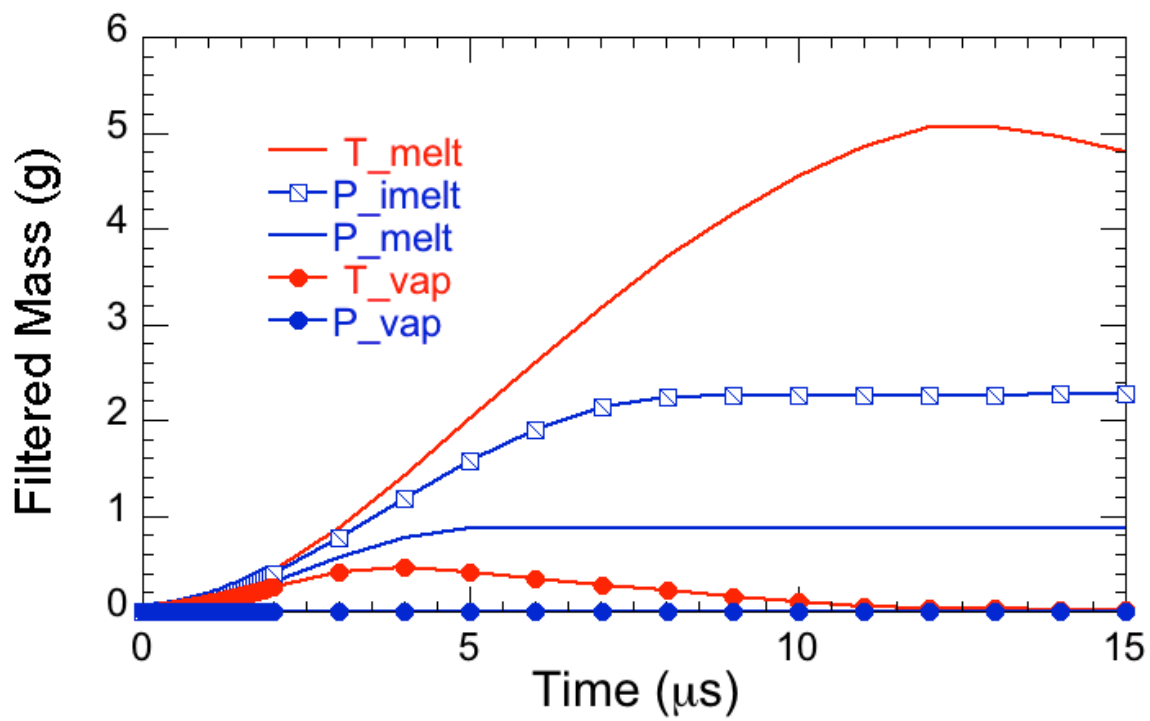


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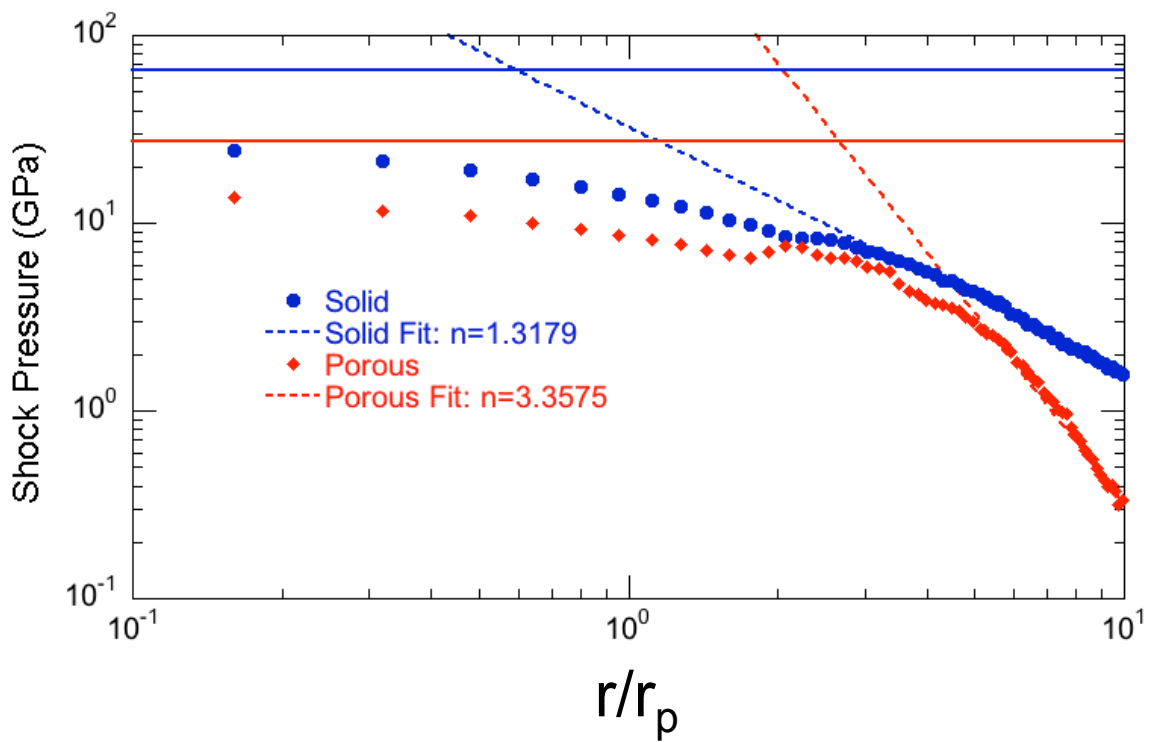


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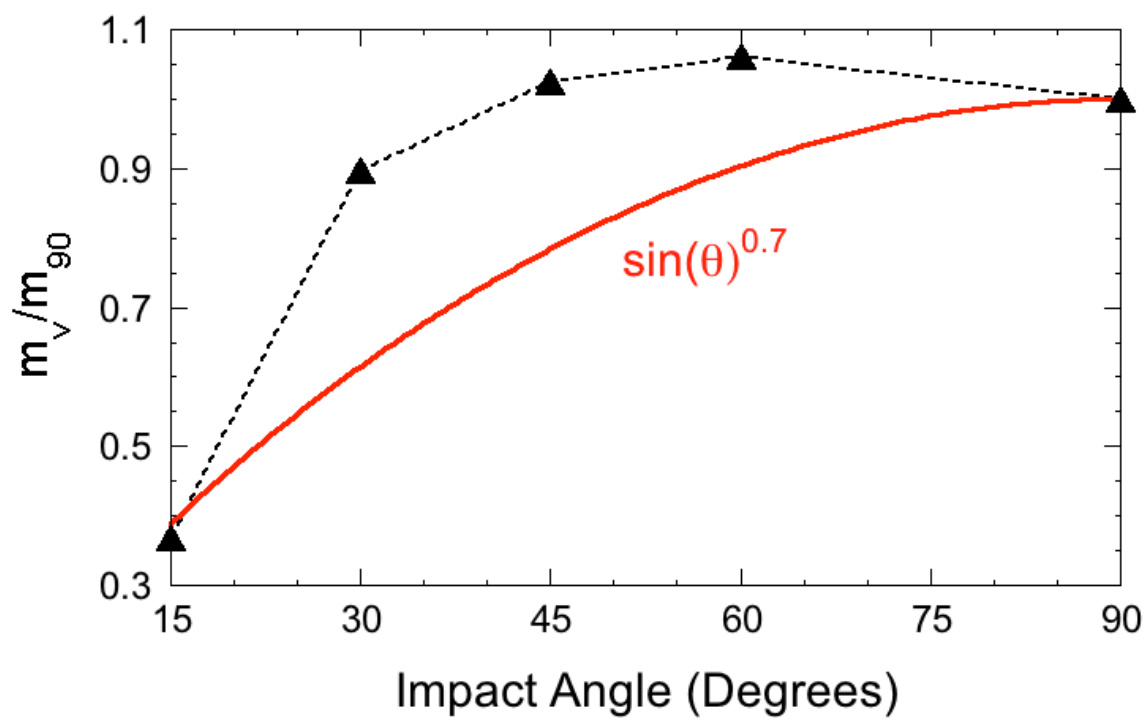


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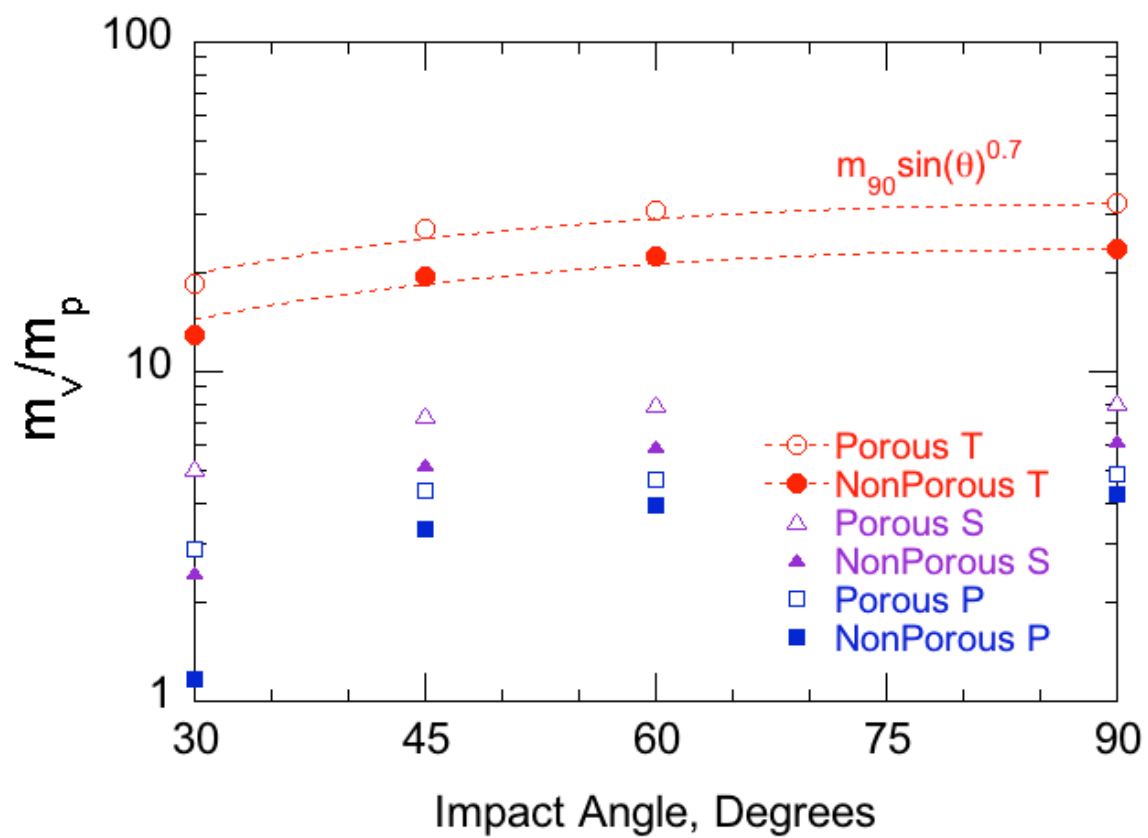


Figure 2.11

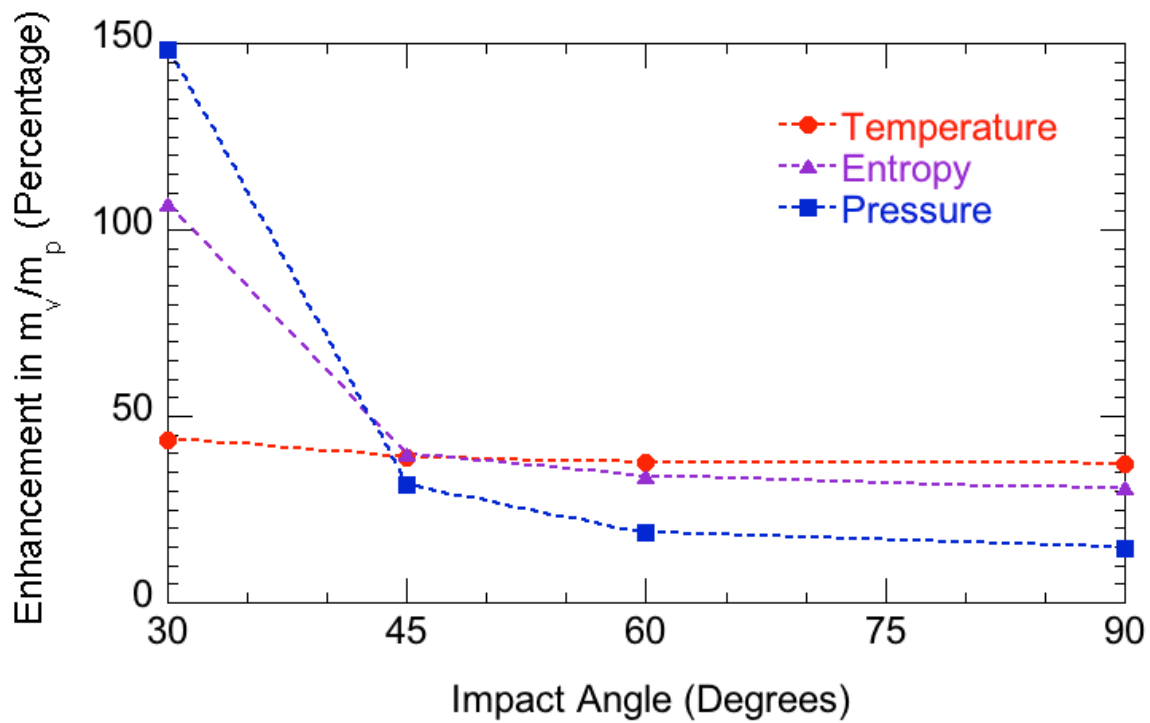


Figure 2.12

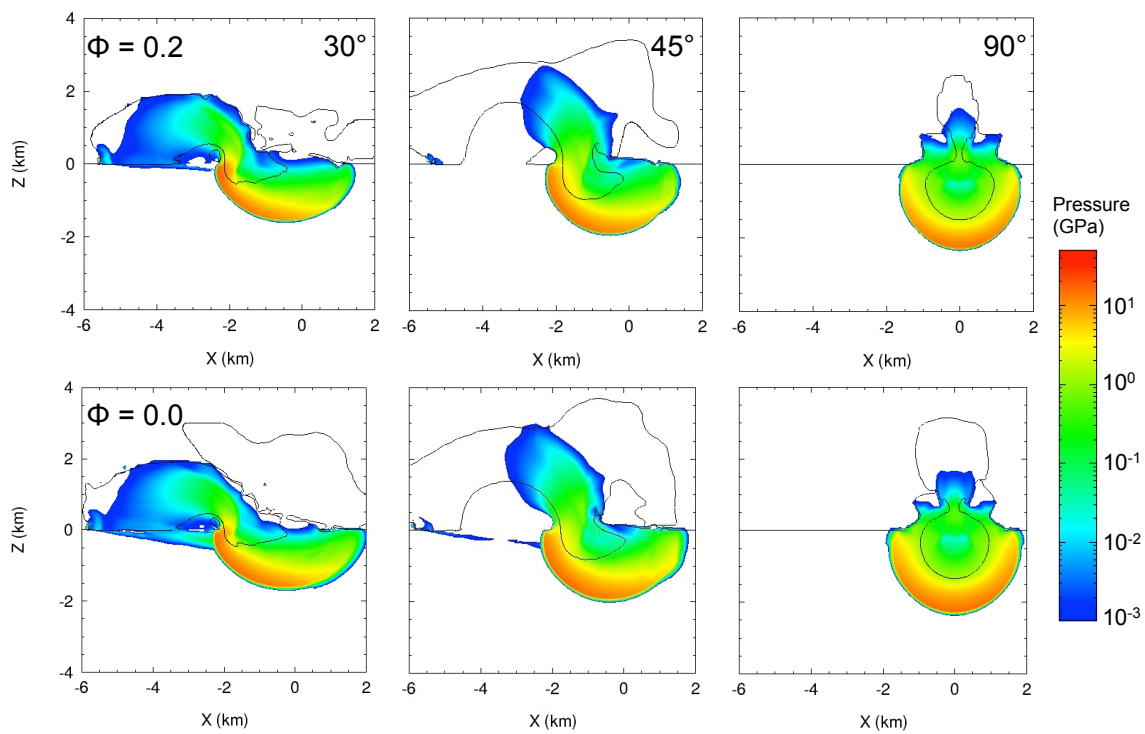


Figure 2.13

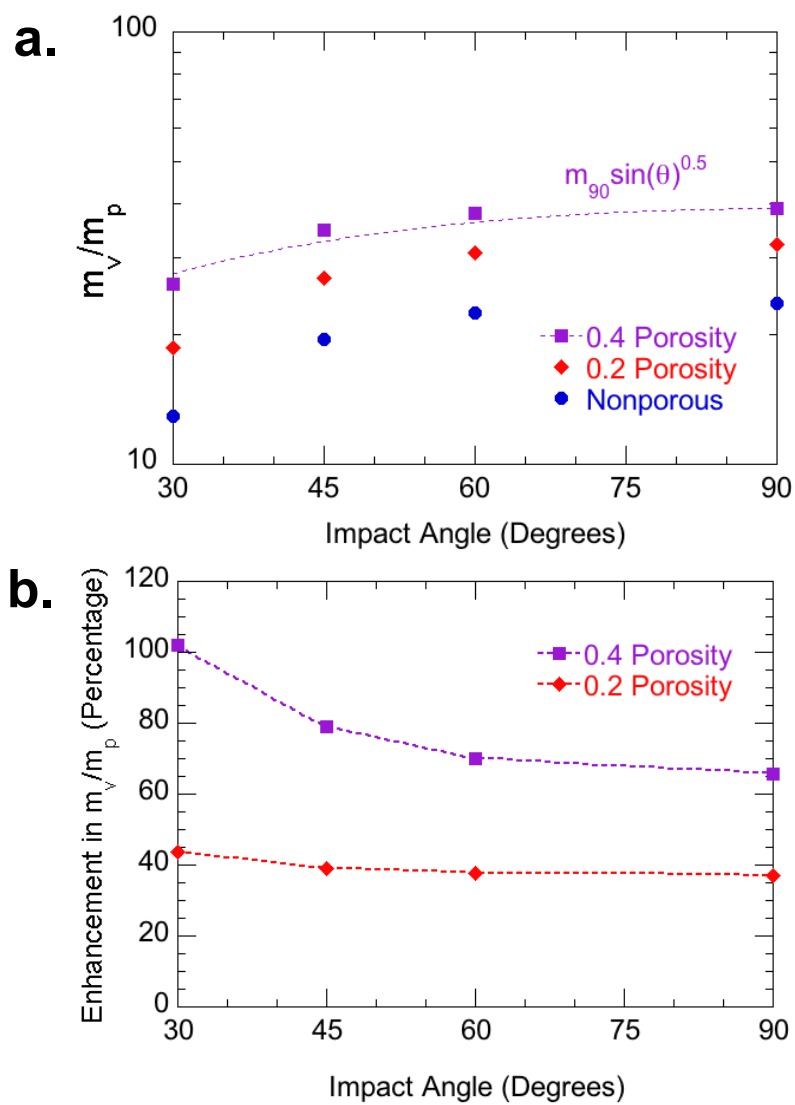


Figure 2.14

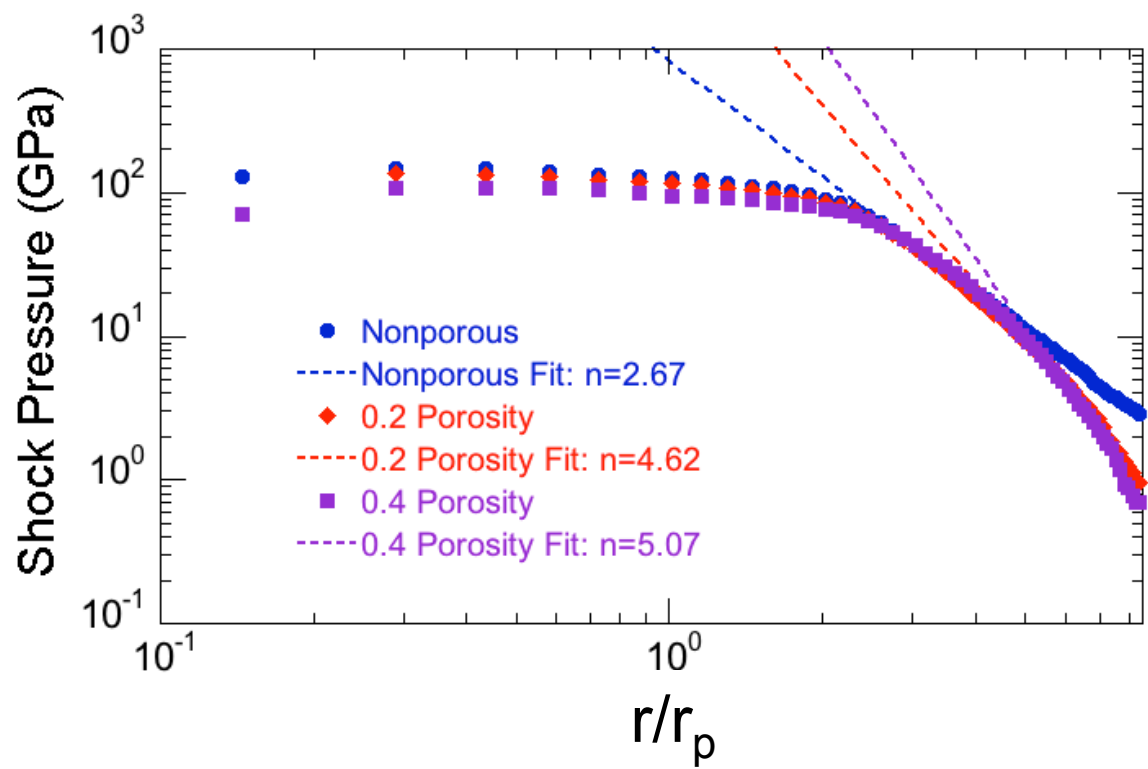


Figure 2.15

Chapter 3. Impacts Into Porous Water Ice: Time-resolved Transient Crater Growth and Non-proportional Scaling

Megan Bruck Syal and Peter H. Schultz

Department of Geological Sciences, Brown University

Providence, RI 02912

Abstract

Hypervelocity impact experiments into porous water ice, conducted at the NASA Ames Vertical Gun Range, provide the first time-resolved measurements of transient crater formation in icy targets. The effects of impact angle, projectile density, and target porosity on crater development and final morphologies were assessed in low-strength water ice, for comparison with quartz sand, a well-studied particulate material often used to scale laboratory results to planetary-size impact events. Cratering in the ice targets, which were of moderately high porosity (40%), did not conform to point source solutions for crater growth. Instead, the early-stage of crater development, during which projectile effects are important, was of longer duration in the ice targets than in lower-porosity sand. All experiments exhibited non-proportional crater growth, and the transition to non-proportional scaling of crater dimensions was documented for projectiles of different acoustic impedances. The variables controlling non-proportional scaling (velocity, gravity, material properties), combined with experimental data, were then shown to be consistent with transition diameters for pitted craters at Ganymede, Callisto, Mars, and the Moon. Impact angle and projectile-target density ratios controlled final crater dimensions, including diameter-depth trends in ice, sand, and ice-sand mix targets. Low-impedance impactors at lower angles had significant effects on the diameter-depth ratios in ice, an observation relevant to outer solar system craters, which are principally formed by low-density comets ($\rho = 0.4\text{-}0.6 \text{ g/cm}^3$). Additionally, cross-sectional views of crater development in porous ice are compared directly to hydrodynamical model results, demonstrating similarities and subtle differences between the experimental data and calculations using the commonly-applied $P - \alpha$ porosity model.

3.1 Introduction

For many solar system bodies, asteroid and comet impacts occur in surfaces comprised predominantly of water ice. Examples range from the icy moons of the outer solar system, to primitive bodies such as comets, Trojan asteroids, and Kuiper belt objects, and also to inner solar system surfaces, including icy layered terrains on Mars and perhaps even ice sheets at Earth. Accurate interpretation of the cratering record at these bodies depends upon a thorough understanding of cratering mechanics in ice, including any differences in scaling relationships between ice and rock, along with possible differences in geomorphology that may arise from the specific material properties of ice. For example, the presence of surface or subsurface ice is often invoked as an explanation for craters containing central pits at Ganymede and Callisto [Schenk, 1993; Bray et al., 2012] and at Mars [Wood et al., 1978; Barlow and Bradley, 1990]; Fig. 3.1 depicts an example central pit crater at Ganymede. However, the volatile-driven, theoretical processes to explain such models have yet to be observed experimentally.

Central pit craters have long been observed to occur at the Moon [Schultz, 1976, 1988] and Mercury [Schultz, 1988]; recent acquisition of higher-resolution spacecraft data has enabled additional mapping of central pit distributions at both bodies [Xiao and Komatsu, 2013; Xiao et al., 2014]. These observations suggest that central pit formation need not depend on target volatile abundances; rather, the presence of these features can also be attributed to the way in which an impactor initially couples its momentum and energy to the target, which does not require the presence of ice and has been experimentally verified [Schultz, 1988]. Hence, testing the differing cratering styles of icy particulate targets should reveal new information on whether these crater features may be directly related to volatile abundances, or whether other factors, e.g., velocity, gravity, material properties, may contribute to their occurrence.

The central pit craters discussed here are distinct from the much smaller (hundreds of meters), “nested” crater features seen at the Moon [Quaide and Oberbeck, 1968]. As demonstrated experimentally by Quaide and Oberbeck [1968], formation of nested craters is related to the penetration of a more competent substrate beneath a loosely consolidated layer; observed nested crater dimensions at the Moon then place constraints on the lunar megaregolith thickness. In the case of nested crater formation, excavation of the larger, shallower outer portion of the crater is controlled by gravity scaling, while the smaller crater at center resides within the strength regime. Later experiments document how the same process may apply to observations of the Deep Impact crater at Comet 9P/Tempel 1 [Schultz et al., 2007; Schultz et al., 2013b]. In contrast, central pit craters, which are the focus of this work, do not require any compositional or structural transitions with depth in the target.

One-dimensional shock experiments have recently refined existing equations of state for porous and nonporous ices subjected to extreme pressures [Stewart and Ahrens, 2004; Stewart and Ahrens, 2005; Senft and Stewart, 2008]; these efforts allow improved computational approaches to icy impacts. However, very limited experimental data is available for transient crater development and scaling in water ice targets. Hydrodynamical approaches to crater formation, while powerful, are still dependent on constraints provided by experiments, as final crater depths and diameters are controlled by the implementation of realistic material models for strength and porosity. While early-time phenomena, e.g., penetration/compression stage, shock-driven vaporization and melting, are relatively straightforward to simulate, running hydrocode models until the cessation of excavation requires larger domains and is computationally expensive.

For rocky targets within the gravity-controlled size regime, impact experiments conducted into sand [Gault and Wedekind, 1978] or wet sand [Schmidt and Housen, 1987] permit scaling laboratory results up to planetary scale problems of interest.

Estimates of cratering efficiencies at icy bodies generally rely on the assumption that scaling works similarly in rock and ice [Zahnle et al., 2003; Chapman and McKinnon, 1986]. Analogous to the use of low-strength quartz sand targets for examining crater growth within the gravity regime, we can use pre-fractured, low-strength, porous ice to simulate crater formation in ice at larger scales, where gravity dominates over material strength.

The introduction of porosity adds a layer of complication to the problem, as energy partitioned into compression of the target affects ejecta velocities and crater growth at even moderate porosities [Hermalyn and Schultz, 2011]. At the higher porosities typical of many asteroids and comets [Britt et al., 2002; Thomas et al., 2013a,b], the suppression of ejecta and the predominance of compaction-driven crater growth have more extreme effects on crater scaling relations [Schultz et al., 2005; Schultz et al., 2007; Collins et al., 2011; Housen and Holsapple, 2012]. Porosity is now recognized as an important material property not only for small bodies, but also within the megaregolith of the Moon [Wieczorek et al., 2013], icy satellites [Besserer et al., 2013], and some relatively large Kuiper Belt Objects [Brown, 2013]. Hence, accurate accounting of porosity’s effects on crater scaling relations and the additional, irreversible heating generated through target compaction remains an area of ongoing study. A detailed comparison of transient crater growth in porous ice, sand, and ice-sand mixtures allows the applicability of similar scaling relations between rock and ice to be tested, along with providing benchmarking measurements for hydrocode comparisons. The quarter-space geometries used in this study yield the first time-resolved views of crater formation in ice, for comparison with other well-studied granular targets. They also extend observations of transient crater dimensions in porous targets to earlier times than previously achieved, due to the suppression of impact flash effects by the ice.

3.2 Experimental Approach

Impact experiments were conducted at the NASA Ames Vertical Gun Range (AVGR), at incidence angles ranging from 30° to 90° and with $v \sim 5$ km/s. We utilize quarter-space targets, which provide a cross-sectional view of the cratering process through a transparent acrylic window (Fig. 3.2). By impacting the target within one projectile radius of the window's edge, crater growth is visualized at the event's axis of bilateral symmetry, parallel to projectile trajectory. This viewing geometry provides time-resolved information on transient crater depth and diameter, as well as additional information about processes occurring inside of the transient cavity (e.g., vaporization) and compaction within the subsurface.

The principal target material explored in this study was porous ($\rho = 0.5$ g/cm³) water ice, although impact experiments into quartz sand ($\rho = 1.7$ g/cm³) and an ice-sand mixture ($\rho = 0.7$ g/cm³) also were conducted, in order to directly compare ice with other granular materials. The ice and ice-sand mix targets were prepared by sifting commercially-manufactured snow into quarter-space buckets in a 260 K environment. A combination of spherical polyethylene ($\rho = 0.92$ g/cm³) and aluminum ($\rho = 2.78$ g/cm³) projectiles permitted probing the effects of different acoustic impedance (sound speed times density) conditions on crater growth. See Table 3.1 for a complete listing of all experiments reported here. The ice targets were kept at ~ 260 K until just prior to shooting, and low atmospheric pressures ($P \sim 8$ mbar) were maintained in the impact chamber during experiments. Two pairs of high frame-rate cameras (black and white Shimadzu, 2-8 μ s time resolution, and color Phantoms, 24-67 μ s time resolution) tracked crater formation throughout all experiments.

3.3 Transient Crater Growth

3.3 Projectile Effects

Crater growth is known to be *non-proportional*: the diameter to depth ratio rapidly evolves during excavation, as maximal transient depth is achieved well before maximal diameter [Schultz et al., 1981; Schultz, 1988; Schultz and Hermalyn, 2013; Schultz et al., 2013a]. This growth sequence is especially apparent when viewing an impact in a quarter-space setup. Although craters do not exhibit static aspect ratios during formation, within certain regimes (dependent on impactor size, velocity, gravity, and material properties), crater scaling relations for the final diameter-depth ratio are proportional, consistent with point-source scaling [Holsapple and Schmidt, 1987].

However, scaling for final crater dimensions transitions into a non-proportional regime, i.e., the aspect ratio of the transient crater is not independent of impactor size, when the time taken to fully couple a projectile’s momentum and energy to the target, t_c , increases beyond a critical value [Schultz, 1988]. This effect is clearly visualized using impactors of differing material properties. In porous ice targets, projectile effects on crater formation are evident in Fig. 3.2, which contrasts low-density polyethylene ($\rho = 0.92 \text{ g/cm}^3$) with aluminum ($\rho = 2.78 \text{ g/cm}^3$) impactors for the 90° impact angle case. The larger impedance contrast (density times sound speed) between aluminum and porous ice results in a longer coupling time, t_c :

$$t_c = \eta t_p \tag{3.1}$$

where

$$t_p = \frac{2r}{v} \tag{3.2}$$

$$\eta = \left(\frac{\rho_t}{\rho_p}\right)^{1/3} \left(\frac{c_p \rho_p}{c_t \rho_t}\right) \left[1 + \left(\frac{\rho_p}{\rho_t}\right)^{1/2}\right] \quad (3.3)$$

Here ρ_t and ρ_p refer to the target and projectile densities, c_t and c_p indicate material sound speeds, r is projectile radius, and v is impact velocity.

Using material values for aluminum and polyethylene (see Table 3.3), t_c is approximately 6.6 times greater for the aluminum impactor. This permits deeper target penetration and the preservation of the projectile’s signature throughout the impact process, as evidenced by the formation of a central ‘pit’. The onset of occurrence for this pit feature is bracketed by the impedance conditions of polyethylene and aluminum projectiles, thereby providing constraints on the initial conditions necessary to produce pits. The implications of this process for planetary-scale events are discussed in Section 3.7.

Note also that, unlike the polyethylene case, the aluminum projectile (at impact speeds available) actually outpaces the compaction wave set up in the target at early times. The serrated edges of the crater’s interior, which are more prominent for the more massive aluminum impactor, appear to arise from large dynamic pressures within the transient cavity, which are sufficient to mobilize the weak porous snow. This vapor-driven process was also reproduced in computational results (Section 5).

3.3 *Impact Angle Effects*

Impact angle effects on transient crater growth are shown in Fig. 3.3, which contrasts the excavation of ice by aluminum projectiles at 90°, 60°, and 30° from the horizontal. Depth-diameter ratios become increasingly shallow at lower incidence angles, an effect well-characterized in quartz sand targets [Gault and Wedekind, 1978]. At these intermediate crater formation times (66, 200, and 1000 μ s), the ‘footprint’ of the projectile remains but becomes slightly less prominent and observably migrates downrange with decreasing impact angle. However, the flow-field center for oblique

impacts also migrates downrange during excavation [Anderson and Schultz, 2006; Hermalyn and Schultz, 2011], so that the final pit location remains near-center.

Results for final diameter/depth ratios as a function of impact angle, for both aluminum and polyethylene projectiles, are shown in Fig. 3.4. Note that, due to settling of particulate material at the end stages of crater growth, final crater dimensions differ from maximal transient crater dimensions [Schultz et al., 2005]. Depth-diameter results for aluminum impacting sand are consistent with previous experiments into half space targets [Gault and Wedekind, 1978]; this supports the validity of the quarter-space approach. In a manner similar to observed trends in sand targets, crater profiles for ice become significantly shallower at 30° , but there is negligible change in D/d between 90° and 60° . Shallowing out of the final crater profile at 30° is especially pronounced for the low-density polyethylene impactors. This effect is likely to be related not only to the reduced penetration time, t_c , associated with lower-impedance polyethylene, but also to its lower vaporization temperature. As shown in Fig. 3.4b and by Schultz et al. [2005], decreasing projectile to target density ratios, ρ_p/ρ_t , are associated with larger D/d ratios. Despite near-identical ρ_p/ρ_t ratios for 90° aluminum into sand and polyethylene into ice, the final crater dimensions for these two cases differ significantly. Hence, additional material properties, including the way in which the projectile fails, melts, and/or vaporizes, can play an important role in determining final crater dimensions.

Note also from Fig. 3.3 that, as impact angle decreases down to 30° , vaporization is enhanced, consistent with results for half space experiments into porous ice targets [Bruck Syal and Schultz, 2014] and previous studies of impact vaporization [Schultz, 1996]. In porous targets, the jetting phase, which is the earliest and hottest component of vaporization, is partially contained by the early-time penetration cavity before being released. The subsequent expression of impact-generated vapor at the surface differs markedly from nonporous ice targets [Bruck Syal and Schultz, 2014].

3.3 Ice and Sand Comparison

A comparison between 90° impacts of aluminum projectiles into pure porous ice, a porous ice and quartz sand mix (rock mass fraction of 0.42, typical of estimated icy satellite regolith compositions), and pure quartz sand is shown in Fig. 3.5. The signature of projectile penetration, evident in pure ice, is suppressed in the ice-rock mixture, due to a higher target density and the resulting decrease in coupling time, t_c . Additionally, although ejecta velocities were not quantified in this study, the suppression of ejecta with increasing porous ice content, presumably due to energetic losses from target compression, was qualitatively evident in the evolution of the ejecta curtains.

Transient crater depths and diameters were measured over the duration of crater formation time for the impacts shown in Fig. 3.5. Overlapping observations from cameras with 2, 8, 24.37, and 66.67 μs time resolution permit probing a time range that spans five orders of magnitude and reduce timing errors to $< 2 \mu\text{s}$. Diameters were measured relative to the pre-impact surface, so that the final diameter does not represent the rim-to-rim distance but the distance between crater walls at the undisturbed target height. Results for diameter and depth are shown in Fig. 3.6; spatial and time scales for the plots have been non-dimensionalized using impactor radius r and impact velocity v .

The earliest-time observations of crater growth are most easily resolved in the ice target, as impacts into sand produce substantial blackbody radiation through shock and frictional heating of sand grains. This radiation obscures much of the earliest transient cavity formation, as shown in Fig. 3.7 that compares ice (left) and sand (right) at $t = 8 \mu\text{s}$ after impact. While ice targets also undergo substantial irreversible heating, a large fraction of this energy is partitioned into vaporization (and melting) of the ice [Bruck Syal and Schultz, 2014] (Chapter 2). Hence, as noted in previous experimental work [Ernst and Schultz, 2007; Schultz et al., 2007], cold

and volatile-rich targets suppress the total luminosity of the impact flash. This makes icy targets particularly useful for evaluating early-time depth and diameter trends, when the commonly used point source approximation for crater growth [Holsapple and Schmidt, 1987] does not hold. Constraints on early-time crater growth can also be obtained using 3-D Particle Imaging Velocimetry methods [Hermalyn and Schultz, 2011; Anderson and Schultz, 2006], which are less intrusive than the quarter-space approach taken here but difficult to implement over the entirety of crater formation.

3.4 Compaction Wave Propagation

Viewing crater formation in porous particulate targets through a quarter-space geometry reveals not only the evolving transient crater dimensions but also permits the resolution of a propagating compaction wave. The compaction wave is visible as a distinct density contrast within the target’s subsurface and provides additional information on the effects of porosity on crater development. Compacted regions are discernible, for example, in Fig. 3.5, as subtle hemispherically-shaped areas of compressed target material, which surround the transient craters. The velocity and spatial extent of this visibly compacted region of material also provide a convenient metric for benchmarking numerical models of target porosity. Comparisons with results from the shock physics code CTH [McGlaun et al., 1990], described in Section 5, show that the shock front (visible as a pressure contrast) is coincident with the compaction wave (visible as a density contrast) at early times. At later times, however, the shock wave (and eventual acoustic wave) detaches from the compaction wave.

Propagated distances for the leading edges of compaction waves in porous ice, ice-sand mix, and quartz sand (illustrated in Fig. 3.5) are shown in Fig. 3.8. Following the convention of Fig. 3.6 for transient crater dimensions, the vertical distance of the compaction wave front from the impact point, d_c , is normalized by impactor radius r , while the timescale utilizes the dimensionless penetration time vt/r . No-

tably, all three target types adhere to a power law with similar slopes: $n_{ice} \sim 0.30$; $n_{mix} \sim 0.35$; $n_{sand} \sim 0.36$, each fit by a least-squares solution. These power law exponents are much too small to describe expected declines in peak particle velocity, u_p , which, by momentum conservation, should scale as $\sim r^{-2}$; results from large-scale explosions demonstrate that $u_p \sim r^{-1.87}$ [Perret and Bass, 1975]. Hence, the compaction wave front, at later times, is not measuring shock or particle velocity, but another characteristic of target deformation associated with porous compaction. The compaction wave, defined as a “propagating disturbance of the solid volume fraction of the granular material,” [Powers et al., 1989], can travel at both supersonic and subsonic velocities, depending upon shock conditions. Compaction wave velocities transition between these two regimes in the experiments described here; wave speeds decay from several hundreds of meters per second (exceeding target acoustic velocities) to subsonic values of just 60 m/s for the ice and ice-sand mix cases and a mere 30 m/s for the sand case.

Increased porosity is associated with compaction waves traveling to larger radial distances, as the projectile penetrates to greater depths during the initial coupling of momentum and energy. Early-time ($t < 50 \mu\text{s}$) velocities for ice and ice-sand mix compaction waves are similar: ~ 370 m/s. For sand, early measured compaction wave velocities are ~ 250 m/s. The ice-sand mix, despite a significant rock mass fraction of 0.42, compacts more similarly to the pure porous ice. This is particularly apparent at later times, when the sand velocities begin to decay at a greater rate, while the compaction wave extent of the ice and ice-sand targets converge (Fig. 3.8).

3.5 Computational Comparison

3.5 Numerical Approach

Observations of crater formation in quarter-space geometries can serve as a convenient benchmarking tool for numerical models. Previous efforts to compare experiments

directly with computational results have focused on impacts into liquid H₂O and aluminum [Pierazzo et al., 2008]. Here we use the CTH shock physics code [McGlaun et al., 1990] to model laboratory-scale impacts into porous ice, for comparison with our experimental results.

We focused on vertical impacts by aluminum and polyethylene projectiles into pure porous ice (No. 121004 and 121008 in Table 3.1), in order to determine whether the developing transient craters exhibit morphology differences similar to those seen in experiments (see Fig. 3.2). As an initial test, these events are simulated in a 2-D cylindrical geometry, using a flat mesh. Usage of cylindrical geometry is computationally efficient but often introduces artifacts along the problem’s axis of symmetry. Hence, caution is merited when interpreting these results. In future work, these problems will be simulated in a full 3-D geometry, which will require careful selection of Adaptive Mesh Refinement indicators, in order to resolve the transient crater accurately [Crawford, 1999].

In the case of under-dense, porous ice targets, the accuracy of the porosity model is critically important. While the development of new porosity models for usage in shock physics codes is ongoing [Wünnemann et al., 2006; Collins et al., 2011; Güldemeister et al., 2013], as an initial test, we use the $P - \alpha$ porosity model [Herrmann, 1969; Kerley, 1992], which is straightforward to implement in a wide range of hydrodynamic codes. Library coefficients from the ANEOS equations of state [Thompson and Lauson, 1972] were used for water ice and aluminum, while polyethylene was modeled as ANEOS polystyrene with the initial density decreased to $\rho_{poly} = 0.92 \text{ g/cm}^3$. Target strength was included using a combination of a modest 0.1 GPa crush pressure [Stewart and Ahrens, 2004] for the $P - \alpha$ input and a pressure-dependent yield surface with thermal softening and density degradation (referred to as ‘Geo-yield’ in Table 3.2). Key parameters used in the simulations are listed in Table 3.2.

3.5 Numerical Results

Early-time ($t \sim 20 \mu\text{s}$) results for both polyethylene (top) and aluminum (bottom) impactors are shown in Fig. 3.9, which contrasts quarter-space images with numerically-derived density and pressure fields. Note that projectile sizes differ slightly, due to experimental constraints: $r_{al} = 0.2381 \text{ cm}$ and $r_{poly} = 0.3175 \text{ cm}$. At this stage, the shock wave front (edge of the pressure contrast) is coincident with compaction wave front; the compaction wave, represented by the red annulus of dense material in the numerical density plots, is visible as a hemispherical region of compressed material beneath the transient craters. The extent of the compaction waves compare well to the experiments; hence the $P - \alpha$ model is accurately capturing the subsurface compression. The differences in coupling time and penetration depth between the deeply-penetrating aluminum and the lower-density polyethylene are reflected in the density plots, which show a more elongated transient cavity for the aluminum impactor.

Time sequences of results for polyethylene and aluminum impactors are plotted in Fig. 3.10 and Fig. 3.11. The compacted region beneath the target, which is visible in the experiments throughout the time sequence, is only discernible in the CTH results at earlier times (light red hemisphere). This implies that the numerical simulations, with the relatively simple model parameters provided, may under-estimate the degree of post-shock, permanent compaction. Overall, the simulations slightly over-estimate the dimensions of the polyethylene crater while underestimating the dimensions of the aluminum crater. However, details of the differing morphologies are consistent with experiments: smaller diameter-depth ratios, the formation of a pronounced central pit, and enhanced ‘serration’ on the interior of the crater are all observed for aluminum projectiles, both experimentally and numerically. The interpretation of any feature near the central axis, however, should be made with caution, because the computational mesh in that region is prone to numerical artifacts (e.g., high-density

material along the axis at late times in Fig. 3.10 and Fig. 3.11 is not real).

The elongated dimensions seen in the numerical results for transient crater development, which are notably non-hemispherical in shape, are consistent with previous work on highly underdense targets, e.g. ground perlite [Schultz et al., 2007], which contains both micro- and macro-porosity ($\rho \sim 0.3 \text{ g/cm}^3$). Impacts into such target materials are described by deep effective depths of burst, longer penetration times, and non-hemispherical transient cavities. The serrated pattern on the interior of the craters appears to be driven by strong dynamic pressures within the transient crater, as material is vaporized and its expansion is partly contained by the early-time cavity. As the porous ice is relatively weak and easily volatilized, the role of impact-generated vapor in enhancing post-impact erosion of transient crater walls is pronounced. This process is more significant in the aluminum projectile case, due to its higher initial kinetic energy and the early containment of vaporized material within an elongated penetration funnel. Recent work shows that, even at laboratory velocities, impacts into porous water ice produce appreciable amounts of vapor, equal to a few or more projectile masses (depending upon the impact angle) [Bruck Syal and Schultz, 2014] (Chapter 2). The erosive patterns along the edges of the impact crater support the possible contribution of vaporized material in shaping the morphology of impact craters.

3.6 Discussion

The use of porous ice targets serves dual purposes: (1) it allows the visualization of gravity-controlled transient crater growth in water ice, which is not possible for nonporous, crystalline water ice targets (strength-controlled regime); (2) it permits probing the effects of porosity on crater development for a geologically relevant material. Porous water ice, in particular, is useful for constraining porosity effects on transient crater growth during the earliest times of crater formation, since it sup-

presses the luminosity of the impact flash (Fig. 3.7).

Time-resolved diameters and depths for impacts into porous ice, quartz sand, and an ice-sand mix reveal non-proportional growth in all three targets, as final diameters are reached well after final depths are achieved (Fig. 3.6); note that the last plotted data points correspond to final dimensions of each crater. Growth rates monotonically decay, except for ice target depth, which experiences a temporary growth rate increase when the projectile-driven central pit becomes visible at $vt/r \sim 5 \times 10^2$. Point source scaling predicts that, for intermediate times (approximately, $1 < vt/r < 10^4$), crater growth should adhere to a simple power-law, dependent on the parameter μ [Holsapple and Schmidt, 1987; Housen and Holsapple, 2011]. The value of μ describes the relative importance of momentum and energy in determining final crater volume; $\mu = 1/3$ corresponds to momentum scaling and $\mu = 2/3$ corresponds to energy scaling.

The ice and sand-ice targets, however, do not adhere to power law scaling throughout the range of times captured by the experiments. At early times, both crater diameter and depth grow rapidly before converging to a linear trend, e.g., [Schultz and Hermalyn, 2013]. This effect is especially apparent for crater diameter relations, which depart from power-law scaling for longer durations of time. However, depth also exhibits departures from power-law scaling out to relatively large radial distances and times. Although transient crater dimensions from the sand target were not measurable at the earliest times, silicate targets have previously been demonstrated to exhibit departures from the point-source power law [Schultz and Hermalyn, 2013]. While point source scaling allows for an “early-time” regime, where projectile effects are important, the spatial and temporal extent of this regime is generally inferred to be much smaller than the results reported here.

Eventually, the ice and ice-sand mixture cases converge to reasonable power-law slopes. Interestingly, the diameter of the ice-sand mixture converges to a power law at significantly later times than pure ice. This may be related to the way in

which porosity is rapidly crushed out directly beneath the pure ice target. While porosity is also crushed out under the mixed target, this process occurs over a slightly longer time period, as the target sound speed is lower (visible from the propagating compaction wave, discussed in Section 4). Consequently, the decay of crater diameter growth rates occurs over a longer time period for the ice-sand mixture than for pure ice. This ice-sand mixture example, which represents a common surface composition within the solar system, highlights the need to understand how and when point source scaling assumptions may not apply to an impact event. Projectile and target material properties can introduce many competing effects that interfere with the development of a growth pattern that is well-described by a simple power law.

Following the arguments outlined in Holsapple and Schmidt [1987], the power law slope should correspond to $\frac{\mu}{1+\mu}$, as the evolving dimensions are described by:

$$\frac{d}{r} = k \left(\frac{Vt}{r} \right)^{\frac{\mu}{1+\mu}} \left(\frac{\rho_p}{\rho_t} \right)^{\frac{\nu}{1+\mu}} \quad (3.4)$$

Here d is the transient crater depth, r is impactor radius, k is an empirically-determined constant, V is impact velocity, ρ_p and ρ_t are impactor and target density, respectively, and ν is a material parameter usually set to 1/3 [Holsapple and Schmidt, 1987; Holsapple, 1993].

From the results shown in Fig. 3.6, separate μ values were calculated for depth and diameter, using crater measurements obtained at intermediate times (after growth appeared to converge to a power law but before it began to be arrested by gravity). These times differ slightly by material type, e.g., transient diameter for the ice-sand mixture does not converge to a point source-like power law until $vt/r \sim 4 \times 10^4$. For transient diameters, sand was described by $\mu_{sand}^D \sim 0.43$, consistent with previous results for soil-like targets [Holsapple, 1993; Housen and Holsapple, 2011]. The ice-sand mixture was described by $\mu_{mix}^D \sim 0.38$, while pure ice was fit by $\mu_{ice}^D \sim 0.35$.

These results suggest that, with increasing ice content, diameter scaling becomes increasingly skewed towards being momentum-controlled. The opposite trend was seen in crater depth scaling; the best-fitting exponents increased from $\mu_{sand}^d \sim 0.35$, to $\mu_{mix}^d \sim 0.42$, to $\mu_{ice}^d \sim 0.47$, implying that crater depth in ice-enriched targets will deviate towards energy-controlled scaling. These trends may be a consequence of the higher porosity and enhanced role for compaction within the icier targets. Other work has also emphasized the distinction between depth scaling and diameter scaling, e.g., [Schultz and Hermalyn, 2013; Schultz et al., 2013a].

These results indicate that conventional main-stage growth relations, as described in Holsapple and Schmidt [1987], are not easily applied to porous materials. The early-stage growth regime, during which depth and diameter evolution do not adhere to a simple power law, lasts much longer in porous targets (see Fig. 3.6). Additionally, differences between the power law slopes for depth and diameter highlight the phenomenon of non-proportional growth, in which the diameter continues to increase well after maximum depth is attained [Schultz, 1988; Schultz and Hermalyn, 2013]. While this process also occurs in lower-porosity sand, it is exaggerated in the higher-porosity snow and snow-sand mixes.

Recent work on impacts into highly porous materials suggests that compaction-dominated cratering, which requires porosities exceeding 30-40%, resides completely outside of the gravity regime, as final crater diameter will depend only on the crush strength of the material [Housen and Holsapple, 2012]. The porous ice used in these experiments lay right at the edge of this transition porosity (40%). Based on comparisons of final diameters and depths to quartz sand in Fig. 3.6, however, the ice and ice-sand targets appear to lay within the same scaling regime as the quartz sand (gravity-controlled). The principal difference between the sand and higher porosity cases is the final diameter-depth ratios: the elongated coupling time associated with porous targets produce deeper and narrower craters (lower D/d ratios; see Fig.

3.4). Although additional experiments at a range of impact velocities are necessary to place constraints on the scaling relations, these initial results suggest that low-strength porous materials may be best-described by gravity scaling coupled with an early-time growth stage of longer duration. The early time regime will be of even longer (relative) duration for larger-scale impacts, as the coupling timescale, t_c , increases at a greater rate than the timescale of crater formation, t_g [Schultz, 1988, 1992]. The sequence in which crater compaction and excavation occurs ultimately has implications for how ejecta is emplaced at porous surfaces.

Impact angle and density ratio effects on diameter-depth relations provide insight to observations of craters at ice, ice-rock, and rock surfaces. As expected, craters become shallower at oblique angles and for smaller projectile-target density ratios (Fig. 3.4). Enhancements of D/d ratios at lower impact angles are even more pronounced for lower-impedance projectiles; this phenomenon was previously noted in sand targets but is here extended to ice as well. This observation is particularly relevant for cometary impacts ($\rho \sim 0.4\text{-}0.6 \text{ g/cm}^3$), which will dominate impact processing within the outer solar system. As the impact angle probability distribution function predicts that one out of every four events will impact at an angle of 30° or less [Gilbert, 1893; Shoemaker, 1962], the extreme enhancements seen in D/d for $\rho=0.92 \text{ g/cm}^3$ projectiles impacting at 30° may explain some anomalous D/d ratios. Ultimately, these relations can serve as a guide for the physical reasons behind observed variations in D/d at different surfaces.

Interpretation of the measured diameter-depth relations for porous ice and porous ice-sand mixes must consider the additional role of moderately high porosity within these experiments (40%). While inclusion of porosity in our targets was necessary for visualization of crater growth within the gravity-controlled regime and provides insight on the type of compaction likely to occur at highly porous comets or asteroids, it clouds the direct application of these D/d ratios to other, less porous surfaces, such

as icy satellite regoliths. Increased porosity produces smaller D/d ratios, due to longer penetration times (deeper initial burrowing of projectile), so that cometary impacts with a less-porous icy surface would produce significantly larger D/d ratios than the ones reported here. Additionally, these values do not account for post-impact modification or viscous relaxation. However, exploration of a relatively large range of impactor-target density ratios (Fig. 3.4b) illustrates the central role of these material properties in driving final dimensions of craters formed in ice. The low density of cometary impactors provides, at a minimum, a partial explanation for the enhancements in D/d ratios at the icy satellites [Schenk, 1989].

Matching numerical simulations successfully captured some of the trends seen experimentally, including smaller diameter-depth ratios for higher impedance impactors, possible onset of non-proportional scaling, early-time compaction waves, and the mobilization of crater wall material by vapor back pressures in the transient cavity. However, the lens of compacted material seen in the experiments was not retained throughout the numerical simulations, while transient crater sizes were too large for polyethylene but too small for aluminum. This discrepancy may reflect the way in which projectiles fail during impact: polyethylene is likely to fully vaporize, while aluminum will likely experience plastic deformation and/or brittle failure. Overall, the effects of differing impedance matching conditions are well-simulated by the calculations, but additional work on benchmarking of porosity and strength models is necessary to tie the computational approach more tightly to observations of crater growth in porous targets.

3.7 Implications: Central Pit Formation

It is tempting to make a connection between the laboratory-scale pit formation discussed in Section 3.1 and the presence of central pit craters at icy satellites Ganymede and Callisto [Schenk, 1993]. The pit-crater diameter aspect ratio is ~ 0.25 , similar to

Ganymede central pit craters ~ 80 km in diameter [Schenk, 1993]. However, invoking a similar process at planetary scales must first be rigorously justified by crater scaling relationships, as the gravity, velocity, and impactor size conditions differ greatly from experimental conditions.

To test the viability of this process at planetary scales, we introduce a dimensionless parameter, π_t :

$$\pi_t \sim \frac{t_c}{t_g} \quad (3.5)$$

where

$$t_g \sim \left(\frac{D_{at}}{g} \right)^{1/2} \quad (3.6)$$

Here t_g is the gravity-controlled timescale of formation, D_{at} is the apparent transient diameter of the crater (before modification), g is surface gravity, and π_t represents the ratio between shock transit time and gravitational arrest of the process. Porous ice, like quartz sand [Gault and Wedekind, 1977], is very low strength and, hence, these experiments reside within the gravity regime. From laboratory scales ($D_{at} \sim 20$ cm, $g = 980$ cm/s²) we know that the critical π_t lies somewhere between the polyethylene and aluminum cases:

$$3.0 \times 10^{-4} < \pi_t^{cr} < 2.0 \times 10^{-3} \quad (3.7)$$

Using these values, we can extrapolate to the conditions at Ganymede, investigating how the corresponding critical D_{at} (applying a modification factor) compares to the observed transition to pitted craters, near 40 km [Schenk, 1993]. From Eq. (3.1-3.6) it can be shown, for a given π_t^{cr} :

$$D_{at} = \frac{4g\eta^2 r^2}{v^2 \pi_t^{cr}} \quad (3.8)$$

This describes the apparent transient crater diameter at which non-proportional scaling should become important. If the experiments are applicable to Ganymede, calculating D_{at} using the π_t^{cr} values from Eq. 10 (and applying a modification factor) should bracket the 40 km transition.

Since the impactor radius, r , is interdependent with D_{at} , we require a scaling relation. In the gravity regime we define, following Holsapple and Schmidt [1982]:

$$\pi_D = 0.62 \left(\frac{\rho_t}{\rho_p} \right)^{1/3} \frac{D_{at}}{r} \quad (3.9)$$

$$\pi_2 = \frac{3.22gr}{v^2} \quad (3.10)$$

where π_D is a measure of the normalized crater diameter and π_2 is the inverse Froude number. The two are related using the experimentally-determined constants C_D and β :

$$\pi_D = C_D \pi_2^{-\beta} \quad (3.11)$$

If we define:

$$C_1 = 0.62 \left(\frac{\rho_t}{\rho_p} \right)^{1/3} \quad (3.12)$$

$$C_2 = \frac{3.22g}{v^2} \quad (3.13)$$

It can be shown:

$$r = C_3 D_{at}^{\frac{1}{1-\beta}} \quad (3.14)$$

where

$$C_3 = \left(\frac{C_1 C_2^\beta}{C_D} \right)^{\frac{1}{1-\beta}} \quad (3.15)$$

Now we may substitute back into Eq. 11. With $\beta=0.16$ we obtain:

$$D_{at} = \left(\frac{v^2 \pi_t^{cr}}{4g\eta^2 C_3} \right)^{0.724} \quad (3.16)$$

While there is some uncertainty in the impedance match conditions between cometary impactors and the surface of Ganymede, nominally, we can use a typical comet density of $\rho_p = 0.6 \text{ g/cm}^3$ and a slightly higher density for Ganymede's ice-rock mix regolith: $\rho_p = 1.1 \text{ g/cm}^3$. The sound speed contrast is assumed to be of order unity, as both impactor and target are particulate ice and rock mixes. With $C_D = 2.5$, $g = 143 \text{ cm/s}^2$, and $v_{rms} = 18 \text{ km/s}$ (see Table 3 for complete list of values) and using the range of π_t^{cr} found experimentally, we obtain:

$$7.3 \text{ km} < D_{at} < 48 \text{ km} \quad (3.17)$$

Note that these are the apparent transient crater diameters. Applying a reasonable modification factor for this size range (~ 1.6) [Holsapple, 1993] yields final crater diameters:

$$12 \text{ km} < D_{at} < 77 \text{ km} \quad (3.18)$$

This result, which does indeed bracket the central pit transition diameter at Ganymede, suggests that non-proportional scaling, which has been well characterized in quartz sand targets and is extended here to low-strength ice targets, may contribute to the diverse crater morphologies observed at icy satellites. Furthermore, the functional dependence of transition diameter, D_{at} , described in Eq. (3.16) can be used to test whether the predicted gravity and velocity scaling is consistent with the onset of central pit craters at Callisto, Mars, and the Moon. We consolidate the material properties in Eq. (3.16) into a simplified impedance factor, C_i , adjusting for

surfaces dominated by asteroidal impacts (here, Mars and the Moon) with $C_i^a=0.305$, from the nominal $C_i^c=1.0$ for cometary impactors. Transition diameter, D_t , should then scale as:

$$D_t \sim C_i \left(\frac{v^{1.723}}{g^{0.862}} \right) \quad (3.19)$$

Note that, since the amount of post-formation modification will increase at greater sizes, this relation may under-predict transition diameters for very large values of $v^{1.723}g^{-0.862}$. Observed transition diameters for each planetary surface [Schultz, 1988; Schenk, 1993] are well-fit by this gravity-velocity relation; Fig. 3.12 b depicts a linear least-squares solution for Eq. (3.19) when applied to these four bodies. This scaling relation provides a superior fit to the conventional g^{-1} scaling used for central pit transition diameter, shown in Fig. 3.12 a.

An interesting case is dwarf planet Pluto (along with Charon), where very low impact speeds (~ 1.9 km/s) [Zahnle et al., 2003] will elongate shock coupling times. Pluto’s predicted transition diameter, shown on Fig. 3.12 b, is similar to Mars: ~ 4 km. Imaging by the New Horizons spacecraft during its 2015 close approach should reveal the first resolvable impact craters at Pluto [Young et al., 2008]. While Pluto’s material properties are unconstrained and may differ from the jovian satellites, we anticipate the presence of central pit craters and expect that the transition to pitted craters will occur at a smaller crater diameter than for Ganymede or Callisto.

Alternative models for central pit origins invoke processes related to impact melt generation in icy targets [Senft and Stewart, 2011; Bray et al., 2012; Elder et al., 2012] or differences in ice rheology at depth [Schenk, 2002]. The volatility of ice may play a secondary role in pit formation, but the occurrence of central pits within relatively dry regoliths (Moon, Mercury) argues that the onset of pit development is likely tied to a process unrelated to volatile abundances. Additionally, the lack of pit features in the case of the polyethylene impactor suggests that ice content alone is not sufficient to induce pit formation. In other words, pit development is not controlled by target

properties alone, but by a combination of projectile and target properties, impact velocity, and gravity. The consistency with which non-proportional scaling relations predict observed transition diameters (Fig. 3.12) indicates an important role for this process, independent of other contributing factors.

3.8 Conclusions

Time-resolved studies of impact crater formation in porous, icy targets reveal strong departures from point-source solutions for transient crater growth. This effect is magnified in icy targets, relative to quartz sand targets, due to the longer timescales for projectile energy and momentum coupling. The consequences of non-proportional growth are most clearly displayed under oblique impact conditions, where crater diameter-depth relations depend upon a combination of projectile and target properties and impact angle. These effects can lead to anomalously shallow craters for oblique, low-density impactors, with consequences for observed crater morphologies within the cometary impact-dominated outer solar system.

Accurate accounting of porosity, which controls cratering processes ranging from melt and vapor generation [Bruck Syal and Schultz, 2014] to ejecta velocity distributions [Hermalyn and Schultz, 2011], is an important aspect of numerical approaches to planetary impact problems. The quarter-space experiments described in this study are well-suited to benchmarking of shock physics codes, due to the cross-sectional view of the cratering process. Comparison of transient crater dimensions and compaction wave propagation to numerical results demonstrated good agreement at earlier times but some subtle differences at later times. Future work will apply these benchmarking methods to a wider array of strength, porosity, and equation of state models. These details are particularly important for the lower velocity impacts that dominate the impact processing of trans-Neptunian objects, many of which possess significant porosity [Brown, 2013].

While all craters are described by non-proportional growth, these experiments also documented the transition between proportional and non-proportional scaling for ice targets. This transition, which can be implicated in the development of central pit morphologies, has been well-investigated in silicate targets [Schultz, 1988], but this is the first such example for ice impacts. As demonstrated by projectiles of differing impedances, pit formation depends on a critical coupling timescale, which must be large, relative to the timescale of gravity-controlled growth.

Constraints from laboratory-scale results also allow predictions to be made for the onset of central pit morphologies at various planetary bodies. The conventional parameter for scaling the onset diameter is $\sim g^{-1}$, but we derive a more nuanced relation (Eq. 3.19), which captures the relative importance of velocity, gravity, and material properties for determining initiation of pit development. This relation is shown to fit a wide range of planets and satellites and predicts that craters at dwarf planets Pluto and Charon should exhibit central pits, transitioning to this crater type at smaller diameters than Ganymede and Callisto (Fig. 3.12). The arrival of the New Horizons spacecraft at the Pluto-Charon system in July 2015 will permit this hypothesis to be tested.

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Tables

Table 3.1: List of Experiments

Shot	Target	Projectile ^a	θ	v (km/s)	m_p (g)
121004	Ice ^b	Alum	90°	5.22	0.158
121005	Ice	Alum	60°	4.87	0.159
121006	Ice	Alum	30°	4.85	0.158
121007	Ice	Poly	30°	5.23	0.127
121008	Ice	Poly	90°	5.30	0.128
121009	Ice-Sand ^c	Alum	90°	5.18	0.158
121012	Sand ^d	Alum	90°	4.94	0.159

^aAlum: $r = 0.2381$ cm; Poly: $r = 0.3175$ cm

^b $\rho = 0.5$ g/cm³

^c $\rho = 0.7$ g/cm³; rock mass fraction: 0.42

^d $\rho = 2.78$ g/cm³

**Table 3.2: Computational Parameters:
Laboratory-scale Simulations**

Parameter	Value
T_{init}	260 K
v_{impact}	5.22-5.30 km/s
r_{al}	0.2381 cm
r_{poly}	0.3175 cm
ρ_{H_2O}	0.50 g/cm ³
Projectile EOS	ANEOS Aluminum ANEOS Polystyrene ^a
Target EOS	ANEOS Water-Ice
P- α Crush Strength ^b	0.1 GPa
Geo-yield Y	0.185 GPa
Geo-yield dY/dP	0.7
Geo-yield Poisson	0.33
Resolution	10 cpr ^c

^a $\rho_0 = 0.92$ g/cm³

^bStewart and Ahrens [2004]

^ccells per projectile radius

**Table 3.3: Key Parameters:
Central Pit Scaling**

Parameter	Value
c_{al}	5.38 km/s
c_{poly}	2.08 km/s
ρ_{al}	2.78 g/cm ³
ρ_{poly}	0.92 g/cm ³
g_{Gany}	143 cm/s ²
v_{Gany}	18 km/s
ρ_{comet}	0.60 g/cm ³
ρ_{Gany}	1.1 g/cm ³
$\rho_{asteroid}$	2.6 g/cm ³
ρ_{inner} ^a	2.6 g/cm ³
c_{comet}/c_{Gany}	~ 1
$c_{asteroid}/c_{inner}$	~ 1
C_D	2.5 ^b
β	0.16K11

^aMars and Moon regolith

^bKraus et al. [2011]

Figure Captions

Figure 3.1: Crater Isis, a 75 km-diameter central pit crater at Ganymede. Note that nearby, smaller craters, do not contain central pits; the minimum crater diameter for which central pits occur at Ganymede is ~ 40 km [Schenk, 1993]. This minimum size is referred to as the transition diameter, D_t , in the present work. While the occurrence of central pits at icy satellites such as Ganymede and Callisto is often attributed to the presence of water ice, similar features are observed at the Moon and Mercury. This suggests that the formation mechanism may not be controlled by target volatile abundances. Voyager 2 images 20636.38, 20640.33. Image credit: NASA/P.M. Schenk, D. Gwynn, J. Tutor.

Figure 3.2: Illustration of projectile impedance effects on transient crater growth and final morphology for 90° impacts into porous ice. The lower-impedance polyethylene impactor (top) produces the classical, bowl-shaped transient crater (although crater growth still occurs in a non-proportional sequence, with final depth achieved prior to final diameter). The higher-impedance aluminum impactor (bottom), however, couples its energy and momentum to the target over a longer duration of time, t_c (Eq. 3.1), leading to non-proportional scaling for the final crater dimensions. This effect is evidenced in the onset of a central ‘pit’ morphology, which records the signature of the projectile. The ramifications of this process for planetary scale impact events are discussed in section 6. Images are at 66, 200, and 1000 μs after impact (left to right).

Figure 3.3: Impact angle effects on transient crater growth for aluminum projectile impacts into porous ice targets. Red arrows indicate trajectory direction for impactor. At lower incidence angles, crater depth-diameter ratios become shallower and the signature of the projectile becomes less pronounced as the crater flow-field

center migrates downrange during excavation. Early-time vaporization is also enhanced at oblique angles, as seen in the first image from the 30° case.

Figure 3.4: Final crater diameter to depth ratios, D/d , plotted as a function of impact angle (a) and ratio of projectile and target densities (b). Impact angles in (b) are 90°, unless noted otherwise. As expected, craters become shallower (larger D/d) as impact angle decreases to 30° and as the density ratio ρ_p/ρ_t decreases. The difference between 90° and 30° impacts is greater for the low-density polyethylene impactors, which not only have smaller coupling times, t_c , but are also more easily vaporized upon impact. Both these effects likely contribute to enhanced shallowing out at lower angles. As seen in (b), although the density ratios between polyethylene into ice and aluminum into sand are similar, the D/d values for 90° impacts differ. Hence, the way in which the projectile fails, melts, and/or vaporizes contributes substantially to final crater morphology.

Figure 3.5: Comparison of vertical impacts into porous ice (top), porous ice and quartz sand mix (middle), and quartz sand (bottom) by aluminum projectiles, imaged at $t = 66, 200, 400$, and $1000 \mu s$ after impact. The inclusion of some sand in the mixture case (rock fraction: 0.42) sufficiently decreases coupling time, t_c , to suppress the formation of a central pit. The evolving depths and diameters of these craters, shown in Fig. 3.6, provide insight to the conditions under which non-proportional growth and scaling occur.

Figure 3.6: Crater diameter (a) and depth (b) as a function of time for 90° impacts of aluminum projectiles into three different particulate targets: porous ice, quartz sand, and ice-sand mixture (rock mass fraction of 0.42). Crater diameter, D , and depth, d , are normalized by impactor radius, r , and timescale is normalized by im-

pact velocity v and r . Non-proportional growth occurs in all three materials (crater diameter continues to grow after maximum depth is reached). Power law growth marked by the time intervals for which the points adhere to straight lines, shown by the least-squares fits (dotted lines). The ice and ice-sand mix cases deviate more strongly than sand from the power law growth predicted by point source solutions [Holsapple and Schmidt, 1987]; in these porous materials, the coupling of impactor energy and momentum occurs over longer time scales, so that main stage growth (intermediate between early and late stage growth) occupies a smaller fraction of total crater formation time; this is especially apparent in the diameter plots. The depth evolution of the ice target demonstrates a growth surge near $vt/r \sim 5 \times 10^2$, due to the central pit coming into view (see Fig. 3.5). For all materials, the very earliest time points should be interpreted with caution, as quarter-space-related obscuration may affect measurements.

Figure 3.7: Comparison of impacts into porous ice (left) and quartz sand (right) at $t = 8 \mu\text{s}$ after impact. Enhanced blackbody radiation, associated with the sand impact, impairs accurate resolution of the transient crater at early times. In contrast, the impact flash associated with the ice target is less intense and of shorter duration, allowing crater growth to be traced to earlier times.

Figure 3.8: Distance propagated by hemispherical compaction waves, d_c , normalized to the impactor radius r , as a function of normalized time variable vt/r . At the earliest times, the compaction wave is coincident with the shock wave (see Fig. 3.9); at later times, it represents the slow movement of a density contrast through the target subsurface. While waves propagate more efficiently through increasingly porous targets, all three cases are described by similar power law exponents, n . Compaction of ice-sand mixture is more similar to pure ice than to pure sand (early and late-time

velocities of ice and mixture are more aligned, and final radial distances converge).

Figure 3.9: Comparison between 90° impacts of polyethylene (top, $t = 20 \mu s$) and aluminum (bottom, $t = 22 \mu s$) into ice with computational results for density (center) and pressure (right). At these early times, the compaction wave front (density contrast) is still aligned with the shock wave front (pressure contrast). The longer target penetration times associated with aluminum contribute to vertically-elongated penetration cavities, both experimentally and numerically. While, for the aluminum impactor, the transient crater shape appears more hemispherical in the experimental results than in the numerical calculation, the lack of elongation in the experimental image may be related to the slight offset of the crater center from the acrylic window. At longer times (see Fig. 3.11) and also in other underdense targets, e.g., perlite [Schultz et al., 2007], substantial elongation of the transient crater is visible.

Figure 3.10: Time sequence of experimental results for 90° impact of polyethylene into ice, compared with numerically-calculated density plots. The compaction wave, while visible throughout all frames of the experiment, rapidly decays away in the numerical results, suggesting an underestimation of post-shock target packing. While the general shape of the transient craters are consistent, the experimental dimensions are appreciably smaller than the calculated dimensions. Note that center line artifacts are visible (higher density material along the $y = 0$ axis), due to the 2-D cylindrical geometry used in the model.

Figure 3.11: Time sequence of experimental results for 90° impact of aluminum into ice, compared with numerically-calculated density plots. Longer penetration times for aluminum than for polyethylene result in evidence for non-proportional scaling in both the experiment and model results. Signatures of the aluminum projectile’s ‘foot-

print', as it burrows into the target subsurface at a faster rate than the compaction wave, are evident even at early times in both image sequences. Additionally, high dynamic pressures associated with intense vaporization of ice appear to contribute to the development of the ragged/serrated patterns on the crater's interior. Ice's volatility and low strength permits efficient erosion by impact-generated vapor during early crater formation. Note that center line artifacts are visible (higher density material along the $y = 0$ axis), due to the 2-D cylindrical geometry used in the model. The elongated shape of the numerically simulated transient crater is consistent with impacts into underdense targets, such as ground perlite, which are described by a greater depth of burst [Schultz et al., 2007].

Figure 3.12: (a) Transition diameters plotted as a function of the conventional metric for central pit onset, the inverse of surface gravity (g^{-1}). The relation shown in (b), which incorporates the effects of impact velocity and material impedances, provides an improved fit to observations of central pit craters throughout the solar system. (b) Observed transition diameters for central pit crater morphologies are plotted for the Moon, Mars, Ganymede, and Callisto as a function of the derived scaling factor $C_i \left(\frac{v^{1.723}}{g^{0.862}} \right)$ from Eq. 3.19. As predicted by the scaling arguments outlined in Section 6, these bodies all adhere to a line. Least-squares fit is shown, with predicted transition diameter for dwarf planet Pluto plotted.

Figures

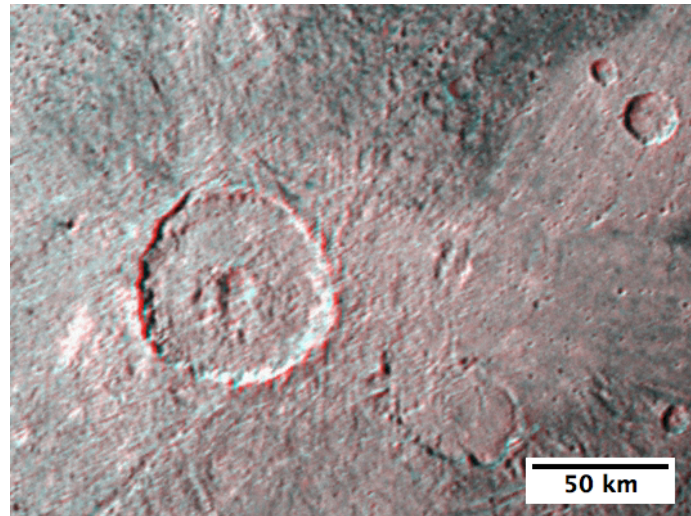


Figure 3.1

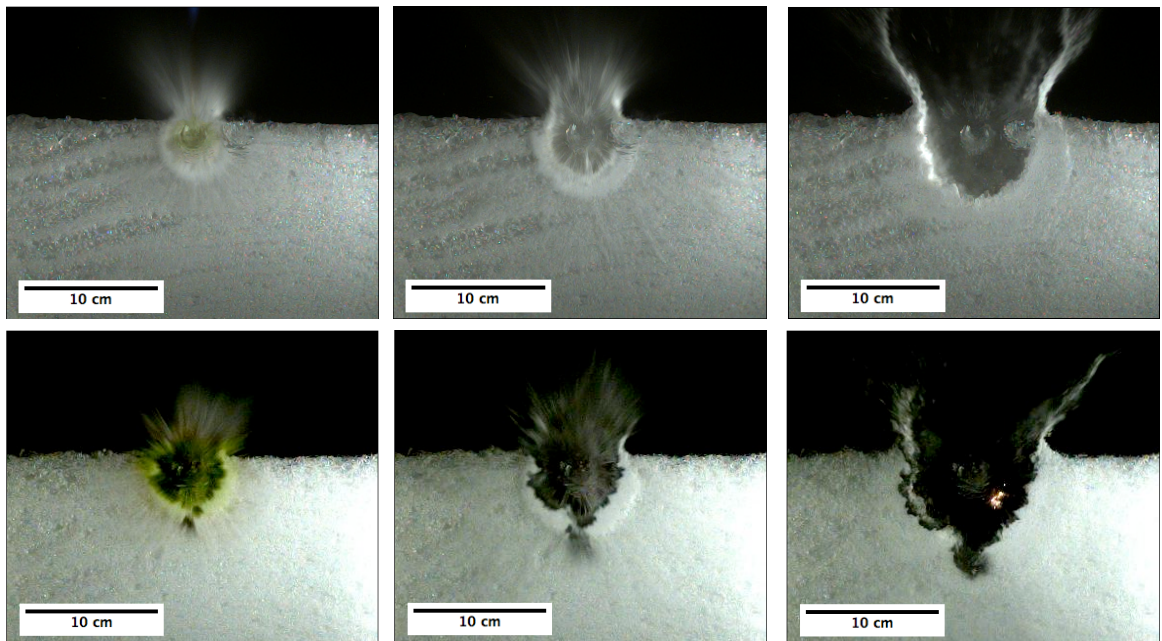


Figure 3.2

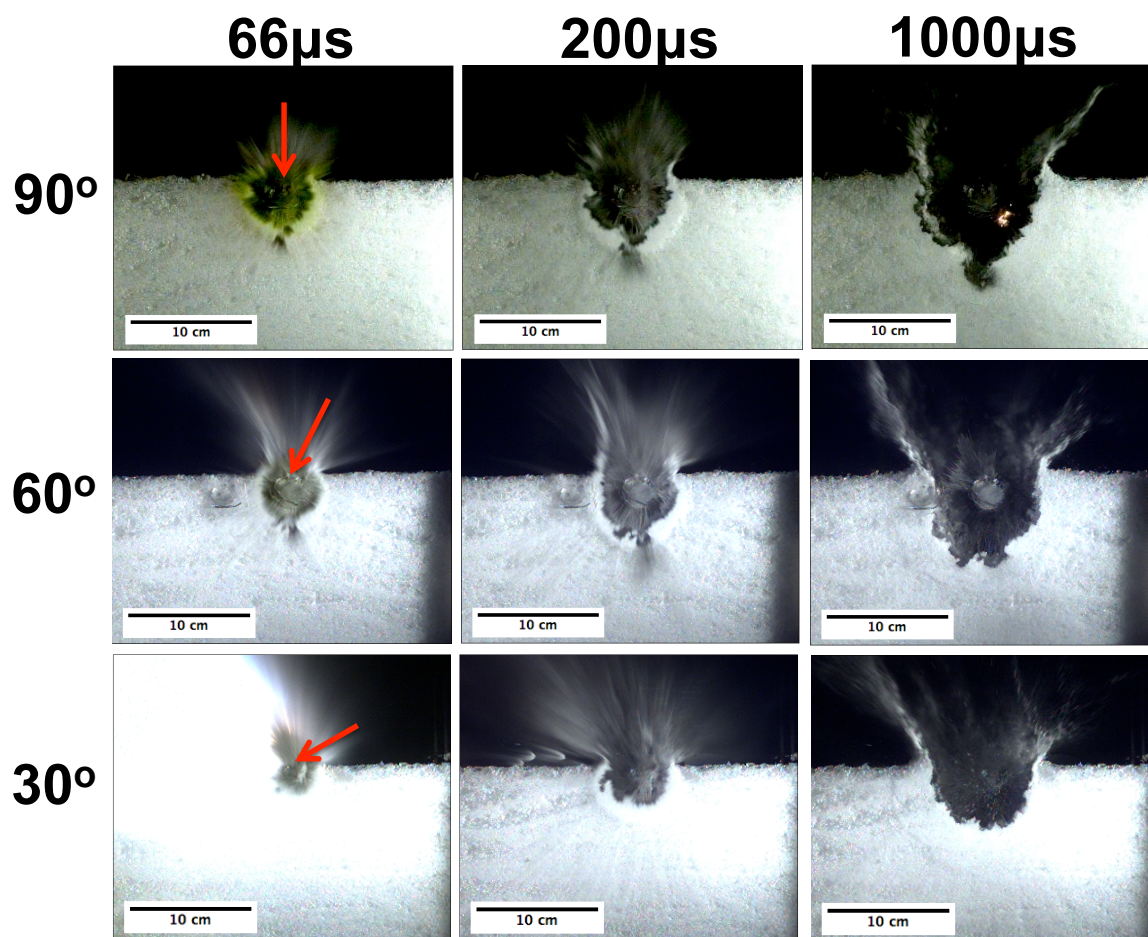


Figure 3.3

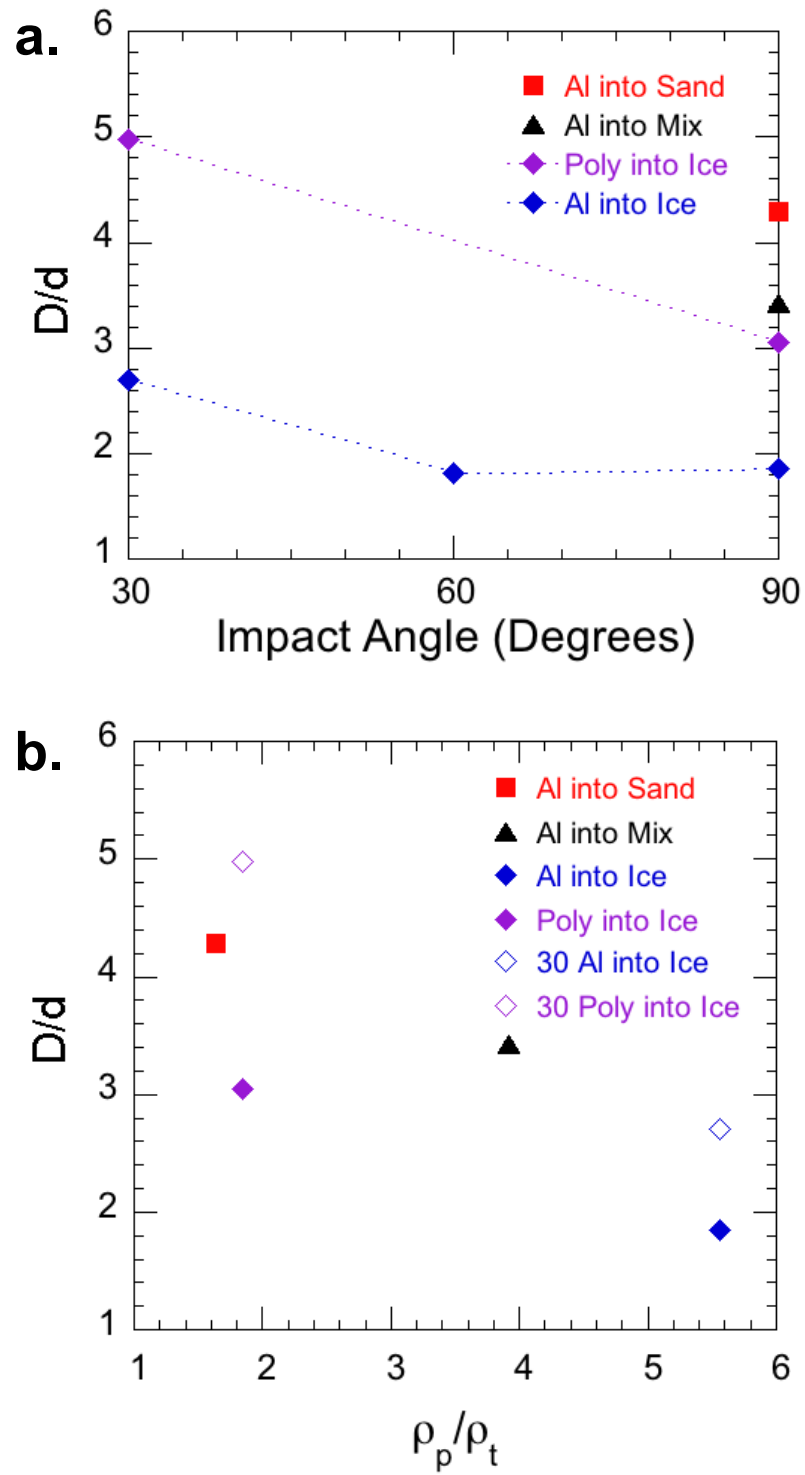


Figure 3.4

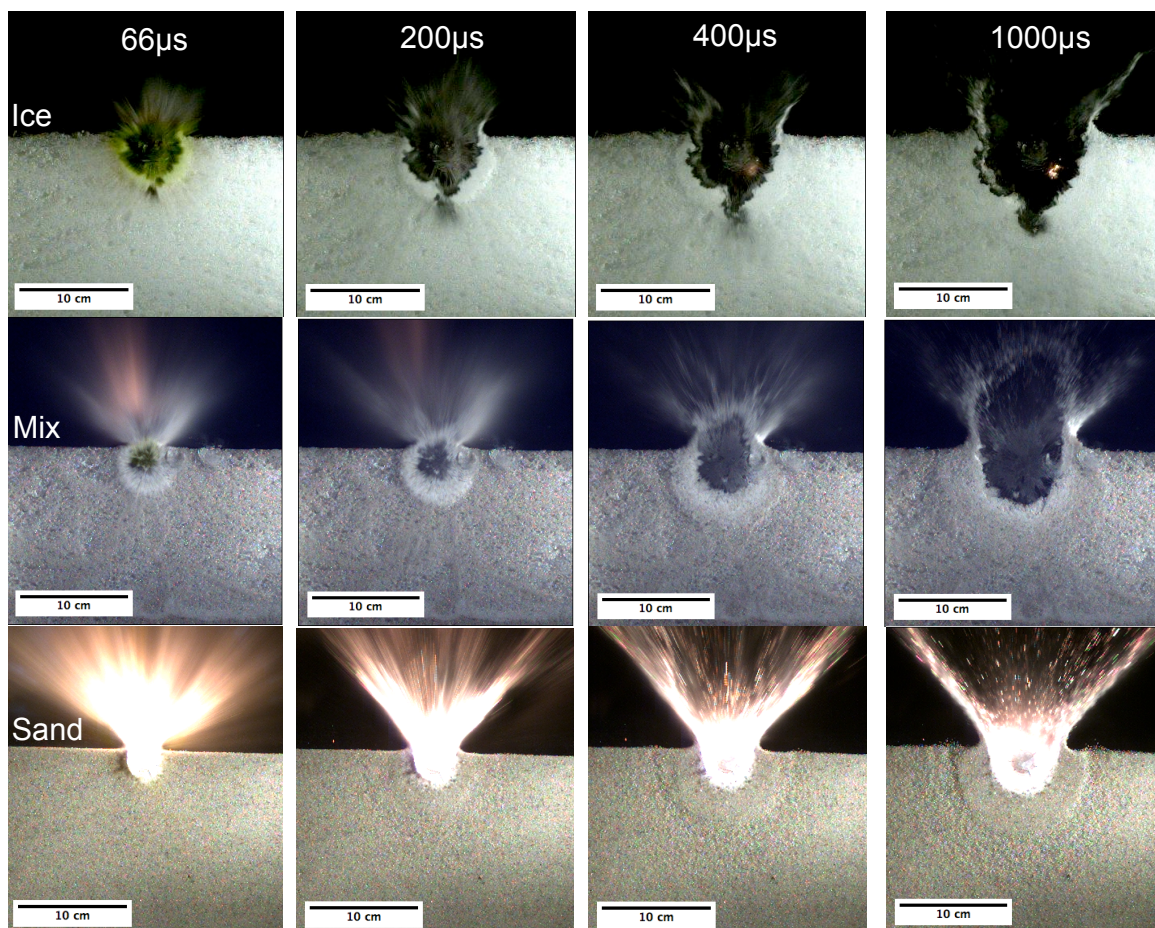


Figure 3.5

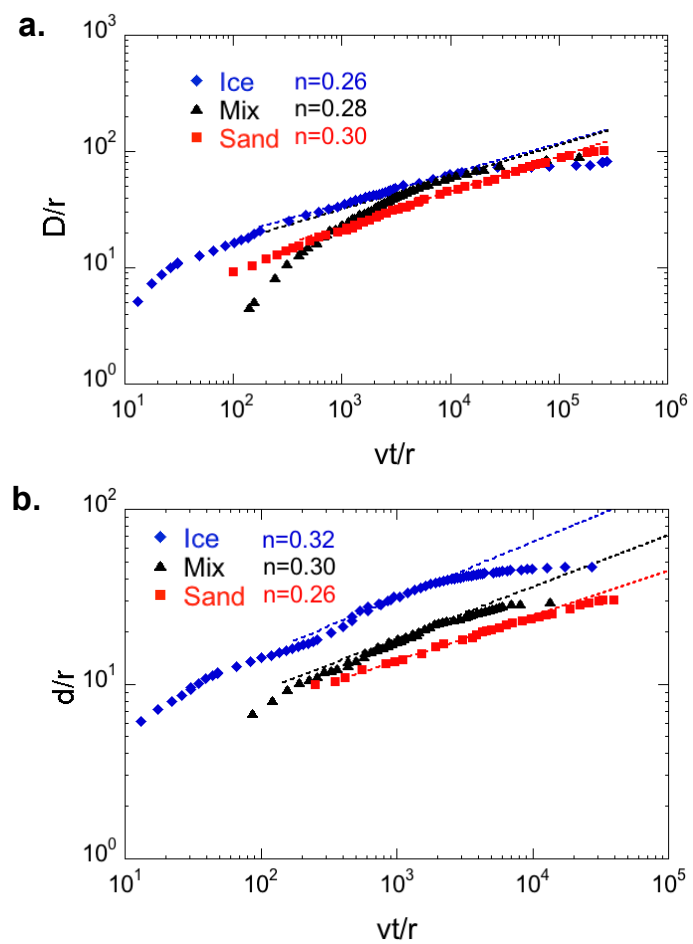


Figure 3.6

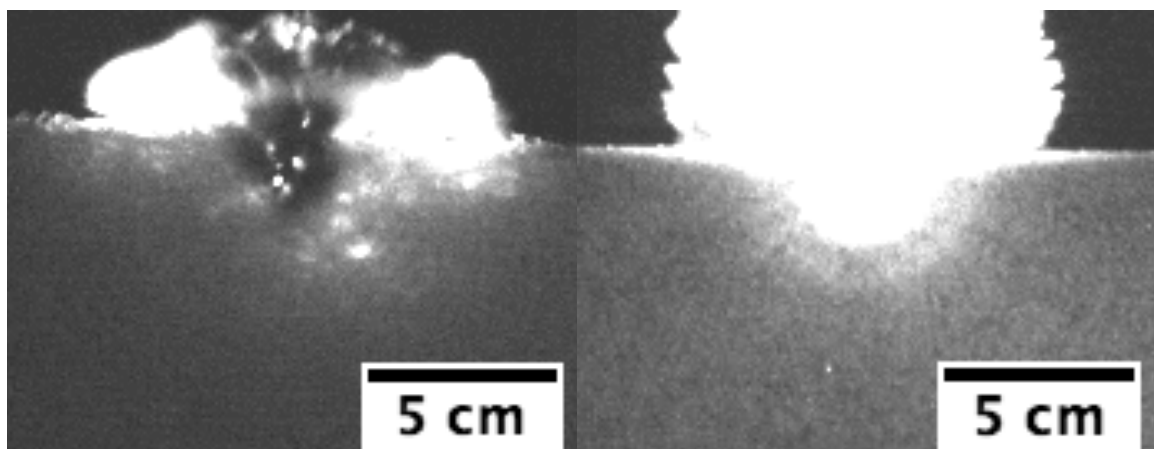


Figure 3.7

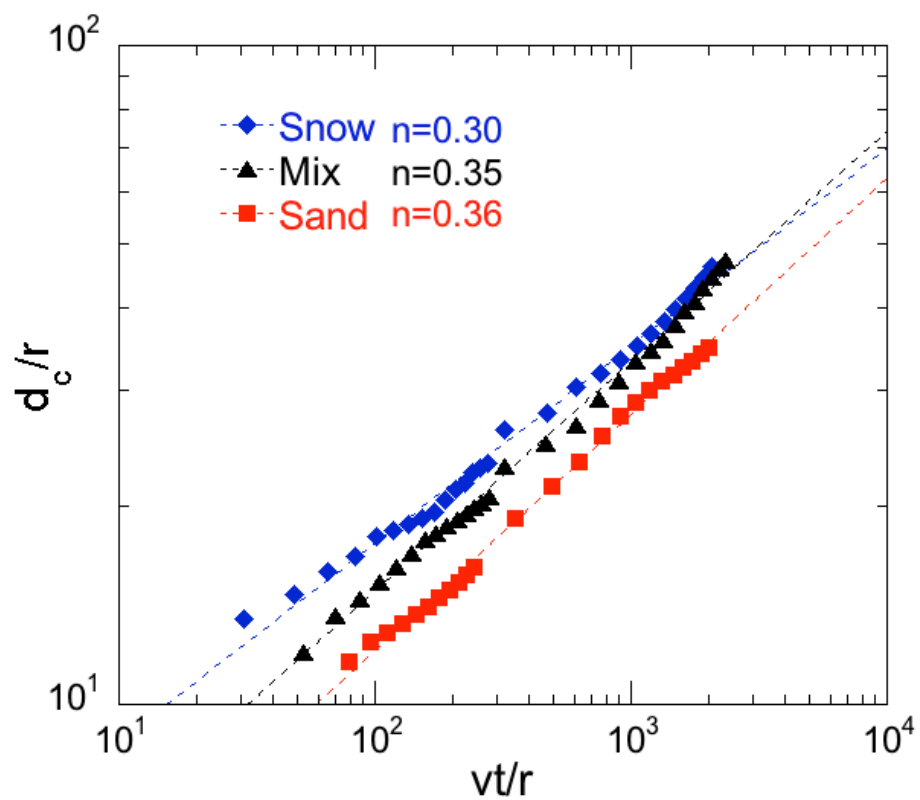


Figure 3.8

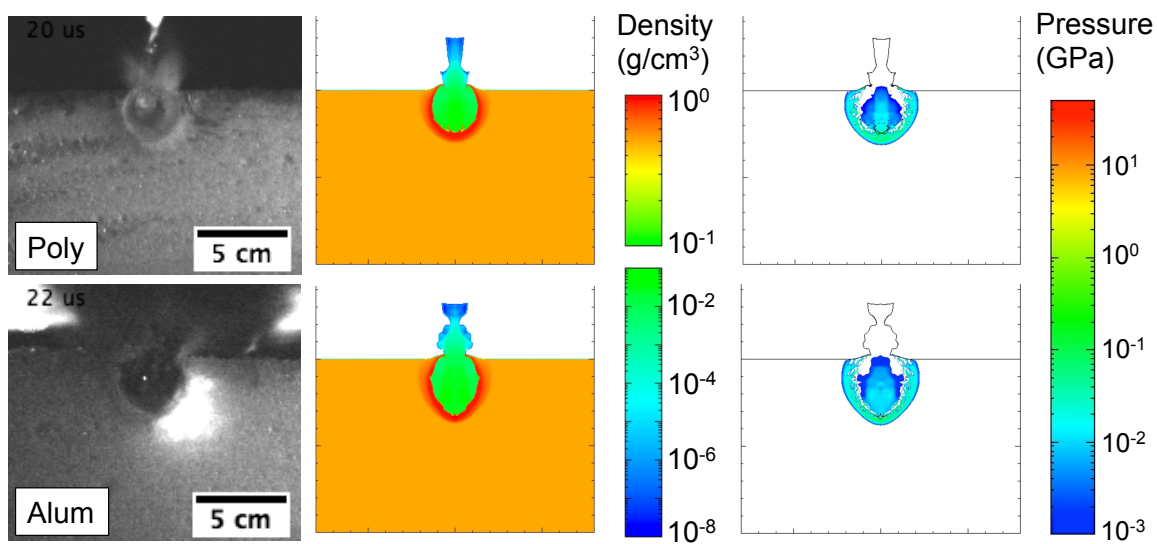


Figure 3.9

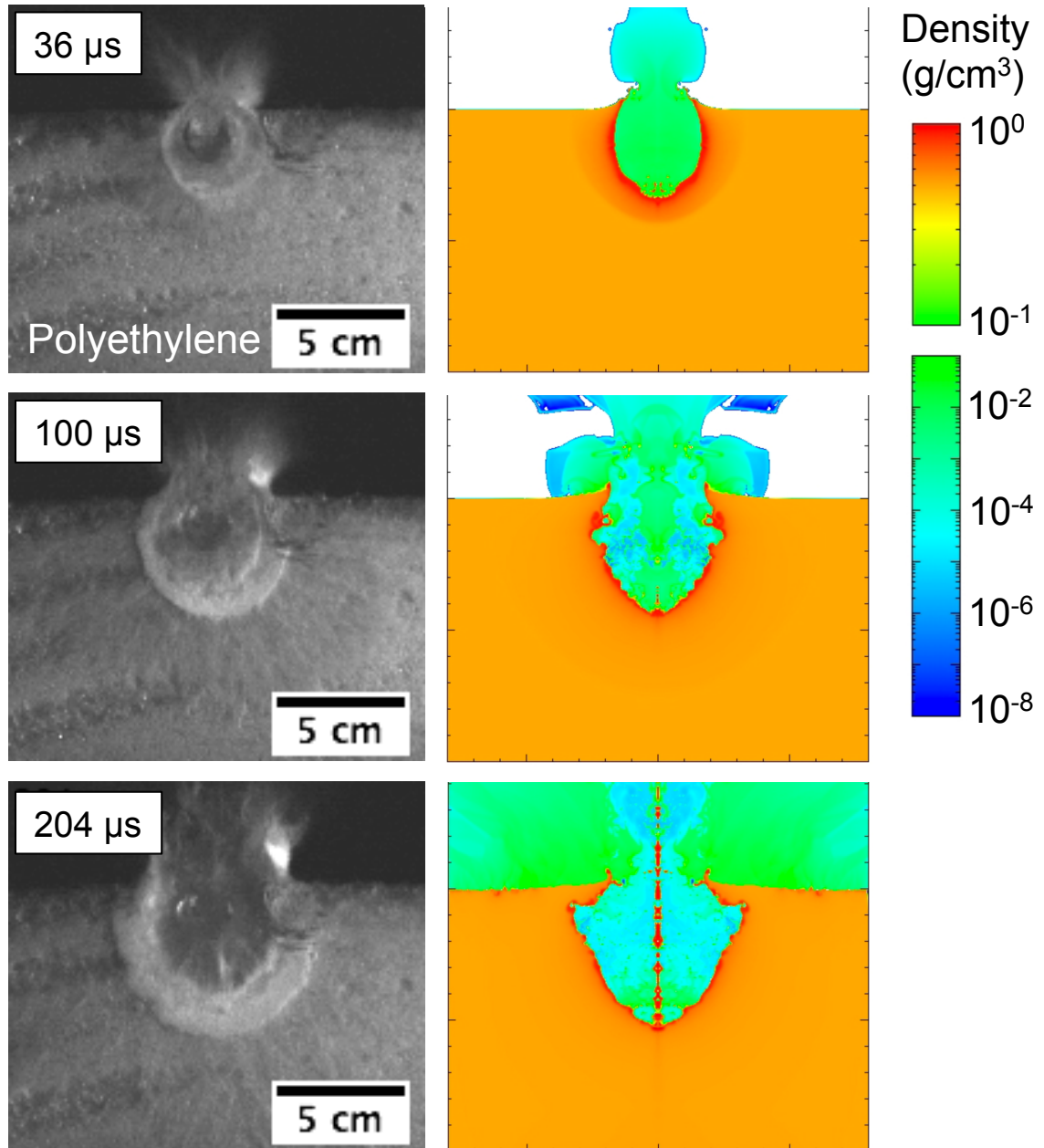


Figure 3.10

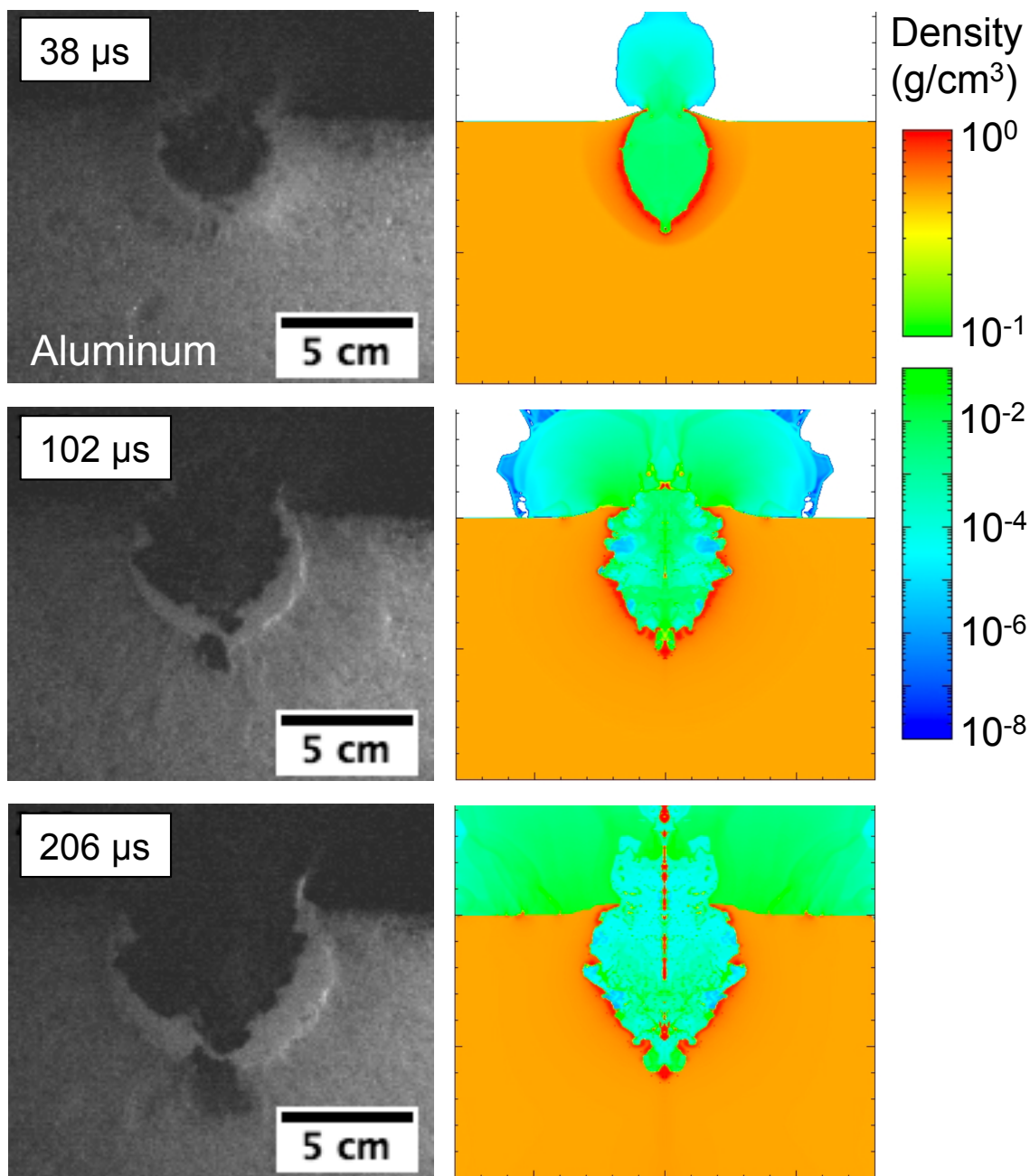


Figure 3.11

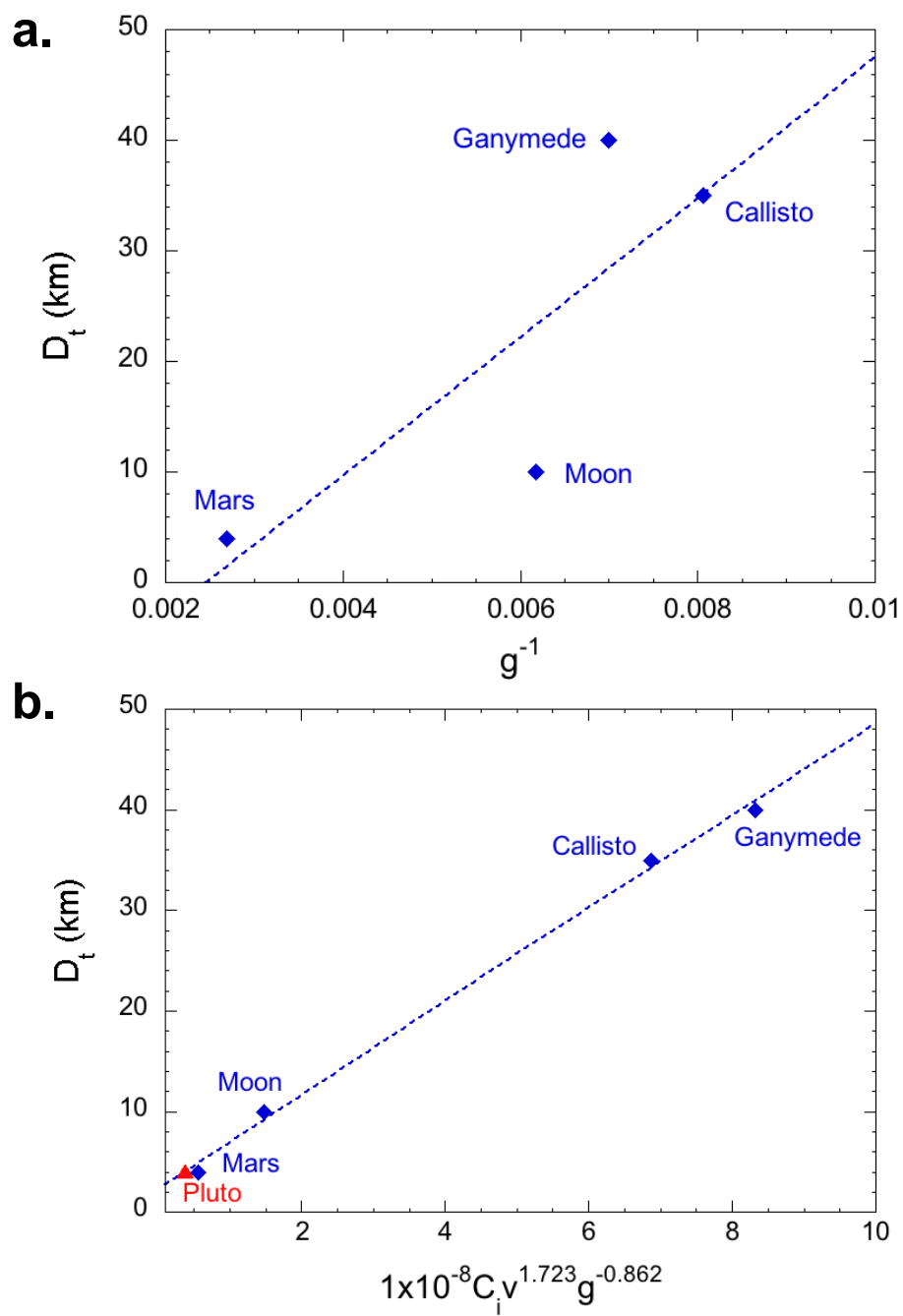


Figure 3.12

Chapter 4. Lunar Regolith Processing by Cometary Impact: Implications for Lunar Swirl Formation

Megan Bruck Syal and Peter H. Schultz

Department of Geological Sciences, Brown University

Providence, RI 02912

Abstract

Relatively recent cometary impacts at the Moon should leave unique traces of their origins: high impact velocities and volatile abundances, combined with the presence of a dust- and ice-laden coma, will thermally and mechanically process the lunar surface in ways distinct from the impact of an asteroid. Here we numerically assess the consequences of a cometary impact at the Moon by analyzing the combined effects of a collision by the nucleus and inner coma. Cometary impacts are found to entrain the finest fraction of lunar soil grains ($<10\ \mu\text{m}$) over regional scales ($\sim 1000\ \text{km}$), produce large masses of vaporized material, and generate transient magnetic fields that exceed the Earth's field strength by a factor of 10^4 . This combination of processes is consistent with a mechanism to generate lunar swirls: diffuse, meandering disturbances in brightness and regolith texture that curl across much of the lunar farside and are also associated with magnetic anomalies. Previous observations of swirl features indicate that bright regions are not only fresh but also possess a peculiar, altered regolith structure, which can be produced by the removal of fine soil grains. Regional scouring by an impacting comet explains both observations: large dynamic pressures expose fresh subsurface material while entraining the smallest grains within a near-surface flow of dusty plasma. Darker lanes observed within swirl regions are interpreted as melt and/or vapor deposits. Finally, the intense magnetic fields generated during high-speed cometary impacts provide an explanation for correlations between swirl locations and magnetic anomalies.

4.1 Introduction

Although the surface of the Moon almost certainly bears the scars of both asteroidal and cometary impacts, the relative contribution of comets to the observed cratering record remains uncertain. Quantifying the fraction of near-Earth impactors of cometary origin is complicated by multiple apparent source regions for these icy bodies. The scattered disk of the Kuiper Belt has been implicated as the supplier of Jupiter Family Comets, which represent a dynamical class of Short Period Comets (SPC) [Duncan and Levison, 1997], while Long Period Comets (LPC) trace to the distal, isotropic Oort Cloud [Oort, 1950; Dones et al., 2004]. Most observable LPCs are dynamically “new”; LPCs with perihelia < 3 A.U. are estimated to be injected at a rate of 12 comets per year [Wiegert and Tremaine, 1999]. The injection of comets into the inner solar system, however, is sensitive to time-varying perturbations by galactic tides and stellar impulse [Fernandez and Ip, 1987; Matese and Lissauer, 2002; Dones et al., 2004]. Consequently, prior enhancements in cometary flux, on the order of 10^7 year intervals, are dynamically plausible.

Even in the absence of such flux enhancements, the estimated combined flux of SPC and LPC impactors is significant: Weissman [2006] calculates an Earth impact rate of 1.4×10^{-7} comets/yr for cometary nuclei exceeding 2 km in diameter. Scaling this to the Moon, using the ratio of effective cross sections between the two bodies [Hartmann, 1977], the mean time interval between 2-km comet impacts at the Moon is $\tau \sim 1.1 \times 10^8$ years. Applying a reasonable power law exponent of $q=1.5$ to the population of cometary nuclei [Snodgrass et al., 2011], yields a mean time interval of $\tau \sim 3.8 \times 10^7$ years between 1-km comet impacts at the lunar surface. Because bolide impacts follow a Poisson distribution, such mean time intervals are not unreasonably long within the context of lunar surface ages. Hence, the Moon may record one or more relatively fresh cometary impacts. The surface expression of such an event is

expected to differ from the statistically dominant asteroid impacts, due to a combination of factors: greater impact velocities, differing acoustic impedance conditions, higher volatile content, and the presence of the encompassing gas and dust-enriched coma.

The reality of LPC impacts (or near-impacts) in the inner solar system has recently been reinforced by observations of newly-discovered comet C/2013 A1 (Siding Spring), which will make a close pass of Mars in late 2014. Even at a miss distance of 40 Mars radii, the high encounter velocity ($v \sim 56$ km/s) and presence of large particles within the coma are being carefully studied to constrain potential threats to Martian spacecraft and thermal effects on Mars' atmosphere [Ye and Hui, 2014]. Occurrence of such events on human time scales underscores the likely role of comets in resurfacing inner solar system bodies in the past.

Recent work to address cometary collisions at the Moon focused on comets as delivery vehicles for lunar volatiles [Ong et al., 2010; Stewart et al., 2011], a problem for which it is sufficient to ignore the tenuous coma surrounding the nucleus. However, including coma effects is of interest for assessing potentially broad, regional effects on lunar regolith processing; this problem has been previously considered [Schultz and Srnka, 1980], in regards to lunar swirl formation. Here we extend the prior work of Schultz and Srnka [1980], by using modern hydrodynamical models to consider the compression of the inner coma at various spatial scales, incorporating the impact of the icy and porous comet nucleus, and assessing the combined consequences of the impacting cometary complex for the lunar surface. This work is also informed by a wealth of data from recent flyby missions to comets, which add new physical insights into conditions within the inner coma and nucleus regions of comets, including estimates of: material strength, density, porosity, spatial variation of activity, and the lofting of large, icy particles [A'Hearn et al., 2005; Schultz et al., 2007; A'Hearn et al., 2011; Hermalyn et al., 2013; Kelley et al., 2013].

Finally, within the context of new observational constraints, we reexamine the cometary collision hypothesis for the origins of lunar swirls and compare this mechanism to other proposed models. Lunar swirl regions, such as the Reiner Gamma feature, shown in Fig. 4.1, are typified by bright meandering, diffuse patterns, interspersed with darker lanes. Swirl complexes do not correlate with any resolvable topographic variation and appear draped across the surface of the Moon. Additionally, swirls are associated with local paleomagnetic anomalies [Hood et al., 1979, 2001; Richmond and Hood, 2008]. The cometary collision hypothesis detailed in Schultz and Srnka [1980] emphasized the unique photometric properties of the swirls, as evidenced by significant swirl brightening under large phase angles ($> 150^\circ$), in understanding swirl origins. Such strong forward scattering required a process that altered the texture of the upper surface structure without affecting other properties, e.g., radar reflectivity, thermal properties, and topography. Consequently, the origin (and disappearance) of the swirls must be related to the physical processing of the regolith, rather than solar wind darkening. Moreover, in the cometary impact model, the swirl-forming process was directly linked to the generation of localized magnetic fields, rather than arising as a consequence of a pre-existing field. The new results reported here provide deeper insights into the mechanical, thermal, and magnetic effects of a cometary impact at the Moon and directly relate these results to observations of lunar swirl regions.

4.2 Impact of Cometary Coma: Analytical Approach

As a comet travels nearer to the sun, water ice and other minor volatile species within the comet nucleus begin to sublimate, generating a gravitationally unbound atmosphere or “coma” [Whipple, 1950]. Entrained within the outgassing species are a variety of solid phases, including dust of both silicate and organic composition and icy grains. Additionally, the EPOXI Mission flyby images of Comet 103P/Hartley

2 revealed individual, larger aggregate particles of ice and/or dust (radii up to 0.2 - 2.0 m) within the inner coma. The lofting of these particles is also very likely driven by sublimation at the nucleus [Hermalyn et al., 2013; Kelley et al., 2013; Bruck Syal et al., 2013]. For cometary impacts within the inner solar system, the sublimation-driven distribution of both vapor and particulate materials within the near-nucleus region comprises a secondary effect on the impact process, in addition to the impact of the solid nucleus. Hence, a complete treatment of a lunar cometary impact will consider the dual effects of coma and nucleus hypervelocity collisions.

4.2 Heat Transfer Effects

Cometary atmospheres, although very tenuous, may have substantial effects on planetary surfaces when colliding at the high encounter speeds typical of comets, particularly in the case of LPCs, a population described by a root mean square velocity of 52 km/s at the Moon [Hartmann et al., 1981]. A useful analogy, first described in Schultz and Srnka [1980], is the ablation of meteoroids within the high-altitude, low-density portions of Earth’s atmosphere: the impact of a low-density cometary atmosphere with grains at a planetary surface can be considered a simplified version of this meteoroid ablation problem, with the inertial frames reversed.

For grains of sufficiently small sizes ($< 100 \mu\text{m}$, appropriate for lunar regolith particles), the energy transfer rate from the atmosphere is:

$$\sigma T^4 = \frac{a\rho v^3}{8} \quad (4.1)$$

where T is the grain temperature, σ is the Stefan-Boltzmann constant, a is the accommodation coefficient, ρ is the atmospheric density, and v is the impact velocity [Ceplecha and Padevč, 1961]. Using a conservative $a = 0.8$ [Opik, 1958], comet impact velocity $v = 60 \text{ km/s}$ (approximate encounter velocity of Comet Siding Spring),

and a lunar soil melting temperature of $T_{melt} = 1473$ K [Taylor and Meek, 2005], the critical coma density to melt lunar soil grains can be derived:

$$\rho_{crit} = \frac{8\sigma T_{melt}^4}{av^3} = 1.24 \times 10^{-11} \text{ g/cm}^3 \quad (4.2)$$

Water vapor densities of this order are present in the innermost regions of cometary comae [Tenishev et al., 2008]. Hence, the collision of the low-density H_2O atmosphere is sufficient to induce surface modification, independent of dust and ice particle effects. The spatial extent to which a coma, compressed against the lunar surface, will achieve such densities, depends upon the water production rate of the impacting comet.

High-resolution flyby images of cometary nuclei [Keller et al., 1986; Soderblom et al., 2002; Sekanina et al., 2004; Farnham et al., 2007; A’Hearn et al., 2011] reveal dramatic directional variability in the outgassing patterns and jet activity within cometary comae. Increasingly sophisticated kinetic methods now utilize numerical models of both the gas and dust activity within cometary atmospheres, thereby connecting the detailed density distributions of each species back to the comet nucleus [Tenishev et al., 2008, 2011]. A widely-used analytic approximation for the radial decay of coma gas density is the isotropic *Haser* model [Haser, 1957]:

$$N_P = \frac{Q_P}{4\pi r^2 v_p} \exp(-r/R_p) \quad (4.3)$$

where, for a given parent species, N_P is the number density, Q_P is the production rate, r is the radial distance, v_p is the characteristic gas outflow velocity, and R_p is the parent species length scale. Although the *Haser* model is simplified, relative to detailed observations of comet activity, it provides a useful framework in which to explore coma effects on cometary impacts. As water is the primary volatile constituent ($\sim 90\%$) and will dominate any mechanical effects during a hypervelocity collision, we will concern ourselves only with the water production values. Some comets, how-

ever, may have additional, non-trivial contributions from CO₂ and CO to inner coma densities [A'Hearn et al., 2011; A'Hearn et al., 2012; Ootsubo et al., 2012] .

For H₂O, $R_p > 5 \times 10^4$ km [Combi et al., 2004]; hence, for the higher-density inner coma distances of interest ($r < 1000$ km), the exponential term in Eq. (4.3) is inconsequential and the density distribution simply decays as r^{-2} . For a nucleus radius R_{nuc} , a simple integral reveals that 90% of the coma's total mass is contained within $r < 10R_{nuc}$.

Two representative water production rates are useful to consider: $Q_P = 10^{27}$ mol s⁻¹, which, for a 1-km diameter comet nucleus, is equivalent to a Halley-like production rate per unit surface area [Keller et al., 1986]; and $Q_P = 10^{29}$ mol s⁻¹ which is a closer analogue to the recent interloper from the Oort cloud, Siding Spring [Ye and Hui, 2014]. Using $v_p \sim 0.7$ km/s for near-nucleus H₂O, the resulting radial density distributions are plotted in Fig. 4.2 a. Note that while the less-active coma density exceeds the critical ablation value (Eq. 4.2) for only the very innermost region, the high-activity coma remains above ρ_{crit} for a significant radial distance.

Upon hypervelocity compression at a planetary surface, continuity of flow implies that the near-surface densities of these atmospheres should be significantly enhanced: a first order approximation of the density enhancement can be obtained by dividing the impact velocity ($v_{imp} = 60$ km/s) by the thermal expansion velocity of water of ($v_p = 0.5$ km/s; applicable after impact) for a factor of 120 increase. This transient enhancement in atmospheric densities at the lunar surface, as the coma collides before, during, and after the comet nucleus, allows for ablative processes to operate out to greater distances. Applying this to both water production cases (Fig. 4.2 b.), the zone of critical density for soil ablation should extend out to $r \sim 6$ km for the less active case and to $r \sim 60$ km for the higher activity example.

4.2 Particle Mobilization Effects

In addition to the transfer of kinetic into thermal energy upon impact of the lunar surface, the hypervelocity collision of an atmosphere will induce mobilization of entrained particles. The maximum particle size that the compressed coma will entrain depends upon the dynamic pressure, $q = \frac{1}{2}\rho v^2$, within the flow. A critical grain size can be calculated by balancing the gravitational ($F_g = mg$) and drag ($F_D = \frac{1}{2}\rho v^2 C_D A$) forces exerted on a particle, where C_D is the drag coefficient (0.47 for a rough sphere), and A is the cross-sectional area. This allows derivation of an expression for maximum mobilized particle diameter, d_{max} :

$$d_{max} = \frac{3\rho_{gas}v^2C_D}{4g\rho_{grain}} \quad (4.4)$$

Using this expression, we can calculate, for both comet activity cases shown in Fig. 4.2, the maximum mobilized grain sizes (once entrained) as a function of radial distance from the impact site. The results, using the 1-D flow continuity approximation for surface density enhancement (factor of 120), are plotted in Fig. 4.3. Remarkably, the higher activity case mobilizes 10 μm sized grains out to a radial distance of 1000 km; the moderate activity case lofts 10 μm grains out to distances exceeding 100 km and 1 μm grains out to approximately 350 km.

Entrainment conditions (Eq. 4.4) are distinct from the initial mobilization of surface materials, due to boundary layer effects [Greeley and Iversen, 1987]. Injection of particles into the near-surface flow of compressed coma gas can be achieved through many small, simultaneous impacts by particles within the coma. Once mobilized and entrained, this two-phase flow can scour the surface out to large regional distances.

These estimates are lower limits on dust entrainment potential, as they do not account for non-isotropic enhancements in gas and dust densities (e.g., within cometary jets, sub-solar regions, and diffuse regions of greater activity at the nucleus) or the

additional scouring by comet dust and ice grains embedded within the coma. Additionally, as the natural impact angle probability distribution favors oblique angles [Gilbert, 1893; Shoemaker, 1962], the asymmetric, downrange-directed buildup of coma density can permit mobilization out to even larger length scales. From geometric effects, for an impact angle of 45° , the sector of lunar surface in contact with a coma's inner 1000 km would subtend an angle of 69° , covering distances greater than 2000 km. A three-dimensional assessment of coma density buildup and the effects of concurrent coma and nucleus impact motivates a numerical treatment of the problem, using a hydrodynamic code.

4.3 Impact of Coma and Nucleus: Numerical Approach

Three-dimensional simulations were carried out with Sandia National Laboratory's CTH Shock Physics code [McGlaun et al., 1990], in order to more fully characterize the physical effects of a coma and nucleus collision with the lunar surface. The CTH suite of codes utilizes a Lagrangian deformation step to solve the conservation equations (coupled with an applicable equation of state and constitutive model as necessary) with a subsequent remapping to an Eulerian grid. For computationally expensive simulations, Adaptive Mesh Refinement (AMR) [Crawford, 1999] allows optimal resolution of physically interesting regions within the problem domain, while conserving CPU usage; all three-dimensional simulations reported here utilized CTH's AMR capabilities. CTH is commonly used to solve planetary-scale problems of interest, including the collision of asteroids or comet nuclei with planetary bodies under various initial conditions. Here we use CTH to simulate the collision of both the solid (nucleus) and gas (coma) components of a comet.

The rarefied conditions within the unbound atmospheres of comets require kinetic approaches, such as the Direct Simulation Monte Carlo (DSMC) method detailed in Tenishev et al. [2008], in order to accurately model much of the coma. However, in

the near-nucleus region, the mean free path of water molecules is sufficiently low ($\lambda < 1$ m) to permit a hydrodynamic description of the gas flow. An approximate radius for the collisional region is defined by:

$$R_{coll} = \frac{Q_{gas}\sigma}{4\pi v} \quad (4.5)$$

where Q_{gas} is the water production rate, σ is the water collisional cross section ($\sigma \approx 10^{-19}$ m²), and v is the water outflow velocity ($v \sim 700$ m/s) [Combi and Smyth, 1988]. For the lower activity case discussed in Section 2 ($Q_P = 10^{27}$ mol s⁻¹), the region in which a hydrodynamic approach remains valid extends out to $R_{coll} \sim 11$ km; for the larger production rate ($Q_P = 10^{29}$ mol s⁻¹), $R_{coll} \sim 1100$ km. Hence, depending on the activity level of a comet, usage of a hydrocode, such as CTH, to investigate the effects of an impacting, shocked coma is valid out to significant regional scales.

Simulations were carried out over a range of length scales in order to focus on different aspects of the concurrent impacts of the coma and nucleus. At the largest scales (coma radius, $R_c = 1000$ km), the surface curvature of the Moon is important and included in the target geometry. At the smallest scales, the isolated effects of the impacting solid nucleus are considered, independent of the coma. Impact angle was maintained at 45°, which corresponds to the peak in the impact angle probability distribution function [Gilbert, 1893; Shoemaker, 1962].

Comet nuclei of 1 km in diameter were used in all simulations. This choice is based upon the reasonably short mean time interval between comet impacts of this size ($\tau \sim 38$ Myr; see Introduction) and recent observations of several dynamically new comets with nuclei in this approximate size range, including ISON (C/2012 S1) [Knight and Walsh, 2013] and C/Siding Spring [Ye and Hui, 2014]. Moreover, scaling constraints are placed upon the nucleus size by the crater diameter of Goddard A ($D = 12$ km), which is presented in Schultz and Srnka [1980] as a possible nucleus impact

site for the generation of the Mare Marginis swirls. While we are concerned with the problem of an impacting comet in general, we are also exploring the problem's direct applicability to formation of lunar swirl regions. Using the scaling relations detailed in [Holsapple and Housen, 2007], the ratio between crater and impactor radius, for a crater residing in the gravity regime, with $\mu = 0.55$ (i.e., wet sand) is approximated:

$$\frac{R}{a} = 0.93 \left(\frac{ga}{U^2} \right)^{-0.2157} \left(\frac{\delta}{\rho} \right)^{0.3137} \quad (4.6)$$

For $g = 1.622 \text{ m/s}^2$ (lunar gravity), $\delta = 0.5 \text{ g/cm}^3$ (comet nucleus density), $\rho = 2.55 \text{ g/cm}^3$ (lunar regolith density), $U = 60 \text{ km/s}$ (impact velocity), and an impactor radius of $a = 0.5 \text{ km}$, $R/a \sim 13$. Wet sand scaling constants are appropriate for describing cratering in lunar regolith at gravity-controlled crater sizes, due to the low strength and reduced porosity (relative to dry sand). While variations in final crater geometry due to uncertainties within the impact velocity, incidence angle, and material properties are likely, a nucleus of 1 km in diameter is the approximate size necessary to generate a Goddard A-size crater.

The semi-analytical ANEOS equation of state [Thompson and Lauson, 1972; Melosh, 2007] was used for all materials in the calculations. As comet impact velocities are relatively high, significant volumes of melted and vaporized projectile and target material will be generated in cometary collisions; hence, the choice of an equation of state with accurate phase boundaries is particularly important. Phase boundaries and phase change behavior during hypervelocity impact for the ANEOS equations of state for water ice [Pierazzo et al., 1997; Turtle and Pierazzo, 2001; Barr and Citron, 2011] and SiO_2 [Melosh, 2007] have been well documented in prior studies, so that these material models are amongst the best choices for modeling cometary and lunar regolith materials.

The comet nucleus was modeled using a modified version of the ANEOS water ice

equation of state, which accounts for the generation of molecular water vapor (as opposed to purely monatomic vapor). Density measurements of cometary nuclei demonstrate that these ice-rich bodies contain very significant void space, with bulk densities of $\rho \sim 0.5 \text{ g/cm}^3$ [Thomas et al., 2013a,b]. Accounting for this porosity in comet impacts has important consequences for both the initial acoustic impedance matching conditions at the target and for the additional irreversible heating of the cometary ices from the PdV work done during the crushing out of pore space [Zel’dovich and Raizer, 1966]. Recent experimental and numerical work using water ice of comet-like porosities demonstrates the enhanced heating and vaporization associated with compaction of porous water ice [Bruck Syal and Schultz, 2014]. Hence, the molecular water ice EOS included a P-alpha model [Herrmann, 1969; Kerley, 1992] with a small crush strength of 0.1 GPa [Stewart and Ahrens, 2004] and an initial reference density of $\rho \sim 0.5 \text{ g/cm}^3$. The initial temperature was set to 50 K, which is representative for all but the uppermost meters of cometary nuclei [Priyalnik et al., 2008].

The gaseous coma was implemented by applying a tabulated radial profile of decaying water vapor densities to the ANEOS water EOS (see Fig. 4.2). The resulting distribution is an isotropic, spherical coma with a $\rho \sim r^{-2}$ dependence. All results are for the higher-activity case $Q_P = 10^{29} \text{ mol s}^{-1}$; As the outflow velocity of water vapor from the nucleus ($v \sim 0.7 \text{ km/s}$) is small, relative to the encounter velocity of the comet ($v = 60 \text{ km/s}$), the entire coma and nucleus complex was assigned the same initial velocity of 60 km/s. The lunar surface was modeled with the quartz equation of state described in Melosh [2007], using the default EOS parameters; the reference density of $\rho = 2.65 \text{ g/cm}^3$ is consistent with lunar crust density measurements derived from gravity data [Wieczorek et al., 2013]. For the comparative case of an asteroidal impactor, the same quartz EOS was utilized to model the projectile. In this case, a P-alpha model with a moderate amount of porosity ($\phi = 0.25$) and a crush strength of 1 GPa [Lindholm et al., 1974] was used to decrease the reference density to $\rho =$

2.0 g/cm³, consistent with typical asteroid densities [Britt et al., 2002].

4.4 Impact of Coma and Nucleus: Results

4.4 1000-km Coma Impact

The largest spatial scale considered was the inner 1000 km of the coma, which is just less than the collisional radius for $Q_P = 10^{29}$ mol s⁻¹. The geometry of the problem is shown in Fig. 4.4; for these initial conditions, lunar soil will contact the coma complex out to 1400 km downrange and 730 km uprange from the impact point. In order to focus available numerical resolution on the compression of the coma, nucleus impact effects were not modeled here but are treated in subsequent calculations. Adaptive mesh refinement indicators were used to focus highest cell resolution in regions of coma compression and density enhancement near the lunar surface.

The development of a near-surface bow shock and density buildup was apparent at even early times (see Fig. 4.4); the final scale height of this shock layer was approximately 100 km. Compressed densities, extending radially from the impact point out to ~ 1000 km, were near $\rho = 10^{-13}$ g/cm³. Once particles are ejected from the surface, e.g., by impacting cometary dust grains, these coma densities are sufficient to entrain any particles smaller than ~ 30 μ m. The transient atmosphere, which engulfs a large fraction of the Moon's surface, is sustained on the order of 10² s.

The ionized fraction of water vapor molecules, calculated internally by ANEOS using the Saha ionization equation, was also tracked throughout the simulation. Upon impact of the surface at 60 km/s, the vapor was predicted to be effectively completely ionized (Fig. 4.5). While this may be a slight over-estimation, ionized fractions exceeding 0.1 are consistent with laboratory studies of the plasma generated by 40 km/s impactors [Dietzel et al., 1972]. Comets possess significant magnetic fields, e.g., flyby measurements of comet Halley revealed 70-80 nT field strength [Riedler

et al., 1986], and cometary dust particles are electrostatically charged [Mendis and Horányi, 2013]. Consequently, impact ionization and compression of the inner coma will affect the dusty plasma environment near the lunar surface and contribute to the amplification of the coma’s pre-impact magnetic field. Charge separation associated with ejecta-plasma interactions from the impacting nucleus, a process detailed in Crawford and Schultz [1999], will make additional contributions to transient magnetic field strength. The characteristics of the near-surface flow of dust and gas may then be controlled by magnetohydrodynamics.

4.4 100 km Coma Impact

An intermediate case considered was the impact of the inner 100 km of the coma, coupled with the impact of a nucleus 1 km in diameter. This allowed tracking both the very near-nucleus enhancements in coma density (perhaps sufficient, as detailed in Section 2, to melt lunar soil grains), and the expansion of the vapor plume generated by the nucleus impact. Again, impingement of the water vapor molecules against the surface created a bow shock. Densities within the bow-shock, prior to nucleus impact, were $\rho \sim 10^{-11}$ g/cm³ or greater, consistent with analytical predictions; the scale height for these densities was approximately 5-10 km (see Fig. 4.6).

While the large spatial scale of the problem precluded very high resolution of the impacting nucleus, general characteristics of the vapor plume expansion can be traced. The emergence of the vaporized comet nucleus from the transient cavity within a tenuous ambient atmosphere created a downrange enhancement of the plume at early times. This is likely related to the coupling of the plume’s energy with the transient atmosphere at sufficient length scales and the pre-existing, radially-directed velocity field at the lunar surface, which is established by the coma impingement. Figure 4.7 depicts vapor densities 0.65 s after nucleus impact (3.1 s after initial contact of inner 100 km of coma); note the differing density scale, compared with Fig. 4.6.

Downrange gas density enhancement occurs from the coupled impacts of coma and nucleus. This effect leads to vapor plume surface interactions out to greater distances, thereby increasing the likelihood of extended reprocessing of the Moon’s uppermost layer.

4.4 *Nucleus Impact: Comparison with Asteroid*

The impact of a comet nucleus, even in the absence of an enveloping coma, will affect the lunar surface in ways distinct from the impact of an asteroid. The most important distinction between cometary and asteroidal impacts are the differing impact velocities: asteroids are described by an approximate median velocity of $v \sim 16$ km/s [Yue et al., 2013], in contrast to the much higher velocities typical of comets ($v_{rms} \sim 52$ km/s) [Hartmann et al., 1981]. As target vaporization efficiency scales approximately as $M_t/M_p \sim v^2$ e.g., Bjorkman and Holsapple [1987], and vaporized target mass, M_t , scales only linearly with projectile mass M_p , the low densities of cometary nuclei, relative to asteroids, $\rho \sim 2.0$ g/cm³ [Britt et al., 2002], are more than compensated by the increased impact velocity. Additionally, the high volatile content of comets ensures that the entirety of the nucleus is likely to also be vaporized and contribute to impact vapor plume formation. For asteroids, however, portions of the impactor will more easily remain below critical vaporization pressures.

In order to contrast cometary versus asteroidal impact, a pair of impact simulations were performed: one using a 60 km/s comet (porous ice), the other using a 15 km/s asteroid (porous quartz, $\rho = 2.0$ g/cm³). Quartz was chosen for: (1) its accurate phase boundaries; and (2) its grain density, $\rho = 2.65$ g/cm³, which aligns with carbonaceous chondrite grain densities and allows a moderate bulk porosity assignment ($\phi = 0.25$) to achieve an initial asteroid density of $\rho_0 = 2.0$ g/cm³ [Britt et al., 2002]. Initial calculations comparing the asteroidal and cometary impactors focused on tracking the post-impact expansion of any vaporized material, which could con-

tribute to downrange surface modification. For this purpose, a larger computational domain was used and the problem extended to longer time scales. Subsequently, higher resolution simulations (25 cells-per-projectile-radius) were used to quantify the total vaporized mass (vaporized projectile and target masses were computed separately) for each case. Vaporized mass was also quantified for an additional, lower velocity (10 km/s) asteroidal impact. The known sensitivity of phase change calculations in shock physics codes to mesh size [Pierazzo et al., 1997; Barr and Citron, 2011] motivates the higher resolution for these simulations. As the specific entropies and associated final release temperatures necessary to achieve vaporization are reached at early times during the impact process, these higher resolution cases are run for a shorter duration of time.

Plots of the dynamic pressure, $q = \frac{1}{2}\rho v^2$, for the 15 km/s asteroid and 60 km/s comet nucleus in Fig. 4.8 illustrate potential downrange interactions of vaporized material with the lunar surface. Both the initial time to penetrate the target and the downrange-directed velocity of the vapor plume front scale with impact velocity. This motivates plotting the asteroid impact at times scaled up by the factor of four velocity difference between the two bodies; a more relevant comparison between the consequences of cometary and asteroidal impacts is obtained this way. Even at early times, the cometary impact exhibits greater dynamic pressures and a more expansive vapor plume. The near-surface dynamic pressures from the cometary vapor plume significantly exceed the asteroidal case. The prolonged zone of large dynamic pressures is driven by the higher velocity, which not only vaporizes a larger mass of projectile and target but also establishes a plume with greater internal energy, which then expands at higher velocities. The lower-energy and lower-mass vapor plume associated with the asteroid impact expands vertically and disperses over relatively small distances from the impact site. Hence, for asteroidal impacts at the Moon, little (if any) scouring by vaporized material is expected; moreover, the region of vapor

interaction with the surface is likely to be overprinted by ejecta deposits within a few crater radii.

The larger internal energy of the comet-generated vapor plume can be observed in early-time temperature plots for both cases, shown in Fig. 4.9. Ambient lunar surface temperature was set to 200 K (mean equatorial temperature); blue material corresponds to 300 K, while red material exceeds 10,000 K (likely associated with significant ionization and the development of conducting plasma). Even though the plume from the asteroid impact is hot (species heated to a few 1000 K), a larger fraction of the species vaporized in the comet impact have temperatures exceeding 10,000 K. Additionally, the larger speed with which this material moves downrange allows for an extended zone of the lunar surface to be engulfed by the heated gas, which may generate additional melt and vaporization.

In order to robustly compare the vaporization efficiency of different impactors, a mass filtering algorithm tracked the cumulative mass of vaporized target and projectile materials using final release temperature criteria [Quintana et al., 2013]. Although the critical pressure method has been widely used in the planetary science community for impact-generated phase changes [Pierazzo et al., 1997], comparisons between experimentally determined impact vapor masses and computational results indicate that the final release temperature metric may provide greater accuracy under certain conditions, particularly in the case of oblique impacts [Bruck Syal and Schultz, 2014].

For the comet impact case, the nucleus mass, $M_P = 2.618 \times 10^{14}$ g, was completely vaporized within 0.025 s after impact and generated significant target vaporization: $\sim 30M_P$, using $T_{crit} = 3157$ K for the lunar crust [Melosh, 2007]. The asteroid impacting at 15 km/s ($M_P = 1.0472 \times 10^{15}$ g) was $\sim 90\%$ vaporized and generated an additional $0.7M_P$ of target vapor. The 10 km/s asteroid, however, was $\sim 10\%$ vaporized and did not vaporize much of the lunar target material ($< 0.005M_P$). Totals for each case are shown in Table 4.1; these results are consistent with the

expected velocity scaling [Bjorkman and Holsapple, 1987].

The net effect of larger vapor volumes for cometary impacts is an enhanced, downrange-directed vapor plume, which interacts with the lunar surface over broader areas. The emergence of this vapor plume from the early-time crater occurs in a transient atmosphere generated by shock compression of the water vapor coma. While this atmosphere is still tenuous, for the $Q_P = 10^{29} \text{ mol s}^{-1}$ case, the near-impact density is $\rho \sim 2 \times 10^{-7} \text{ g/cm}^3$, which is within an order of magnitude of atmospheric densities at the Martian surface, for example. Previous numerical and experimental work demonstrates the non-trivial effects of such atmospheric densities on the evolution of impact-generated vapor plumes and ejecta [Wrobel et al., 2006; Schultz, 1992]. The coupling of the plume energy to the atmosphere (and subsequent deceleration) results in the downrange-directed vapor cloud remaining nearer to the surface, promoting the mobilization of particles out to greater distances [Schultz and Wrobel, 2012]. Moreover, recent experimental and numerical work using Martian atmospheric conditions demonstrates the scouring effects associated with impact vapor plumes [Quintana and Schultz, 2014], which is observed on the surface [Schultz and Quintana, 2013].

4.5 Discussion

4.5 Dust and Ice Particle Effects

Although the calculations detailed in Sections 3 and 4 illustrate the potential for coma-driven surface interactions, this approach also neglects the significant dust and ice components within the coma, which are initially entrained by sublimating gas but eventually evolve to ballistic trajectories. Constructing a full numerical model of a two-phase, particle-laden flow impacting a surface at hypervelocity speeds is outside the scope of this study (and outside the capabilities of most hydrocodes), but available experimental data and theory provide insight for the likely effects of these entrained solids.

An analogous problem is the impingement of a two-phase, particle-laden jet. The dynamics of impinging jets are especially well suited to this problem, since, for a comet, the areas of greatest dust particle density are often collimated into jets, due to spatially varying outgassing and the ballistic decoupling of particles within a few nucleus radii [Combi et al., 2012; Bruck Syal et al., 2013]. Far from the nucleus ($\sim 10^7$ km), solar radiation pressure shapes the distribution of dust [Finson and Probst, 1968]. This extends the model by Schultz and Srnka [1980], who proposed an analogy with the effect of the descent stage of the Apollo LEM, which scoured the surface and created similar scarring of the lunar regolith with forward-scattering properties.

As a jet impinges against a solid wall, the flow is rapidly redirected, establishing a radial flow along the surface. Experimental study of jet impingement on walls demonstrates the complexities within such flows, particularly for obliquely impinged jets: in the case of a 45° impingement, downrange momentum flux at the wall is enhanced by a factor of 6 (relative to 90° case), due to strong azimuthal variations [Donaldson and Snedeker, 1971]. The presence of a recirculation bubble in the near-impingement region, a flow characteristic now known to be generated by interactions between differentially-directed shocks within the jet, is evidenced by complex streamlines at the plate surface [Donaldson and Snedeker, 1971; Kalghatgi and Hunt, 1976; Klinkov, 2005]. These streamlines were originally visualized using grease streak photography [Donaldson and Snedeker, 1971], as illustrated in Fig. 4.10, which reveals the transition between inward- and outward-directed flows.

Gas impingement drives the formation of a high-density bow shock, similar to those modeled at lunar scales in Section 4. Solid particles within the jet are decelerated upon encountering the bow shock layer; particles of sufficiently high inertia will impact the surface, while others will be entrained in the radially propagating wall jet. Impacting particles may also reflect off of the surface and subsequently be carried away within the wall jet [Klinkov, 2005]. The process of continual reflection in

the near-impingement region produces a cloud of particles; those arriving after cloud development may be scattered away. Cloud formation is sustained by the presence of the recirculation (stagnation) bubble. Velocities measured by Klinkov [2005] for both radially-flowing and reflection-cloud particles are shown in Fig. 4.11.

These experimental constraints on dust-laden jet impingement provide additional mechanisms by which a dust and ice coma may scour the lunar surface. The geometry of an obliquely impacting cometary coma will generate many interacting shocks, which establishes conditions for recirculation bubble formation. Since this flow feature was shown experimentally to imprint streamline patterns on easily-mobilized surfaces, its presence during a cometary impact also may generate flow-like patterns. The sinuous character of lunar swirls is consistent with this hydrodynamic scouring process.

Ice particles, both small and large, will vaporize upon impacting the bow shock and/or lunar surface, contributing to enhanced densities within the transient atmosphere engulfing the Moon. Comet-derived silicate and carbonaceous dust particles, however, may exhibit much of the same behavior as seen in the laboratory-scale jet impingement studies, including dust cloud formation within the bow shock layer and radial mobilization out to large distances. Dust that impacts the lunar surface will generate significant ejecta, further loading the radial flow with additional dust. In concert, these processes promote scouring and redistribution of particles, in fluid-like patterns, over prolonged distances.

4.5 Lunar Swirls: Background

The striking appearance of lunar swirl complexes, first noted in Lunar Orbiter images [El-Baz, 1972; Schultz, 1976], has generated continual interest in their formation mechanism(s) over the past several decades [Schultz and Srnka, 1980; Hood and Schubert, 1980; Starukhina and Shkuratov, 2004; Garrick-Bethell et al., 2011]. Swirl patterns are characterized by looping high- and low-albedo features; the dark

regions, often termed “lanes,” are optically darker than surrounding lunar terrain, whether the swirls occur in the maria or the highlands [Schultz and Srnka, 1980]. The bright and dark patterns typically do not correlate with topographic relief and appear to have been surficially emplaced. Differing illumination conditions reveal that the bright areas are forward scattering, making the swirls particularly apparent, relative to surrounding terrain, at large phase angles [Schultz and Srnka, 1980]. More recent photometric studies support the observation that lunar swirl regions are described by unusual phase functions [Pinet et al., 2000; Kreslavsky and Shkuratov, 2003; Kaydash et al., 2009], which depart from expected values at large phase angles. This effect implicates physical alteration of the regolith structure has occurred; since optical maturity only controls sub-micron regolith structure, the swirls’ high reflectance cannot be attributed to differential space weathering effects [Schultz and Srnka, 1980; Kaydash et al., 2009]. Additionally, these unusual photometric properties, which are distinct from the porous and open “fairy castle structure” of lunar soil in its equilibrium state [Hapke and Hoen, 1963], require that the swirls are geologically young features. Cross-cutting relationships with nearby craters [Schultz and Srnka, 1980] and spectroscopic evidence [Bell and Hawke, 1987] further suggest that the swirls are < 100 Myr old. Because swirls superpose young craters, e.g., King [Schultz and Srnka, 1980], some examples could be much more recent.

The spatial extent of swirls varies by location; complexes with apparent origins in the Mare Marginis and Mare Ingenii regions extend for hundreds of kilometers across the lunar surface. Near-side swirl Reiner Gamma (5° S, 60° W), the most well-studied example, is relatively isolated but may represent a distal, eastward extension of the Mare Ingenii system due to a split cometary body impacting the Moon over a few days, resulting in the crater Goddard A (Mare Marginis swirls) and O’Day (Mare Ingenii swirls) [Schultz and Srnka, 1980]. In addition to their anomalous visual appearance, swirls have attracted attention due to their association with lunar pale-

omagnetic anomalies [Hood et al., 1979, 2001; Richmond and Hood, 2008]. Global swirl locations and the local magnetic field anomalies associated with each swirl region were recently documented by Blewett et al. [2011]. Notably, a one-to-one correlation between magnetic anomalies and swirls is not present. While each swirl area appears to have at least a weak anomaly associated with it, many areas of enhanced magnetic field strength lack discernible swirl features.

Additionally, for swirls such as Mare Marginis, the area of enhanced magnetism is localized to just one section of the entire swirl structure and does not extend to the great distances traversed by these wispy features (e.g., compare mapped magnetic anomalies of Richmond and Hood [2008] to mapped swirls of Schultz and Srnka [1980]). Further puzzling is the observation that stronger anomalies, such as the field at Gerasimovich (28 nT at 30 km altitude), which is the largest magnitude anomaly at the Moon, may possess more subdued and less extensive swirl patterns than relatively weak magnetic anomalies, i.e. Mare Marginis (6 nT at 30 km altitude) [Blewett et al., 2011]. This complex relationship between swirl occurrence and magnetic field strength suggests that swirl formation is not a purely passive process, driven by the prior existence of local lunar magnetism.

4.5 Cometary Origin of Swirls

Any proposed mechanism for swirl generation must be able to explain several observations: (1) occurrence of highly forward-scattering, bright, looping patterns; (2) alteration of regolith texture, without affecting the night-time thermal/radar signatures; (3) large spatial extent (swirl complexes stretch up to 1000 km in length); (4) relative youth, <100 Myr; (5) narrow dark lanes that cross-cut bright regions; (6) association with local coherent paleomagnetism [Schultz and Srnka, 1980]. The physical consequences of a relatively recent cometary impact at the Moon, including large-scale (100-1000 km) scouring by inner coma gas and dust, additional regional

(>100 km) scouring by an efficiently vaporized comet nucleus (and any associated swarms of ~ 0.1 -1.0 m ice/dust aggregates) and lunar target material, transport of ejected/entrained dust in a near-surface, radial flow of ionized cometary plasma, and significant transient magnetic fields, is consistent with each of these observations.

The calculations in Section 2 and 3 demonstrated that, while coma densities may be sufficient to melt lunar soil grains at relatively near-impact locations (<5-50 km), mobilization of lunar particle sizes up to $\sim 30 \mu\text{m}$ in diameter may extend to 1000 km from the center of the comet complex. Hence, the effects of coma impact, compression, and radial flow over the lunar surface are dominantly scouring effects. Any near-impact melting by the coma will be dwarfed by the large volumes of melt and vapor generated in the nucleus impact, which is predominantly directed downrange. Nucleus-generated vapor, comprising ~ 30 times the initial projectile mass, produces large near-surface dynamic pressures, capable of creating extended zones of scouring. This plume-driven processing is unique to high-velocity cometary impactors; asteroidal impacts occur at velocities too low to generate vapor-related modification of the surface out to similar length scales. Additional scouring efficiency is provided by impacting dust and ice particles embedded within the inner coma. The tortuous appearance of the swirls is consistent with scouring achieved by this near-surface, hydrodynamic flow (perhaps a magnetohydrodynamic flow, due to large fractions of ionized species).

Both the high albedo and unique regolith texture observed within bright areas of lunar swirls can be explained by a scouring mechanism, which unearths fresh, high-reflectance material while producing variations in particle size, through the preferential lofting of smaller dust grains. Compositional analyses of lunar swirls, carried out over the past three decades using both telescopic and spacecraft-acquired spectroscopic data, repeatedly demonstrate that the bright regions are consistent with fresh material [Bell and Hawke, 1987; Pinet et al., 2000; Kramer et al., 2011; Blewett

et al., 2011], consistent with a relatively recent scouring event. However, immaturity arguments alone are not sufficient to account for the anomalous regolith textures indicated by the swirl’s peculiar phase functions [Schultz and Srnka, 1980; Pinet et al., 2000; Kreslavsky and Shkuratov, 2003; Kaydash et al., 2009]. The regolith structure in swirls, revealed by photometric and spectroscopic study, is consistent with removal of the finest fraction of grains, $< 45 \mu\text{m}$ [Pinet et al., 2000]. These observations require mechanical alteration of the lunar soil’s equilibrium state, a constraint accommodated by the dynamic pressures calculated in this study, which mobilize particles of this size to large distances.

The dark lanes associated with swirls, known to be darker than nearby, undisturbed surfaces [Schultz and Srnka, 1980], could represent melt products. The large volumes of melt and vapor produced during comet nucleus collision (Table 4.1) will be expressed at the lunar surface as darkened, glassy agglutinates. The generation and ejection of melt, in addition to the condensation of vaporized material, will occur concurrently with the scouring process, producing entangled patterns of looped dark and bright features. In addition, comets are extremely enriched in dark, carbonaceous material, relative to asteroids; comet dust contains an order of magnitude greater carbon content than carbonaceous chondrites [Jessberger et al., 1988]. This significant refractory component may further lower reflectance values within melted and vaporized material, as carbon is known to disproportionately reduce albedo in mixtures [Clark, 1983].

Very significant transient magnetic fields are predicted to be produced during the impact of a comet by at least two effects: (1) compression of comet’s intrinsic magnetic field [Gold and Soter, 1976; Schultz and Srnka, 1980]; (2) charge separation between nucleus-generated plasma and ejected material [Crawford and Schultz, 1999]. The compression mechanism should amplify a comet’s magnetic field by four orders of magnitude, achieving field strengths of $B \sim 10^{-3}$ Tesla for a Halley-like magnetic

field. Using experimental results from impact-generated magnetic fields, the impact of a 1 km asteroid ($v = 20$ km/s) has been calculated to produce fields of $B \sim 0.03$ Tesla field at a distance of 100 km from the impact site [Crawford and Schultz, 1999]. A 1-km comet nucleus will generate a much larger field, as the field strength scales linearly with with impactor mass but very strongly with velocity [Crawford and Schultz, 1999]:

$$B = 9 \times 10^{-20} \frac{mv^{3.6 \pm 0.1}}{x^2} \quad (4.7)$$

where B is magnetic field strength, m is projectile mass, v is impact velocity, and x is radial distance from impact (SI units). Scaling by the density of a comet nucleus, $\rho = 0.5$ g/cm³, and the factor of three increase in impact velocity, the field generated by a comet should be near $B \sim 0.4$ T, approximately four orders of magnitude larger than Earth’s magnetic field. Hence, this second mechanism may dominate the transient fields produced during cometary impacts. Importantly, the impact-generated field would also be coherent. Near-swirl paleomagnetism can then be acquired via shock or thermal remanent mechanism.

As swirls are formed through surface scouring and melt draping processes in the comet model, heated material will cool through the Curie point, locking in thermal remanent magnetism. In this way, the comet-driven formation of swirls is expected to be associated with local paleomagnetism, although the field strength may vary with the local abundance of magnetic remanence carriers. A very surficial origin of the swirl magnetic anomalies is consistent with magnetic field observations, which constrain the source to be a a depth of < 1 km [Nicholas et al., 2007].

4.5 *Alternate Models*

Three primary formation hypotheses for swirls are discussed in the current literature: (1) impact of one or more cometary bodies [Schultz and Srnka, 1980; Bell and Hawke, 1987; Starukhina and Shkuratov, 2004]; (2) solar wind standoff model [Hood and

Schubert, 1980; Hood and Williams, 1989]; (3) electrostatic lofting and redistribution of dust particles [Garrick-Bethell et al., 2011]. The coma-nucleus hypothesis tested in the present paper differs from the cometary impact model outlined in Starukhina and Shkuratov [2004], which attributes swirl formation to impacting swarms of cometary debris, without invoking effects from an extended coma.

The solar wind standoff model attributes swirls’ brightness to a pre-existing, sub-surface magnetic field, which shields the local lunar surface from the darkening effects of solar wind particles. The solar wind, however, is only one component of the lunar “space weathering” environment; vaporization associated with continuous micrometeoroid bombardment at the Moon also contributes to the darkening and reddening effects of space weathering [Hapke et al., 1975; Keller and McKay, 1993; Keller and McKay, 1997]. These spectral changes are now known to be principally controlled by the accumulation of sub-micron metallic iron in the rims of vapor-coated soil grains [Pieters et al., 1993; Pieters et al., 2000; Hapke, 2001; Noble et al., 2001]. Such a process will continue unabated by the presence of lunar magnetic anomalies; this is particularly troublesome when noting that ancient basin ejecta, dating to 3.6 - 3.9 Gyr, are the suggested sources of magnetic anomalies within the standoff model [Hood and Artemieva, 2008]. This formation sequence requires swirls to retain their spectrally fresh appearance over billions of years of micrometeorite processing. Additionally, their unique photometry [Schultz and Srnka, 1980; Pinet et al., 2000; Kreslavsky and Shkuratov, 2003; Kaydash et al., 2009] does not support an optical maturity process, as argued by Schultz and Srnka [1980].

Swirl emplacement through the electrostatic lofting of dust [Garrick-Bethell et al., 2011], a recently suggested mechanism, also relies on the pre-existence of paleomagnetic anomalies to shape swirl formation. In this model, lunar dust particles accumulate positive or negative charge through diurnally varying processes, including the photoelectron effect and low-density plasma interactions [Farrell et al., 2007]. Dust is

then assumed to be lofted twice daily, at the evening and morning terminators. Prior evidence for dust lofting at sunrise and sunset (particularly at sunrise) includes data from the Lunar Ejecta and Meteorites experiment, deployed during the Apollo 17 Mission [Berg et al., 1976]. Subsequently, depending on its charge, the dust is either repelled from or attracted to regions subjected to electrostatic charge separation, e.g., magnetic anomalies associated with lunar swirls [Garrick-Bethell et al., 2011]. As the lofting process only affects the finest fraction of dust ($< 10 \mu\text{m}$), which tends to be enriched in feldspar [Taylor et al., 2001; Pieters and Taylor, 2003], the optical properties of the swirls are attributed to the emplacement of fine, feldspathic material, which is preferentially lofted and deposited in swirl-like patterns. The dark lanes are then suggested to represent areas of zero electric field, where normal weathering processes occur. Subsequent work to model the magnetic anomalies at the Reiner Gamma and Airy swirls suggests that dark lanes correspond to areas of enhanced vertical field [Hemingway and Garrick-Bethell, 2012].

The charged dust lofting model addresses some of the weaknesses of the solar wind standoff hypothesis, by maintaining the spectral freshness of swirls through replenishment on shorter time scales and possibly accounting for a unique regolith microstructure (separate from optical maturity effects). One key question, however, is how to explain the spatial scales of many swirl patterns, which often extend far beyond any associated magnetic anomalies (e.g., Mare Marginis, Mare Ingenii). If lofted dust is settling only in regions of substantial electrostatic potential, the draping of swirl features over tens of thousands of square kilometers, mostly free of substantial magnetic anomalies, is difficult to explain. Additionally, as sunrise dust lofting is a global process, one would expect a cleaner correlation between magnetic anomalies and swirl occurrence: wherever charge separation occurs, locally lofted dust should accumulate. However, as detailed by Blewett et al. [2011], not all magnetic anomalies are associated with the presence of swirls.

Another issue is the role of the small size fraction in the regolith. The swirls' optical and spectroscopic properties have been shown to be consistent with the *removal* of the finest fraction ($< 45\mu\text{m}$) of particles [Pinet et al., 2000]. This implicates a scouring mechanism, rather than an emplacement process. If repulsive forces are invoked to explain the swirl observations (absence of small particles due to local charge separation), the problem of spatial extent still remains. Additionally, the presence of dark lanes, which are lower in albedo than surrounding terrain, is not fully explained by the electrostatic model.

Finally, the model described by Starukhina and Shkuratov [2004] invokes a swarm of small cometary impactors to generate swirl features and has some elements in common with the cometary model explored in this paper. Both approaches recognize the resurfacing potential of impact-generated vapor plumes, which will possess significantly larger dynamical pressures than the tenuous, impacting coma. Also, the detection of many $\sim 0.1 - 1$ m clumps of icy/dust aggregates in the inner coma of Hartley 2 [Hermalyn et al., 2013; Kelley et al., 2013] suggests that comet swarm impacts may commonly accompany the impact of a full-sized nucleus, such as the one modeled in the present study. Indeed, Schultz and Srnka [1980] noted the presence of small fresh dimple craters associated with some swirls.

However, hypervelocity collision of the inner coma's dusty plasma environment is not considered in the Starukhina and Shkuratov [2004] model. This simplification limits the regional extent of vapor plume scouring; the absence of an atmosphere at the Moon allows vapor, even from oblique impacts, to quickly expand away from the surface, unlike the situation at Mars [Schultz and Wrobel, 2012]. The contiguous, flow-like character of the swirls is also inconsistent with the impact of many smaller, isolated icy bodies. Perhaps most importantly, consideration of a larger nucleus impact, which occurs within an environment of impacting inner coma plasma and charged dust particles, provides a mechanism to generate magnetic anomalies associ-

ated with swirls; the comet swarm model does not address the association between swirls and magnetic anomalies.

The model presented here, however, differs slightly from the original model by Schultz and Srnka [1980]. While that model also included impacts by the nucleus, it attributed the swirls to interactions with denser jet regions within the inner coma. Here we argue that this process may not be sufficient to heat the necessary mass of material above the Curie point, and that the transient magnetic fields generated by a cometary impact would dominate any field amplification effects upon the compression of the coma with the surface. Rather, it is the interaction between the relatively hot and high-density vapor generated by the nucleus collision and the distal dust scouring effects of a dust-laden coma that driving the necessary conditions for regional scouring and heating. Furthermore, we add in the role of hydrodynamic instabilities and electrostatic interactions, which are promoted within the dusty plasma environment during a coma/nucleus impact, as possible contributors to the final resurfacing patterns.

4.6 Conclusions

Cometary impacts at the Moon, while less frequent than asteroidal impacts, should occur on short enough time scales (< 100 Myr) to be observable in the lunar cratering record. The Reiner Gamma, Goddard A, and Ingenii swirls, however, may represent the last recent events. Characteristics unique to these icy bodies, including higher velocities, greater volatile contents, and the development of dusty plasma atmospheres at inner solar system orbital locations, result in a distinct set of physical consequences upon impact with the lunar surface. These include: local ($\sim 5 - 50$ km) heating of lunar surface by inner coma plasma; regional ($\sim 100 - 1000$ km) scouring of ~ 10 μm particles; formation of a near-surface, radial flow of near-fully ionized coma gas; massive vapor plumes ($> 30M_P$) with high internal energies and large, near-surface

dynamic pressures; and transient magnetic fields of $B \sim 0.4$ T, approximately 10^4 times Earth's field strength.

These effects should produce traceable signatures of cometary impacts, including: elongated areas of disturbed regolith, in which the heavily-weathered topmost layer has been removed, revealing fresh, bright material; unique regolith textures within scoured areas; fluid-like patterns of scouring, associated with hydrodynamic instabilities in rapidly-flowing plasma; areas of darker material, signifying melt and condensed vapor deposition (may also be carbon-enriched); and local paleomagnetism, driven by charge separation between hot plasma and impact materials.

Each of these signatures is consistent with available photometric, spectroscopic, and magnetic observations of lunar swirl regions, which have puzzled planetary scientists since the first Lunar Orbiter images of swirls were acquired. While a cometary origin for swirl regions is not a new idea [Schultz and Srnka, 1980; Bell and Hawke, 1987; Starukhina and Shkuratov, 2004], this contribution is the first to test such a model with modern hydrodynamical methods, along with new constraints from comet flyby missions and lunar remote sensing studies. While we favor the comet collision model for swirl formation over other competing models, we anticipate the role of future missions in testing this hypothesis; tighter constraints on the depth of magnetization, along with *in situ* structural and compositional information are necessary to fully unravel the secrets of the swirls. Results from the present study, however, will enable new discussion about key processes in future contributions, including new constraints on age, emplacement processes, photometry, and other possible clues.

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Tables

Table 4.1: Vaporized Projectile and Target Masses

Impactor	v (km/s)	M_{proj}^a	M_{targ}	M_{total}	M_{total} (g)
Comet ^b	60	1.0	30	31	8.12×10^{15}
Asteroid ^c	15	0.9	0.7	1.6	1.68×10^{15}
Asteroid	10	0.1	0.0	0.1	1.05×10^{14}

^aReported in units of projectile mass, M_P , unless otherwise noted

^b $M_P = 2.618 \times 10^{14}$ g

^c $M_P = 1.047 \times 10^{15}$ g

Figure Captions

Figure 4.1: The Reiner Gamma formation, located at 7.5°N 59.0°W , is a well known lunar swirl region, due to its nearside location; the majority of swirl features are distributed across the lunar farside. Swirls are described by curving, diffuse bright patterns, which are interspersed with darker lanes. These anomalous albedo patterns are associated with local paleomagnetic anomalies of varying intensities; swirl complexes are often extensive and may stretch across thousands of kilometers, well beyond localized magnetic anomalies. Lunar Orbiter 2 image 2215_med. Image credit: NASA/LPI.

Figure 4.2: (a) Coma H_2O vapor density profiles, as calculated from Eq. 4.3, for two cases: $Q_P = 10^{27} \text{ mol s}^{-1}$ (moderate activity) and $Q_P = 10^{29} \text{ mol s}^{-1}$ (high activity) (b) Compressed densities of coma profiles shown in (a), using 1-D continuity of flow approximation for a $v = 60 \text{ km/s}$ impact. Water vapor densities are sufficient to melt lunar regolith grains out to 6 km and 60 km for the moderate and highly active cases, respectively.

Figure 4.3: Maximum diameter of mobilized lunar regolith grains, as calculated from Eq. 4.4, for two cases: $Q_P = 10^{27} \text{ mol s}^{-1}$ (moderate activity) and $Q_P = 10^{29} \text{ mol s}^{-1}$ (high activity). Particles up to $10 \mu\text{m}$ in diameter are mobilized out to distances of 1000 km by the more highly active comet, while the less active comet lofts $10 \mu\text{m}$ particles out to distances exceeding 100 km and $1 \mu\text{m}$ grains out to 350 km.

Figure 4.4: Density evolution for inner 1000 km of water vapor coma ($Q_P = 10^{29} \text{ mol s}^{-1}$), impacting the Moon at $v = 60 \text{ km/s}$, at 45° incidence angle. A bow shock, in which compressed densities of $\rho \sim 10^{-13} \text{ g/cm}^3$ extend out to 1000 km from the

central impact point, engulfs the lunar surface. These densities are sufficient to loft grains of 30 μm or less. Here the nucleus of the comet was not modeled, in order to focus on the compression of the coma.

Figure 4.5: Ionized fraction of water vapor as inner 1000 km of coma impacts the Moon, 17.5 s after initial contact. High temperatures associated with collision at 60 km/s allow efficient ionization as H_2O molecules encounter the lunar surface; the ANEOS equation of state predicts complete ionization under these conditions. Although this is an over-estimate, ionized fractions exceeding 0.1 are consistent with plasmas generated by hypervelocity experiments at $v = 40$ km/s [Dietzel et al., 1972]. Hence, the dusty plasma environment in the inner comae of comets, which contain significant ionized species prior to impact, will likely experience a large amplification in electron density during impact. This creates highly conductive conditions in the plasma, allowing large amplifications of the coma’s magnetic field strength.

Figure 4.6: Density plot of impacting inner 100 km of H_2O coma, shown 2.1 s after initial contact with lunar surface (0.25 s before nucleus impact). The compressed, shocked layer of vapor near the surface is described by densities $\rho \sim 10^{-11}$ g/cm³ or greater, near to the critical density for melting of lunar regolith grains (Eq. 4.2).

Figure 4.7: Density plot of impacting coma and nucleus, shown 3.0 after initial contact of coma with lunar surface (0.65 s after nucleus impact). At this time, ices within the nucleus have been fully vaporized and expand within the transient atmosphere generated by the coma collision. Note the changed density scale, compared with Fig. 4.6, which highlights the larger vapor densities within the nucleus plume. The radial (surface-parallel) flow established by the colliding inner coma plasma, combined with partial containment of the nucleus vapor plume as its energy is coupled

to the surrounding atmosphere, promotes enhanced surface interactions downrange from the impact. This effect, combined with greater total vaporization volumes, allows cometary impactors to scour the lunar surface more efficiently, and over larger distances, than asteroidal impactors.

Figure 4.8: Evolution of dynamic pressure within the vapor plumes generated by asteroidal ($v = 15$ km/s, top) and cometary ($v = 60$ km/s, bottom) impactors. As the projectile penetration timescale (and the timescale of plume expansion) is inversely related to impact velocity, comparative asteroid results are plotted at times scaled by the factor of 4 difference in impact velocity. The comet impact's larger vaporization volumes and higher plume internal energies, both driven by the impactor's high velocity, result in an elongated downrange zone of plume interaction with the lunar surface, in which dynamic pressures remain high. In contrast, materials vaporized by the asteroid are seen to expand and disperse away from the lunar surface at relatively near-impact distances, limiting the extent of surface modification.

Figure 4.9: Temperatures of vapor plumes produced by asteroidal (15 km/s) and cometary (60 km/s) impactors. As the timescale of projectile penetration and plume expansion is inversely related to impact velocity, results for the asteroid at $t = 0.3$ s are compared to the comet at $t = 0.075$ s. Initial lunar regolith temperature is 200 K; blue color corresponds to 300 K and red color corresponds to temperatures in excess of 10,000 K. The comet-generated vapor plume is both more massive (see Table 4.1) and possesses higher internal energy, as evidenced by the large fraction of material heated to greater than 10,000 K. The asteroid-generated plume is described by somewhat lower temperatures (a few 1000 K) and is less massive by about a factor of five. The high internal energy of the comet vapor plume contributes to enhanced downrange ablation and scouring, as it expands at greater velocities and higher tem-

peratures than the asteroidal plume.

Figure 4.10: Hydrodynamic jet impingement against a surface produces complex streamline patterns, through the development of a recirculation or stagnation bubble. Streamlines were imaged (right) using the grease streak method. Transition between regions of inward and outward flow are shown in sketch on left; dotted lines map this transition zone to the experimentally observed streamlines. This impact of dust- and ice-enriched cometary jets with the lunar surface should establish similar flow patterns and mobilize lunar regolith particles in an analogous process. Figure modified from Donaldson and Snedeker [1971].

Figure 4.11: Particle-laden supersonic jets, upon impingement at a wall, generate radially-flowing “wall jets”, in which particles are swept parallel to the solid surface. The development of a recirculation flow (see Fig. 4.10) allows clouds of particles, which are reflected from the wall upon impact, to cluster within the flow. These particles can locally shield the surface from additional particle bombardment or eventually become mobilized in the radial flow. This figure, modified from Klinkov [2005], shows measured particle velocities within an impinging particle-laden jet. Similar particle flow may occur during the impact of a comet’s dust-laden inner coma with the lunar surface. Ejecta generated by impacting dust will load the near-surface flow with additional particles.

Figures

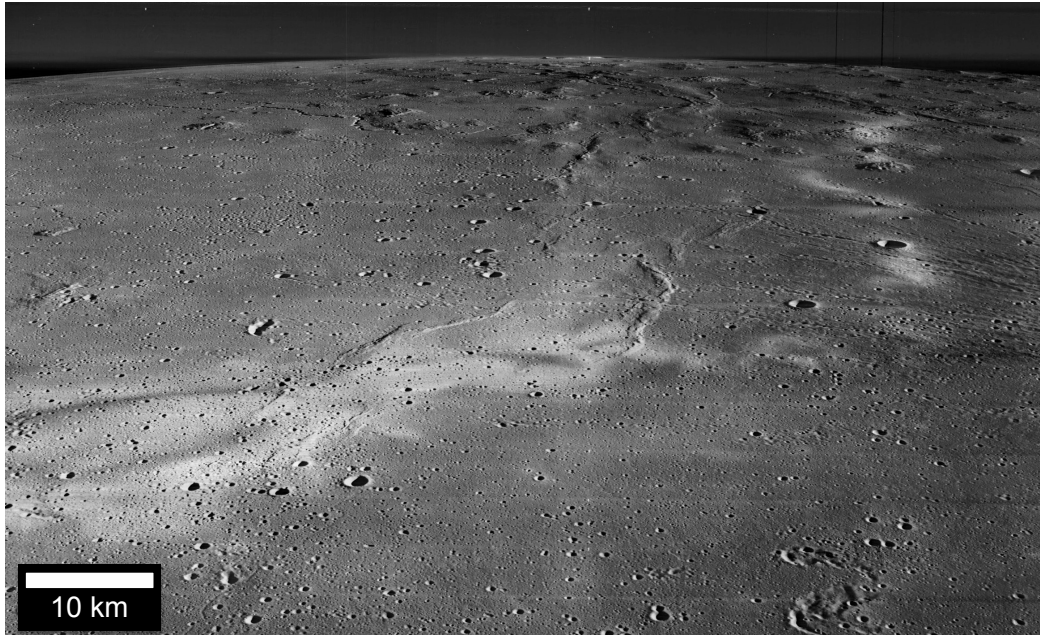


Figure 4.1

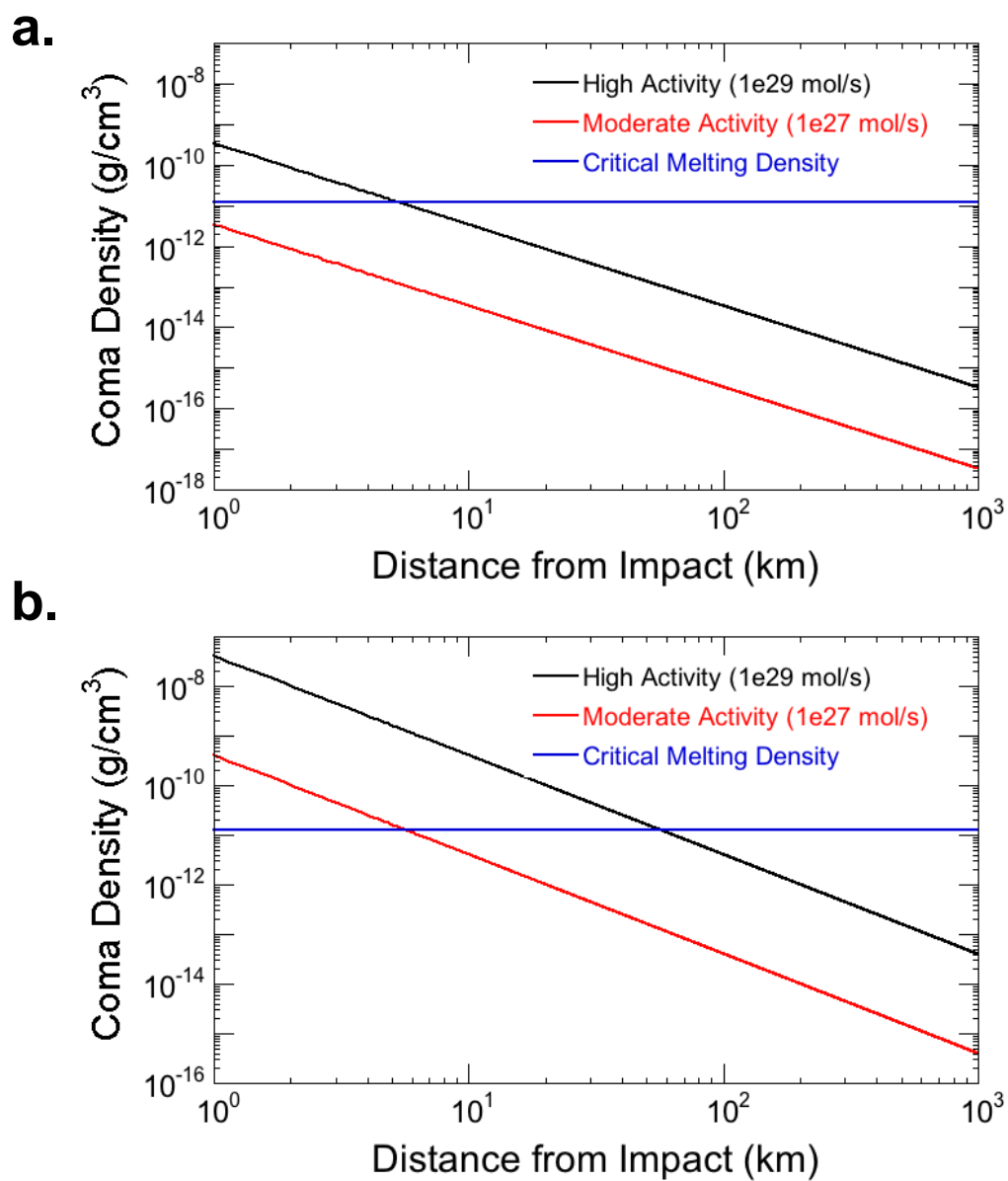


Figure 4.2

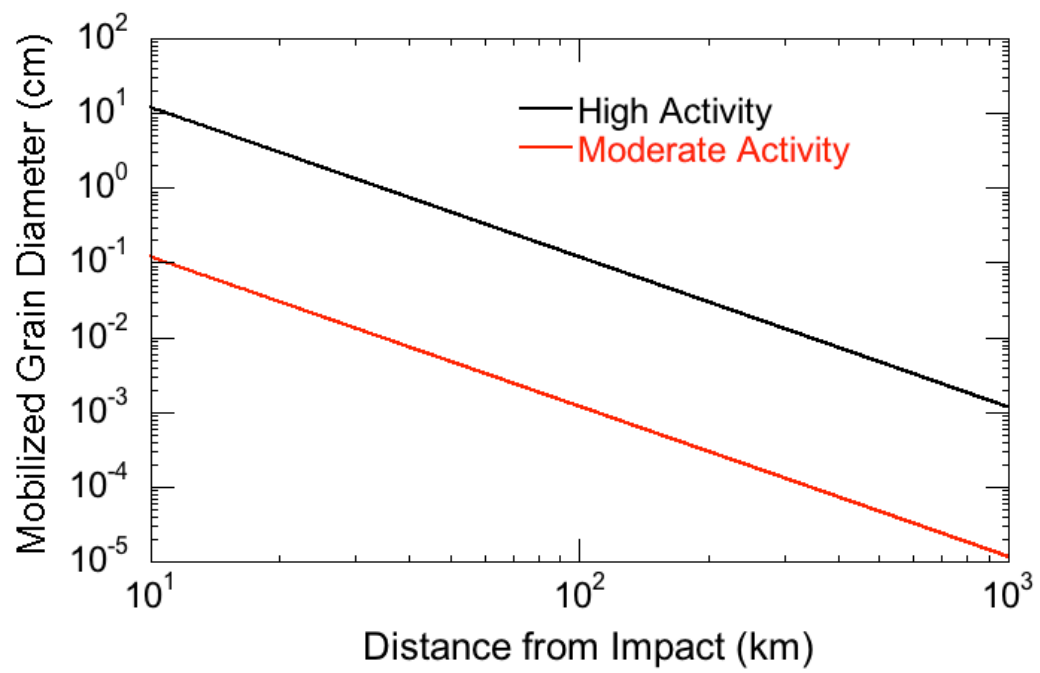


Figure 4.3

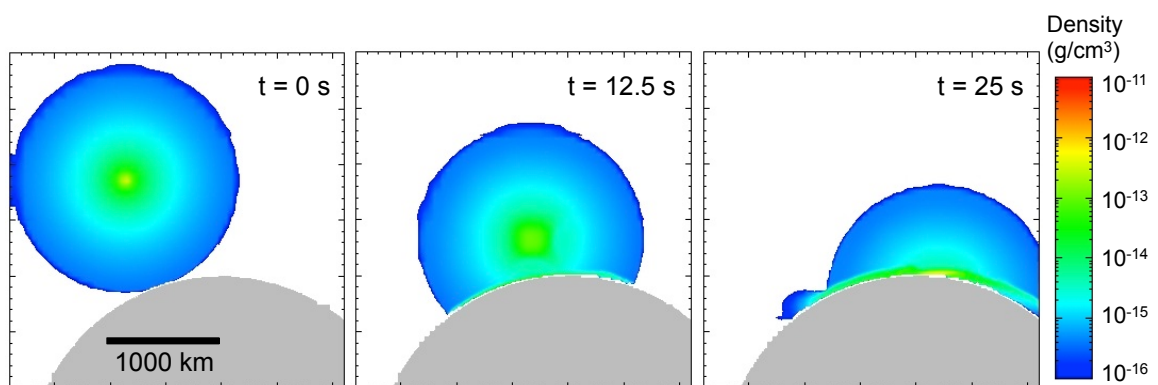


Figure 4.4

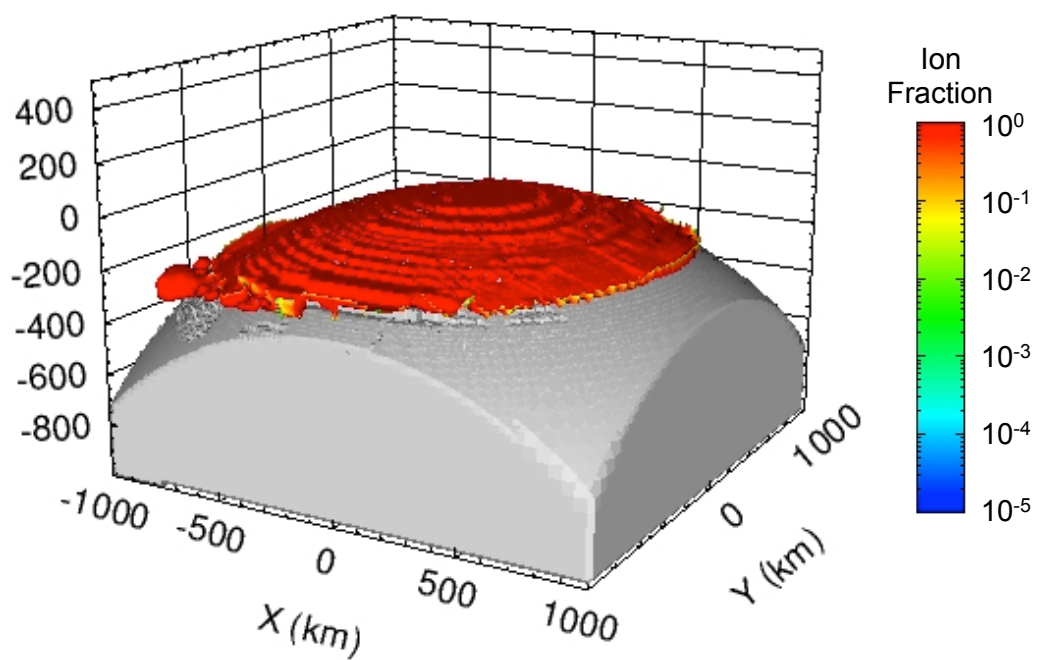


Figure 4.5

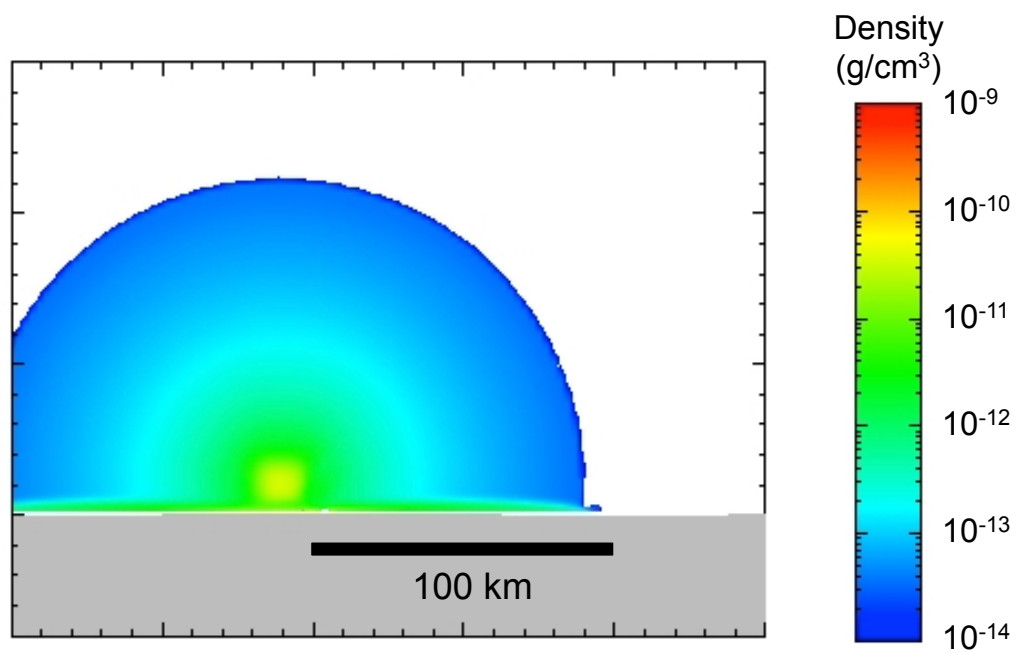


Figure 4.6

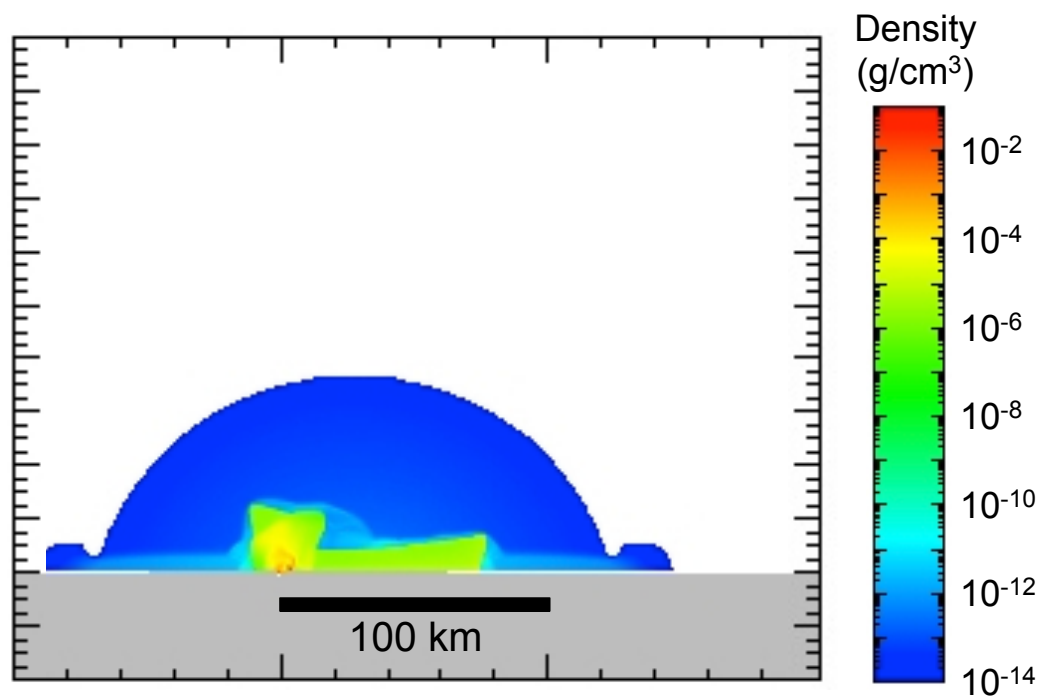


Figure 4.7

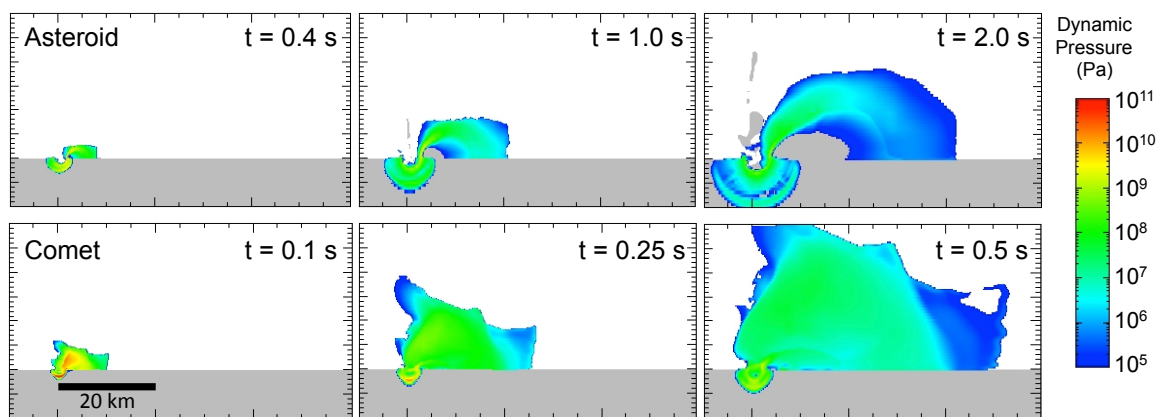


Figure 4.8

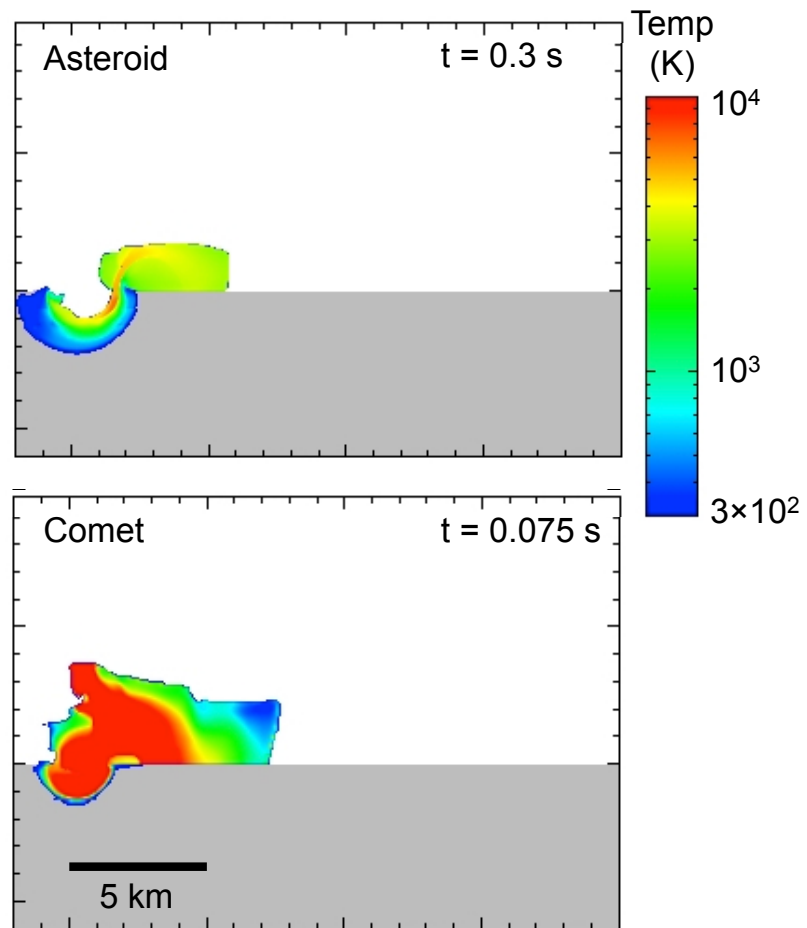


Figure 4.9

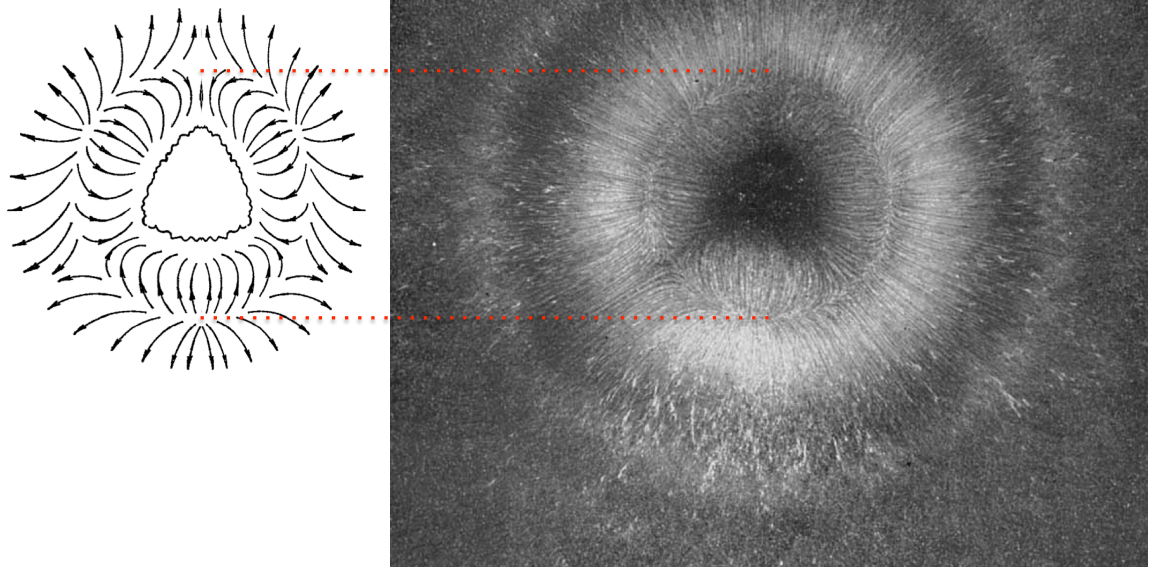


Figure 4.10

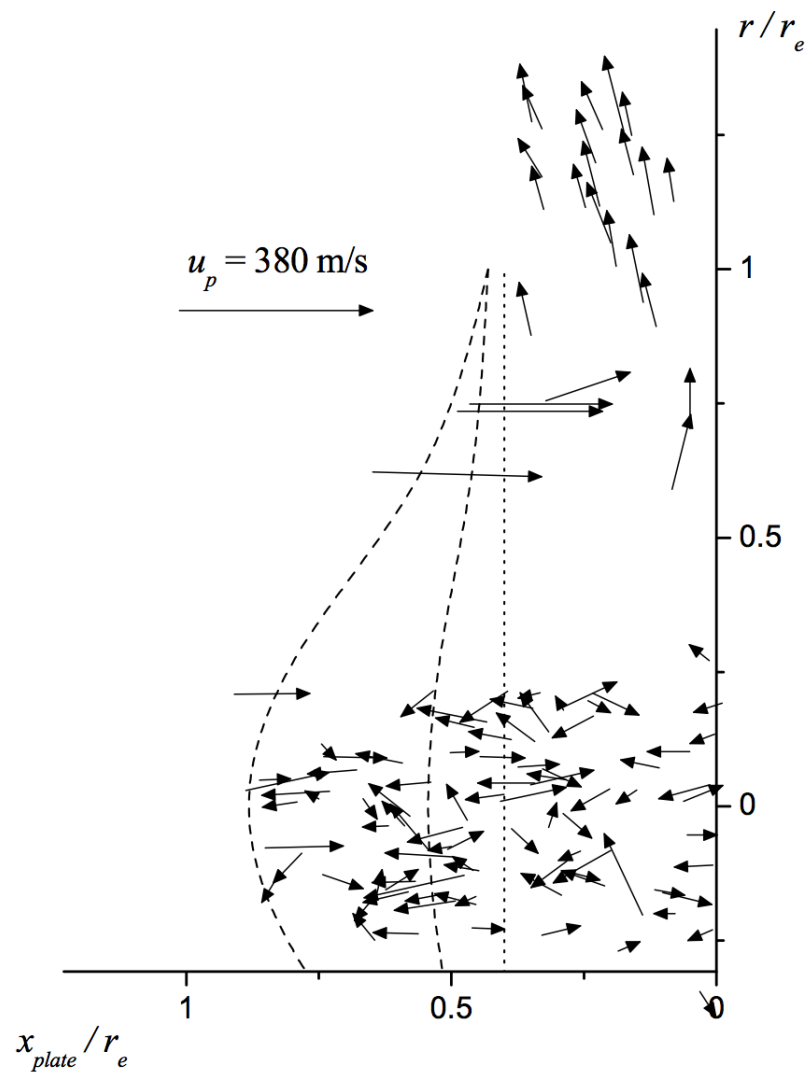


Figure 4.11

Chapter 5. Spatially-Resolved Spectroscopy of Impact-Generated Vapor Plumes

Megan Bruck Syal and Peter H. Schultz

Department of Geological Sciences, Brown University

Providence, RI 02912

Abstract

High-speed spectroscopy of impact vapor plumes is a key experimental strategy for constraining some of the earliest time and hottest processes occurring during hypervelocity impact. Spectroscopic methods are unique in their ability to characterize compositions and temperatures non-intrusively within transient impact-generated vapor. These measurements provide insight into a diverse set of problems, ranging from the processing of planetary atmospheres to the identification of meteorite types by their emission spectra. Here we report on experiments utilizing a recently upgraded spectroscopic system at the NASA Ames Vertical Gun Range (AVGR). Up to six regions of interest within the vapor plume are targeted by separate spectrometers during an impact experiment, thereby allowing spatially-resolved information to be obtained from the plume. Simultaneous operation of cameras (1 million frames per second) provides a complete view of the plume evolution. Observations of target-derived vapor, using easily volatilized particulate dolomite targets, contrast with the plumes arising from projectile-derived vapor (crystalline dolomite projectiles). Results demonstrate the evolution in time and space of spectral content, including Ca, Mg, CaO, MgO, Na, and CO emission. Derivation of temperatures through both relative emission line intensities and blackbody fitting to the background radiation allow cooling sequences to be tracked. Additionally, impact vaporization of a hydrated silicate projectile (serpentine) revealed both atomic (Fe) and molecular (FeO) emission lines arising from the iron content in serpentine. Achieving iron vaporization at these relatively low ($v \sim 5$ km/s) impact velocities has ramifications for the space weathering history of the Moon and other airless bodies. Additional experiments and analysis will provide more detailed information on the plume's thermodynamic state, with the goal of refining existing impact vapor plume theory.

5.1 Introduction

Impact-generated vapor plumes, observed at both planetary [A’Hearn et al., 2005; Schultz et al., 2007; Colaprete et al., 2010; Schultz et al., 2010] and laboratory scales [Schultz, 1996; Sugita et al., 1998; Sugita and Schultz, 1999, 2003a], are produced during the earliest stages of hypervelocity impact, as a combination of projectile and target material undergoes irreversible heating through shock unloading [Zel’dovich and Raizer, 1966] and additional, dissipative processes [Schultz, 1996]. The initial partitioning of impactor energy into the vapor plume is of central importance to numerous planetary problems of interest, including: delivery of projectile material (e.g., volatiles or organics) [Moses et al., 1999; Pierazzo and Chyba, 2002; Ong et al., 2010], target degassing [Schultz and Gault, 1990; Pierazzo et al., 1998], and the impact erosion of planetary atmospheres [Cameron, 1983; Vickery and Melosh, 1990; Shuvalov, 2009]. The thermodynamic state of the vapor plume, particularly the evolving chemistry and temperatures during vapor expansion and cooling, inform models of early solar system processes, such as the Earth-Moon forming impact [Pahlevan and Stevenson, 2007] and chondrule formation [Krot et al., 2005]. Condensation of impact spherules, important tracers of past meteoroid impacts at Earth, is also controlled by the vapor plume cooling sequence [O’Keefe and Ahrens, 1982; Johnson and Melosh, 2012]. Planetary surface materials are also processed through ongoing impact vaporization. For instance, the generation and deposition of impact vapor by micrometeorite bombardment makes important contributions to space weathering effects at bodies without atmospheres [Keller and McKay, 1993, 1994]. In addition, the evolution of cool vapor phases recognized in prior spectral studies may provide clues for the delivery, sequestration, and release of impact-delivered volatiles [Schultz, 2011].

Details of the vapor expansion process are often simplified into a one-dimensional, isotropically-expanding plume model, originally detailed in Zel’dovich and Raizer

[1966]. Construction of self-similar solutions requires an *a priori* assumption for the plume’s initial velocity distribution, after which the remaining flow variables may be calculated, under various assumptions. However, impact plumes observed in experimental studies are far from isotropic, even in the case of vertical impacts (see Fig. 5.1). For oblique impacts, which dominate the bombardment histories of planetary systems [Gilbert, 1893; Shoemaker, 1962], plume morphology deviates even more strongly from the one-dimensional approximation. Analytical models of vapor plume evolution do not distinguish between impactor- and target-derived vapor; moreover, impactor energy is generally assumed to be partitioned proportionally to each component, e.g., [O’Keefe and Ahrens, 1982; Moses et al., 1999; Ong et al., 2010; Johnson and Melosh, 2012]. Differing degrees of shock heating for projectile and target, combined with the initial confinement and release of vapor during the penetration stage, may make this approximation too simple. These observations highlight the importance of additional experimental work to constrain working models of impact vapor plumes.

Extracting quantitative data from experimentally-produced impact vapor plumes requires high-speed instrumentation; for typical laboratory-sized projectiles ($d = 6.35$ mm), vapor plume formation, expansion, and dispersal occurs over $\sim 50 \mu\text{s}$. While modern high-speed cameras can capture plume morphology with $1\text{-}\mu\text{s}$ time resolution, other ultra-fast measurement techniques are necessary to constrain thermodynamic variables of interest. High-speed spectroscopy, enabled by the fast optical gating and high sensitivity of intensified charge-coupled devices (ICCDs), permits non-intrusive probing of the rapidly evolving temperatures and compositions within the plume. Impact vapor plumes are well-known to emit a wealth of spectral information: black-body, molecular, and atomic emission all contribute to experimentally-observed vapor plume spectra [Sugita et al., 1998]. Analysis of such spectral data provides powerful insights to the chemistry and physics within the plume.

Pioneering work on impact spectroscopy at the NASA Ames Vertical Gun Range (AVGR) focused on characterization of the jetting phase, which is the earliest and hottest component of vaporization [Sugita et al., 1998; Sugita and Schultz, 1999; Sugita et al., 2003]. These early studies, which utilized copper projectiles and crystalline dolomite targets, revealed that the jetting phase was described by temperatures ranging from 4000 K to 6000 K; jetting temperatures and the relative contributions of target and projectile material were found to depend upon impact angle. Additionally, the jetting phase was calculated to comprise a small fraction of the total mass of vaporized carbonate. Subsequent impact spectroscopy work analyzed projectile-derived vapor plume interactions with an ambient atmosphere [Sugita and Schultz, 2003a,b]; isolated the role of projectile failure in vapor production [Schultz et al., 2006]; demonstrated CN synthesis through impactor and atmospheric interactions [Sugita and Schultz, 2009]; supported interpretation of large-scale impacts, such as Deep Impact [Schultz et al., 2007]; assessed processes for the temporary trapping of lunar volatiles [Schultz, 2011]; and explored the controlling processes for vaporization [Schultz and Eberhardy, 2014].

Here we extend these established spectroscopic techniques in order to acquire spatially and temporally-resolved spectra of impact vapor plumes. Previous studies compared the spectral evolution of the vapor plume in different directions (above and inside the crater) with 5 μ s snapshots using a high-resolution, single-frame thermal camera [Schultz et al., 2007; Schultz and Eberhardy, 2014]. The present study integrates high-frame rate imaging (up to 1 million frames per second) with simultaneous spectral measurements of the evolving plume. This approach differs from prior work in several key respects: (1) A multi-fiber spectroscopic set-up, coupled with simultaneous high frame rate (2 μ s) imaging of the plume, allows careful tracking of spectral content in time and space. While previous studies have successfully targeted multiple spatial regions within a single impact event [Schultz and Eberhardy,

2014], integrating this information with early-time images of the plume provides new and clearer context for the observations. (2) Upgraded ICCD detectors (Andor iStar 340T) with 16-bit digitization and greater sensitivity allow detection of fainter, cooler molecular species. (3) The spatial areas targeted within the plume extend to much greater radial distances from the impact point (~ 30 cm) than previous measurements [Eberhardy and Schultz, 2003; Schultz and Eberhardy, 2014]. (4) Vaporization of carbonate, particulate targets (dolomite powder) by silicate projectiles is directly contrasted with the vaporization of a dolomite projectile upon impact with a silicate target. An additional, new projectile material, serpentinite, is also utilized to isolate projectile vaporization.

Results from these experiments provide critical ground truth not only for the refinement of analytical and numerical approaches to impact vaporization problems but also for the design and interpretation of missions utilizing kinetic probes. Impact excavator missions, including Deep Impact [A’Hearn et al., 2005; Schultz et al., 2007] and LCROSS [Colaprete et al., 2010; Schultz et al., 2010], are gaining traction as a low-cost, rugged strategy to generate high scientific return. However, optimal engineering of this subsurface exploration approach, along with accurate interpretation of spectral observations obtained during these large scale impact experiments, depends upon additional fundamental work on impact vapor plume development.

5.2 Experimental Approach

Experiments to assess the spectral content of impact-generated vaporization were carried out at the NASA Ames Vertical Gun Range (AVGR), a two-stage light gas gun in Moffett Field, CA. Features of the AVGR facility that are key to this study include its large (2.5 m diameter) target chamber, which allows free expansion of impact vapor, and the ability to shoot into particulate targets at a variety of incidence angles. All experiments were conducted under near-vacuum conditions, $P < 0.8$ mbar,

and impact velocities were maintained at $v \sim 5$ km/s (see Table 5.1 for detailed list). Impact angles for results reported in this study include 30° , 45° , and 90° from the horizontal.

5.2 Target and Projectile Properties

Pyrex projectiles ($d = 6.35$ mm) impacting powdered dolomite targets [$\text{CaMg}(\text{CO}_3)_3$] isolated target-derived vapor; Pyrex contributes minimally to spectral content while dolomite is known to vaporize readily at laboratory velocities and produce rich spectral content in the optical region. Powdered dolomite, although it produces somewhat less luminous emission than crystalline dolomite [Schultz and Eberhardy, 2014], was chosen to provide a more physically realistic proxy material for planetary surfaces, which are often porous and particulate. These material properties affect not only the initial generation of vapor but also its eventual expression at the surface.

In addition to the nominal “half-space” target geometry (impact into the center of a 59.5 cm diameter bucket filled with target material), dolomite powder experiments were also conducted in a “quarter-space” target configuration. Quarter-space experiments provide a cross-sectional view of the impact event through a clear acrylic window, thereby minimizing contributions from projectile disruption [Eberhardy and Schultz, 2003; Schultz and Eberhardy, 2014]. The projectile trajectory is aligned parallel with the window, with impactor penetration occurring within a projectile diameter of the window edge. This effectively splits the impact event in half, down the transient crater’s axis of symmetry. Quarter space techniques allow direct acquisition of spectra from vapor inside of the plume, rather than peering through the entire hemisphere of vapor. Using both approaches provides complementary information.

Experiments focused on the vaporization of projectile materials utilized crystalline dolomite slugs ($d = 6.35$ mm; rounded on the end making first contact) impacting airfall pumice powder ($>70\%$ SiO_2 , $\rho = 1.5$ g/cm³), effectively reversing the silicate-

into-carbonate conditions of the target-focused experiments. These conditions produce a purely projectile-derived vapor plume, as pumice is not readily vaporized at laboratory velocities. Airfall pumice is a particularly useful material for the simulation of planetary regolith properties, due to its compressibility and varied particle texture, with irregularly-shaped grains ranging from 10-100 μm [Schultz et al., 2005]. Information from projectile-derived vapor was also obtained from the impact of a serpentinite $[(\text{Mg,Fe})_3\text{Si}_2\text{O}_5(\text{OH})_4]$ slug ($d = 6.35 \text{ mm}$) into pumice powder.

5.2 *Measurement Techniques*

High frame-rate, visible grayscale cameras provided side and top views of vapor plume development and expansion with 2 μs time resolution. The side-mounted camera was aligned with the side-viewing fore-optics of the spectrometers, for the acquisition of spatially coincident context images. Collection of light for spectroscopic analysis was achieved with six separate telescopes, each providing a 2.5 cm field of view at the impact plane. Fiber optics coupled the focused light signal from the telescopes to a pair of McPherson Czerny-Turner monochromators (each fed by three telescopes), described by 0.35 m focal lengths and $f/4.8$ effective apertures.

Diffraction gratings of 300 lines/mm were selected, providing maximal spectral range (200 nm) and 0.2 nm spectral resolution. While higher spectral resolution is achievable with higher density gratings, 0.2 nm represents significant improvement over previously published studies on spectral emission from dolomite targets [Sugita et al., 1998; Sugita and Schultz, 1999, 2003a]; results from this work demonstrated that resolutions of $\sim 0.9 \text{ nm}$ are sufficient to calculate vapor phase temperatures. The monochromators were tuned to focus on the 440-640 nm wavelength region, which contains an array of emission features from Ca, Mg, CaO, and MgO, the primary vapor phase products from impacts into dolomite.

Spectra were recorded by two Andor iStar 340T ICCD detectors (2048x526),

mounted to each monochromator. The detectors were operated in multi-track mode, enabling the assignment of detector row ranges to different fiber optic input. Hence, each ICCD records three simultaneous spectra from three different spatial regions within the impact vapor plume, each targeted by separate fore-optics. A photodiode signal or a time pulse from projectile velocity measurements triggered each detector; inherent variation in the precision of each method introduces an initial uncertainty in the acquisition times for a given impact. However, post-impact comparison of trigger pulses with the 500,000 frames/s images allows the timing of spectral acquisition to be known to within $< 2 \mu\text{s}$. Depending on the impact conditions, the duration of ICCD exposures varied from 10 to 100 μs ; trigger delays, when desired (e.g., for imaging late-time vapor) were inserted through a built-in digital delay generator.

Calibration was carried out manually for each fiber. Oriel spectral calibration lamps (Ar, Xe, Ne, and Hg(Ar)) provided strong emission lines across the entire observed spectral range for wavelength calibration. Calibration was applied using a cubic interpolation scheme between known strong emission features. In-chamber observations of a Labsphere integrating sphere source were used to apply spectroradiometric corrections to each fiber. While absolute radiative intensities (spectral power, W/nm) can be calculated, we take the conservative approach of reporting intensities in arbitrary units (a.u.), which are sufficient for extraction of compositional and temperature information and for comparison of relative intensities between different fibers.

5.3 Results

5.3 *Excitation Temperature*

Where possible, vapor temperatures have been estimated using relative intensities of neutral calcium emission lines to derive excitation temperatures, as described in detail in Sugita et al. [1998]. In general, this technique is more readily applied to

very hot plasmas, such as the material contained within the jetting phase, than to cooling, later-time vapors, which are dominated by molecular emission and contain fewer atomic emission lines. The reader is referred to Gaydon and Wolfhard [1970] and Griem [1997] for details of the line ratio method and to Sugita et al. [1998] for specifics on the method's application to impact-vaporized calcium, but a brief overview is given here.

Under the assumption of thermodynamical equilibrium, a Boltzmann distribution will apply to the number of atoms, N_u , in a given energy state u :

$$N_u = \frac{g_u}{g_o} \exp\left(-\frac{E_u}{kT}\right) N_o \quad (5.1)$$

where g_u and g_o are the statistical weights of the u and ground energy states, respectively, E_u is the energy of the u state, k is the Boltzmann constant, T is the temperature, and N_o is the number of ground state atoms. A linearized relation between a normalized line emission intensity, $\hat{I}_{lu}$ (where l and u refer to lower and upper electron energy states) and temperature can then be defined:

$$\ln \hat{I}_{lu} = \ln N_o - \frac{E_u}{kT} \quad (5.2)$$

where

$$\hat{I}_{lu} = \frac{I_{lu}}{\frac{h\nu_{lu}}{4\pi} \frac{g_u}{g_o} A_{lu}} \quad (5.3)$$

Here h refers to Plank's constant, ν_{lu} is the frequency of light emission, and A_{lu} is the Einstein A coefficient of the transition. If a sufficient number of emission lines are observed within the targeted wavelength range, a plot of normalized emission intensities against the corresponding E_u/k values can be used to derive an excitation temperature. An example plot is shown in Fig. 5.2, in which a linear fit yielded a temperature estimate of 5900 K.

5.3 Dolomite Targets

A direct comparison between impacts into powdered and crystalline dolomite targets is instructive for investigating the effects of porosity on vaporization and for continuity with previous work. Fig. 5.3 and 5.4 display the time evolution of the vapor plumes and spectra from four different view spots. Each 2.5 cm view spot is distinguished by a circle of a different color, while the impact point is marked with a star. Note that the exposure times differ between some regions; this affects what portion of the plume is being sampled and the interpretation of results.

The jetting phase is expressed much more clearly in the block impact shown in Fig. 5.4, due to the clean coupling at the target-projectile interface. Projectile penetration into the porous dolomite target (Fig. 5.3) results in temporary containment of vapor within the early transient cavity, e.g., [Schultz and Eberhardy, 2014]; the vapor is released later at lower expansion velocities than in the nonporous, crystalline case. Downrange fibers positioned above the target avoid spectral contamination from the high-temperature jetting phase in order to focus on lower-energy vapor components. Hence, several commonalities exist between spectra from each experiment.

The spectra shown in Fig. 5.3a and 5.4a are taken from a point just 5 cm downrange from the impact point and 2 cm above the impact plane (red field of view or FOV). In both the crystalline and powdered cases, the high intensity emissions in the region resulted in saturation of the sodium doublet lines (589.0/589.6 nm), but this saturation does not affect interpretation of the other features. The well-known calcium oxide green and orange bands stretch from 547.3 nm to 556.0 nm and from 598.3 nm to 636.2 nm, respectively [Gaydon, 1955; Pearse et al., 1976]. As these exposures are relatively long, the individual band heads are not resolved but, instead, contribute to a broader continuum of emission features. The green system of magnesium oxide bands, $B^1\Sigma - X^1\Sigma$, are observed in the 480.15-500.73 nm and 514.7-520.6 nm regions. Relatively strong atomic magnesium emission lines at 516.73, 517.27,

and 518.36 nm are also likely contributing to the broad feature near 520 nm. Bands from CO are observed just red of the 589/589.6 nm sodium line, between 590 and 600 nm. Smaller contributions from CO are seen near 452 nm and 475 nm. While the lack of atomic emissions precludes derivation of a line-ratio temperature estimate, these spectra possess a significant blackbody radiation component. Least squares fits to the blackbody emission reveal that the nonporous dolomite vapor is significantly hotter, $T_{npor} \sim 4200$ K, than the porous dolomite, $T_{por} \sim 3300$ K.

Slightly farther above the target (Fig. 5.3b and 5.4b, orange FOV), the impact into the porous target results in more CaO emission than in the solid target. Note that observed atomic and molecular species are only labeled in the topmost plots of Fig. 5.3 and 5.4. Farther downrange, in the green FOV (Fig. 5.3c and 5.4c), contributions to CaO emission further decay. Relatively more emission is observed in the porous case, partly due to the geometry of the plume. Lastly, Fig. 5.3d and 5.4d (yellow FOV) shows spectra obtained uprange from the target. While the emission spectrum was better defined for the nonporous case, the CaO emission in this region was strong, relative to the sodium doublet, in both targets. Based upon previous work [Sugita et al., 1998, 2003], this may suggest that the uprange component is somewhat cooler than the other sampled plume regions, as documented at longer wavelengths in other experiments at lower spectral resolution [Schultz, 2011].

Depending upon experimental conditions, the molecular emissions are resolvable into discrete band heads, rather than the broad features shown in Fig. 5.3 and 5.4. Spectra from a 30° impact into a dolomite powder (half-space) target are shown in Fig. 5.5, again with a time sequence of context images. Spectra obtained from this experiment were much “cleaner,” with relatively little blackbody emission, scatter, or blending between emission lines. Interpreting the spectra requires focusing on a subset of the entire range, since narrow features resolved with 0.2 nm resolution appear crowded when displayed over the entire 200 nm wavelength range. A general

trend with this vapor plume sequence is the predominance of CO emission within the more rarified regions uprange and downrange from the impact point (panels a and d), whereas CaO emission dominates in the denser regions, nearer to the impact point

A quarter-space experiment (Fig. 5.6) complements the half-space experiments described above (Fig. 5.5) for the same impact conditions but with slightly different fields of view. The right-most panels of Fig. 5.6 show some of the more detailed molecular structure with these regions, displaying sequences of both MgO and CaO band heads. Fig. 5.6a (purple FOV) shows how the spectrometer captured the very leading edge of the plume at $t = 20\text{-}30\ \mu\text{s}$. Here, the spectrum was clean enough and contained enough calcium emission lines to extract a Boltzmann temperature estimate, calculated to be $T \sim 3300\ \text{K}$. Later ($t = 60\text{-}80\ \mu\text{s}$), the leading edge of the vapor plume passed through the blue FOV (Fig. 5.6c) and had cooled to $2600\ \text{K}$ (based on the line ratio method). Also notable here are the evolving relative band strengths of the MgO and CaO emission; CaO emission is most pronounced at earlier times, while MgO emission becomes relatively stronger as the vapor plume cools and expands. This effect could be due to either preferential condensation out of the plume or temperature differences in the vapor phase. Closer to the impact point (Fig. 5.6b, red FOV) the leading edge of the plume had already passed. Now a much stronger blackbody component (with greater Mie scattering) dominates, along with stronger molecular emission (CaO and CO). The background radiation is fit by a $T \sim 1500\ \text{K}$ blackbody curve, well below the vapor temperatures estimated; incandescent melt, some of which may have condensed out of the vapor as droplets, is the likely source of this blackbody emission.

A comparative case for Boltzmann distribution-determined temperature is shown in Fig. 5.2, which depicts the Boltzmann plot for a 90° quarter-space impact into powdered dolomite. As peak pressures are highest for vertical incidence impacts, higher vapor temperatures are expected; this trend has been characterized spectroscopically

in prior studies [Sugita et al., 1998]. Here the atomic emissions were sufficient to derive a robust temperature estimate of $T \sim 5900$ K, consistent with previous work and significantly greater than estimates from the 30° experiment shown in Fig. 5.6.

5.3 Dolomite Projectile

The experiments described above focused on emissions created by a refractory projectile (Pyrex) impacting into volatile-rich dolomite. Another experiment reversed these conditions: a carbonate projectile impacting into a silicate target. The vapor plume generated by the dolomite projectile is shown in Fig. 5.7, with various spectra from the plume displayed. While aspects of the overall plume composition were similar to target-derived plumes, e.g., strong Na and green/orange CaO emission, other differences include the absence of discernible MgO bands and a relatively low black-body component throughout the various FOV's. The plume shape and size was also markedly different from target vapor plumes; Fig. 5.8 compares plumes generated in 30° impacts from a dolomite projectile (top) and a dolomite target (bottom).

5.3 Serpentine Projectile

A serpentine projectile launched into airfall pumice at 45° produced a luminous plume from which several interesting spectra were recorded. The iron content of serpentine $[(\text{Mg,Fe})_3\text{Si}_2\text{O}_5(\text{OH})_4]$ was anticipated to contribute to features distinct from the dolomite spectra. Within rarified portions of the plume, both uprange and downrange from the impact point, a unique band system of many densely-packed emission features in the 580-600 nm region was recognized (Fig. 5.9). The system, after careful comparison with molecular spectroscopy literature, was positively identified as the “orange bands” from iron oxide [Pearse et al., 1976; West and Broida, 1975; Cheung et al., 1981, 1982, 1983]. The FeO band structure is particularly complex, with the strongest emission lines occurring in the 580-600 nm region. Within

the entire region of 560-600 nm, however, many different systems overlap. The orange bands of FeO, which were long debated, are now known to arise from a strongly perturbed $^5\Delta_i - ^5\Delta_i$ electronic transition. Fig. 5.9a also shows many atomic Fe emission lines, identified alongside the FeO vapor emission. The vaporization of Fe at laboratory velocities is unexpected, based upon theoretical critical pressures to achieve iron vaporization. The implications for this observation will be discussed in the next section.

5.4 Discussion

Measuring the spectral content of vapor plumes within cooler regions, outside of the plasma-dominated jetting phase, is complicated by several processes, including lower luminosities, condensing droplets, and the lack of atomic emission lines. However, the sensitivity and spectral resolution of our recently upgraded spectroscopic system allows for the identification of previously difficult-to-resolve band heads within molecular emission bands, such as the MgO green system and the CaO green and orange systems. Modeling and fitting these emission features with synthetic spectra will be necessary to derive more robust temperature estimates for many of the cooler vapor regions, which lack sufficient atomic emission for the construction of a Boltzmann temperature plot. Building synthetic spectra for diatomic molecules such as MgO and CaO, for which the various molecular constants are less well characterized than more commonly modeled systems (e.g., C_2 , OH), will require additional theoretical development; this will be a central objective of future work on the project.

The temperature and compositional information revealed in this initial study is instructive, even in the absence of full synthetic spectra calculations. For instance, the higher blackbody temperatures associated with vapor from nonporous dolomite than from porous dolomite (Fig. 5.3 and 5.4) are likely related to the way in which vapor is generated in the penetration cavity of porous materials. While the PdV

work during compression of pore space contributes to additional irreversible heating [Zel'dovich and Raizer, 1966] and, hence, greater vaporization efficiency (greater total vaporized mass), suppression of the jetting phase during the elongated projectile penetration stage results in a vapor plume described by lower temperatures upon release at the surface. The blackbody component for these experiments, which sampled plume chemistry over relatively long timescales, is likely produced by incandescent melt, rather than optically thick vapor. Early condensation products contributing to background radiation may include refractory materials such as carbon.

The results shown in Fig. 5.5 for a 30° impact into dolomite powder (half-space) are suggestive of a transition from CaO-dominated vapor to CO-dominated vapor as the plume moves downrange. The strength of the green and orange CaO bands gradually decays as the plume expands and is imaged at progressively more distal locations, while CO emission is still present. On these time scales, CaO may be forming nucleation centers and condensing out of the plume, limiting contributions at later times [Zel'dovich and Raizer, 1966].

The cooling sequence depicted in Fig. 5.6 provides a complementary quarter space example for the impact conditions shown in Fig. 5.5. Here, the timing of the spectra with the edge of the vapor plume provided more useful information, as the spectral signature was not overprinted with extraneous blackbody or scattering particles. While this event was one of the few for which temperature estimates could be made, it illustrates proof of concept for the tracking of cooling vapor species. The middle (red FOV) spectra from Fig. 5.6, which was taken inside the plume, after the plume's luminous edge had passed ($t = 60\text{--}80 \mu\text{s}$), shows evidence for the condensation of droplets through both an increased blackbody component and the addition of scatter to the spectrum. At these wavelengths, condensing particles, which should be on the order of a few μm in diameter [Zel'dovich and Raizer, 1966], would contribute to Mie scattering in the spectra.

The production and analysis of impactor-derived vapor plumes, using dolomite and serpentinite projectiles, provides new insight on projectile vaporization arising from geologically relevant materials; prior studies have used proxies such as copper [Sugita and Schultz, 1999] or Lexan polycarbonate [Sugita and Schultz, 2003a] to investigate projectile vaporization. A direct comparison between the target- and impactor-derived vapor plumes from 30° impacts, shown in Fig. 5.8, demonstrates subtle differences between the two types of vaporization. The dolomite projectile vapor appears more luminous than the dolomite target vapor, with bright regions of vapor distributed throughout the plume. In contrast, emitting vapors are concentrated along the leading edge of the target-derived vapor plume. The dolomite projectile’s luminosity also persists to longer times. This may evidence higher vapor phase temperatures in projectile vapor plumes, which is consistent with higher peak pressures and shear heating occurring in the projectile than in the target. An additional distinction is the apparent detachment of the projectile plume’s leading edge from the remaining plume; this luminous edge is seen to expand vertically with time, while the target-derived plume’s leading edge exhibits more typical jetting behavior, remaining near to the surface. These differences in post-impact vapor flow characteristics are important when considering the mass of vaporized material remaining below escape velocity for a given impact.

Spectra obtained of the dolomite projectile plume (Fig. 5.7) demonstrated the presence of many of the same species seen in dolomite target plumes: sodium doublet (589.0/589.6 nm), CaO green and orange bands, and CO. While temperature estimates could not be derived from the available spectra, relatively little blackbody radiation suggests that these vapor plumes are hotter than the target-derived plumes. Also, the lack of any MgO features, which tend to become more visible at later times in the plume cooling sequence (Fig. 5.5 and 5.6), may also suggest hotter temperatures. Obtaining temperature estimates from projectile vapor will be a high priority

for future experiments.

Serpentine is a particularly interesting projectile material for vapor plume investigations, due to its significant water content. Impact delivery of water to terrestrial planets will depend upon the way in which the water content of impactors is heated during an impact; hydrated silicates (from impacting carbonaceous chondrites) may have made substantial contributions to the water reservoirs of both the Earth and the Moon [Harris and Schultz, 2011; Saal et al., 2013]. Accurate modeling of mixed materials is a significant challenge for numerical approaches, which rely on mixed equation of state models to approximate the problem. Hence, experiments to quantify post-impact temperatures for various molecular species within the plume could provide novel insight into the problem of water delivery from a heterogeneous mineral mixture.

Although atomic iron emission lines have been spectrally resolved by previous impact experiments at the AVGR [Adams et al., 1997], the serpentine experiments provided the first identification of FeO molecular emission in an impact experiment. Additionally, the iron observed by Adams et al. [1997] arose from target materials (meteoritic compositions), rather than the projectile. In addition to the FeO emission, several atomic iron lines were also present in the serpentine spectra. This is significant for several reasons: (1) using critical pressure criteria, the incipient vaporization of iron should begin at velocities greater than ~ 13 km/s [Kraus et al., 2014], an estimate recently revised downward based upon new experimental results; (2) generation of metallic iron through impact vaporization is an important contributor to space weathering effects at atmosphereless bodies.

The identification of iron and iron oxides in experiments demonstrates that iron may be vaporized at much lower velocities, $v \sim 5$ km/s, suggesting that predictions based upon critical pressure may under-estimate vaporization efficiency. The usage of other metrics to numerically quantify vapor production, including temperature

and entropy [Quintana et al., 2013], may be particularly important at the lower impact velocities typical of laboratory studies. Other factors contributing to enhanced vaporization of iron may include the phyllosilicate structure of serpentine, which could preferentially partition shock energy into the irreversible heating of iron, or a larger role for frictional heating in oblique experiments [Schultz, 1996].

The ease with which iron is vaporized, even at these relatively low velocities, has important implications for space weathering processes at the Moon, where sub-micron metallic iron, deposited on the rims of soil grains, contributes to reddening, darkening, and weakening of spectral features [Pieters et al., 1993; Pieters et al., 2000; Hapke, 2001]. The mean impact velocity at the Moon, $v \sim 17$ km/s [Yue et al., 2013], is more than sufficient to generate vaporized iron. For the micrometeoritic impacts that dominate regolith processing, iron vapor will condense into small, nanometer-sized droplets. Since vaporized mass scales as $m_p \sim v^2$, e.g., [Bjorkman and Holsapple, 1987], the larger velocity impacts will dominate vapor production and deposition on nearby grains. Using analytically-derived scaling relations for spherule production in a silica vapor cloud [Johnson and Melosh, 2012], the estimated spherule size for a 1 cm meteorite impacting at $v = 20$ km/s is $d \sim 0.1$ nm, consistent with the nm-sized metallic iron particles found in amorphous rims of lunar soil grains [Keller and McKay, 1993; Keller and McKay, 1997]. However, as the iron equation of state differs from the SiO_2 equation of state used to derive the scaling relations in Johnson and Melosh [2012], this estimate size may differ slightly for the case of metallic iron.

The results from this study also underscore interpretations of meteoritic emission spectra, which display many of the same features when imaged upon entry to Earth’s atmosphere. The FeO orange bands, in particular, have been recognized relatively recently within the Earth’s night glow [Evans et al., 2010; Saran et al., 2011]. The feature, which was also associated with sodium emission (as seen in the experiments reported here), was attributed to meteors, a hypothesis born out by the impact pro-

duction of FeO from serpentine in these experiments. FeO emission has also been observed within the persistent trains of Leonid meteors [Jenniskens et al., 2000]; the presence or absence of this feature or other iron line emissions has the potential for discriminating between different meteor types. Recent emission spectra obtained from super-bolides, assumed to be cometary from their trajectories, contain many of the molecular species probed in the present study, including: Na, Fe, Ca, and Mg lines [Madieto et al., 2014]. These observations demonstrate yet another application of impact spectroscopy: the interpretation and classification of meteorite impacts through knowledge of their emission spectra.

5.5 Concluding Remarks

Integration of high-speed spectroscopy with high-speed imaging provides an increasingly clear picture of the early-time physics and chemistry occurring in impact-generated vapor plumes. Advancements in ICCD technology allow fainter species to be detected and permit the resolution of individual band heads within previously unresolved, broad molecular emission features. These new capabilities, in addition to greater spatial coverage of the plume with the spectral view spots, reveal the cooling sequence of vapor in both time and space, including the condensation of vapor behind the luminous leading edge of the plume. Additionally, improved spectral resolution allows the use of synthetic molecular spectra to constrain the evolution of much lower temperatures.

Geologically relevant impactors, including dolomite and serpentinite, were successfully tested and spectroscopically measured for the first time. Target- and projectile-derived vapor plumes (each arising from dolomite vaporization) exhibit similar but not identical emission spectra, hence indicating possible differences in the processes controlling vaporization of each material type. Impacts by volatile-rich projectiles into a silicate target generate a more self-luminous vapor plume than the plume

from a silicate projectile into a volatile-rich target. However, the downrange-directed velocity component of projectile-derived vapor is lower than the target-derived case; conversely, the vertically-directed velocity component of the projectile plume expands more rapidly than the target plumes. These differences have implications for the efficacy of impactor delivery to planetary surfaces.

Vaporized serpentinite, a newly-tested projectile material, contained FeO and Fe emission, despite impact velocities ($v \sim 5$ km/s) well below the theoretical limits for incipient vaporization of iron ($v \sim 13$ km/s). Iron vaporization and deposition is a key contributor to space weathering at the Moon: greater vaporization efficiency implicates an enhanced role for micrometeoritic bombardment in the darkening, reddening, and weakening of lunar regolith reflectance spectra (relative to solar wind implantation effects). Additionally, Fe emission can serve as key diagnostic feature in the spectra of meteors entering an atmosphere or striking a planetary surface. New knowledge of the impact-generated emission spectra from various projectile materials has direct application to placing improved constraints on the meteoritic flux at planetary bodies, both at Earth and elsewhere.

This study further represents substantial progress towards building a fuller thermodynamic characterization of impact vapor plumes. Shorter exposures isolating a parcel of evolving vapor will provide a more complete record of the rapidly evolving conditions within impact-generated vapor plumes. These results can then be compared with predictions from numerical and analytic models. Such a strategy, combined with more complete temperature data through synthetic spectra calculations, will build a comprehensive understanding of the vapor plume processes that numerical and analytical approaches cannot yet capture. The ultimate goal of such work is to refine the existing vapor plume theory, so that details including impact angle and target and projectile material properties can be accommodated when considering impact vaporization's effects upon the formation and evolution of planetary

systems. In addition, results from continued impact spectroscopy studies will provide a clearer understanding of regolith processing, assist in the classification of meteor types impacting planetary atmospheres or surfaces, and inform engineering strategies for collecting and analyzing spectral data from future kinetic probe missions.

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Tables

Table 5.1: List of Experiments

Shot	Target	Projectile ^a	θ	v (km/s)	m_p (g)
130815	Dolo (1/4) ^b	Pyrex	45°	5.02	0.298
130816	Dolo (1/4)	Pyrex	60°	5.07	0.298
130819	Dolo (1/4)	Pyrex	90°	5.28	0.298
130820	Dolo (1/4)	Pyrex	30°	5.35	0.298
140107	Dolo ^c	Pyrex	30°	5.28	0.297
140108	Dolo	Pyrex	45°	5.18	0.298
140109	Dolo (cryst.) ^d	Pyrex	45°	4.89	0.298
140112	Pumice ^e	Serp. ^f	45°	4.93	0.453
140116	Pumice	Dolo ^g	30°	4.54	0.595

^a $d = 6.35$ mm

^bquarter-space dolomite powder

^chalf-space dolomite powder

^dhalf-space crystalline dolomite

^ehalf-space airfall pumice

^fserpentine slug

^gdolomite slug

Figure Captions

Figure 5.1: Contrasting vapor plume time sequences for Pyrex projectiles impacting dolomite powder targets at various incidence angles, imaged in quarter-space geometry. Arrows denote impact point and angle; note that impact point for the 30° case occurred 1.5 cm uprange from frame edge. Plume formation, often simplified into a one-dimensional, isotropic problem, is inherently non-isotropic, even for 90° impacts. For the 90° case, early-time jetting phase vapor flows radially outward, near the target surface, in all directions before expanding vertically upward. This pattern of vapor generation is also seen in two-dimensional hydrodynamical models of impacts. For more oblique angles, the projectile penetration timescale, $t \sim 2r_p(\sin\theta)/v$, increases; the jetting phase is mixed within the early-time transient cavity and then released downrange of the impact point. Upon release, the vapor expands upward and downrange; the downrange momentum imparted to the plume become increasingly important at low angles. While vapor phase temperatures are hottest for vertical impacts, plume mass and luminosity is enhanced at oblique angles. Lower peak shock pressures contribute to lower plume energies and temperatures at shallower angles.

Figure 5.2: Example Boltzmann plot for 90° impact of Pyrex projectile into powdered dolomite (quarter-space). Normalized emission intensities for various calcium emission lines are plotted against the energy of the upper state of the given electron transition, divided by the Boltzmann constant, E_u/k . The inverse slope of the best-fit line corresponds to the excitation temperature, calculated here to be $T \sim 5900$ K.

Figure 5.3: Vapor plume evolution for 45° impact of Pyrex projectile into powdered dolomite (half-space), with emission spectra from four different view spots (denoted by colored circles and arrows). Star marks impact point. Note that delay and dura-

tion of spectral observations differ between regions (acquisition times noted in figure). The principal emitting species are marked on the topmost spectrum (corresponding to red FOV), which is just downrange of the impact point and records the strongest emissions.

Figure 5.4: Vapor plume evolution for 45° impact of Pyrex projectile into solid dolomite (half-space), with emission spectra from four different view spots (denoted by colored circles and arrows). Star marks impact point. Note that delay and duration of spectral observations differ between regions (acquisition times noted in figure). The principal emitting species are marked on the topmost spectrum (corresponding to red FOV), which is just downrange of the impact point and records the strongest emissions.

Figure 5.5: Vapor plume evolution for 30° impact of Pyrex projectile into powdered dolomite (half-space), with emission spectra from four different view spots. Note that delay and duration of spectral observations differ between regions (acquisition times noted in figure). Band heads from many different molecular species are resolved (see rightmost panels of Fig. 5.6 for a closer look at individual band systems). The farthest uprange and downrange view spots (a and d; yellow and blue) are dominated by CO emission. Nearer to the impact point, emissions from CaO and MgO are more prominent.

Figure 5.6: Vapor plume evolution for 30° impact of Pyrex projectile into powdered dolomite (quarter-space), with emission spectra from three different view spots. Note that delay and duration of spectral observations differ between regions (acquisition times noted in figure). Spectra a (purple FOV) and c (blue FOV) contained enough calcium emission lines to derive temperature estimates; the vapor near the leading

edge of the plume was found to cool from 3300 K to 2600 K between these view spots. Observations were separated by $\sim 30\mu\text{s}$ ($t_a=20\text{-}30\ \mu\text{s}$ and $t_c=60\text{-}80\ \mu\text{s}$). The rightmost panels depict subsets of the entire spectral ranges, in which the details of the MgO and CaO band systems are clearer. Spectrum b (red FOV) was taken after the plume front passed; condensation appears to have begun, as evidenced by a larger blackbody component, broad molecular emissions, and Mie scattering. The blackbody temperature is estimated to be 1500 K; it likely arises from incandescent droplets.

Figure 5.7: Vapor plume evolution for 30° impact of dolomite slug into pumice, with emission spectra from five different view spots. Note that delay and duration of spectral observations differ between regions (acquisition times noted in figure). The dolomite projectile-derived vapor shares many molecular features with target-derived vapor (see Fig. 5.3), however, MgO is absent.

Figure 5.8: Comparison between vapor plumes from 30° impacts that have arisen from the projectile (top; crystalline dolomite projectile) and from the target (bottom; powdered dolomite target). While the dolomite projectile plume appears more luminous, the target plume expands downrange with a greater lateral velocity. In contrast, the projectile plume appears to expand at greater vertical velocities. Scale bar is 10 cm.

Figure 5.9: Vaporization of a serpentinite projectile produced unusual molecular features far downrange (a) and uprange (b) from the impact point. Spectra captured in these regions sampled rarified gas, outside of the luminous edge of the vapor plume. Both the uprange and downrange spectra contained a region of dense molecular emission from 580-600 nm. This emission was identified as FeO, a species that has recently

been identified in the Earth's night glow [Evans et al., 2010], attributable to impacting meteoritic material, and has also been observed within the persistent trains of the Leonid meteors [Jenniskens et al., 2000]. Many atomic iron lines are also present; a subset have been identified in spectrum a (blue FOV).

Figures

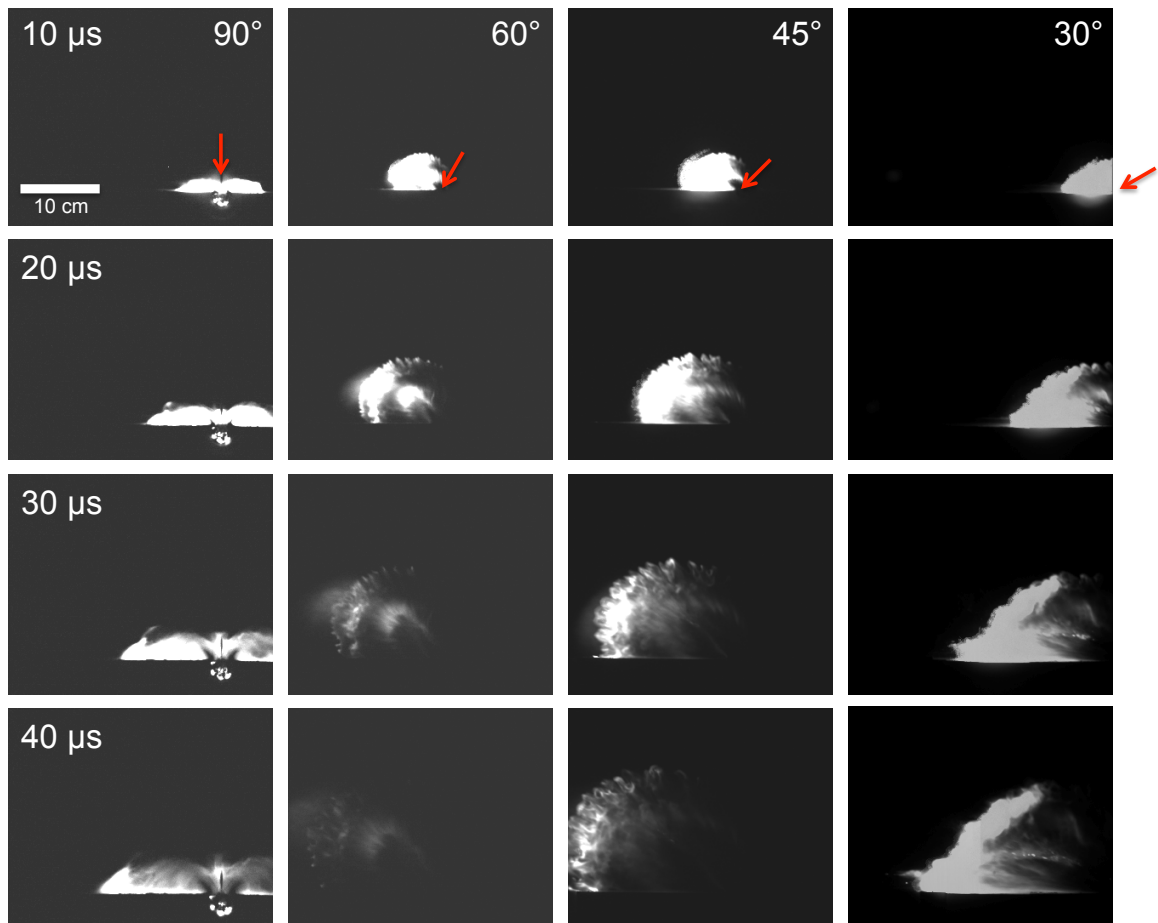


Figure 5.1

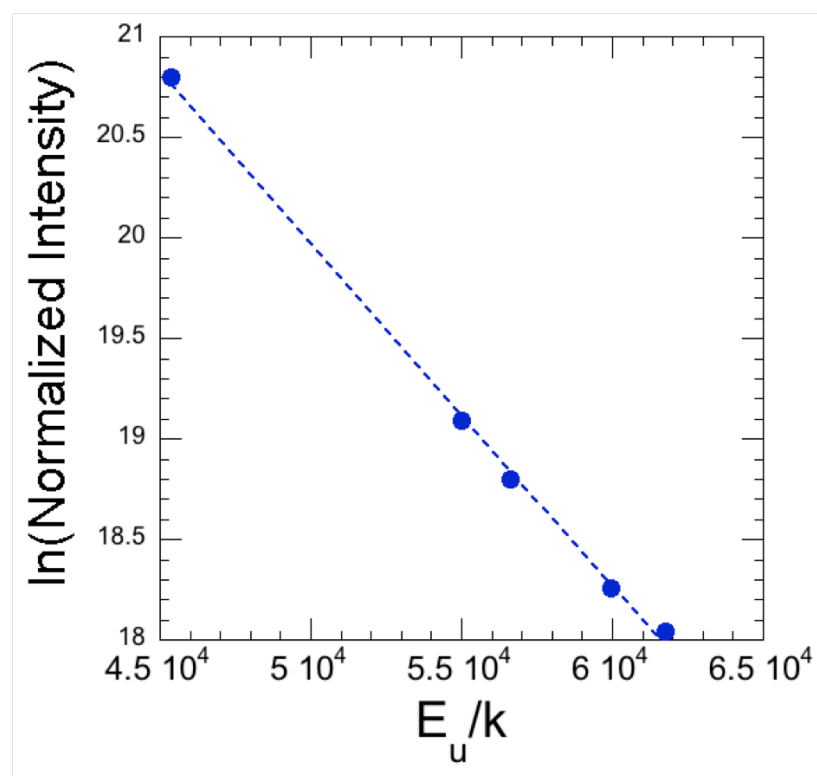


Figure 5.2

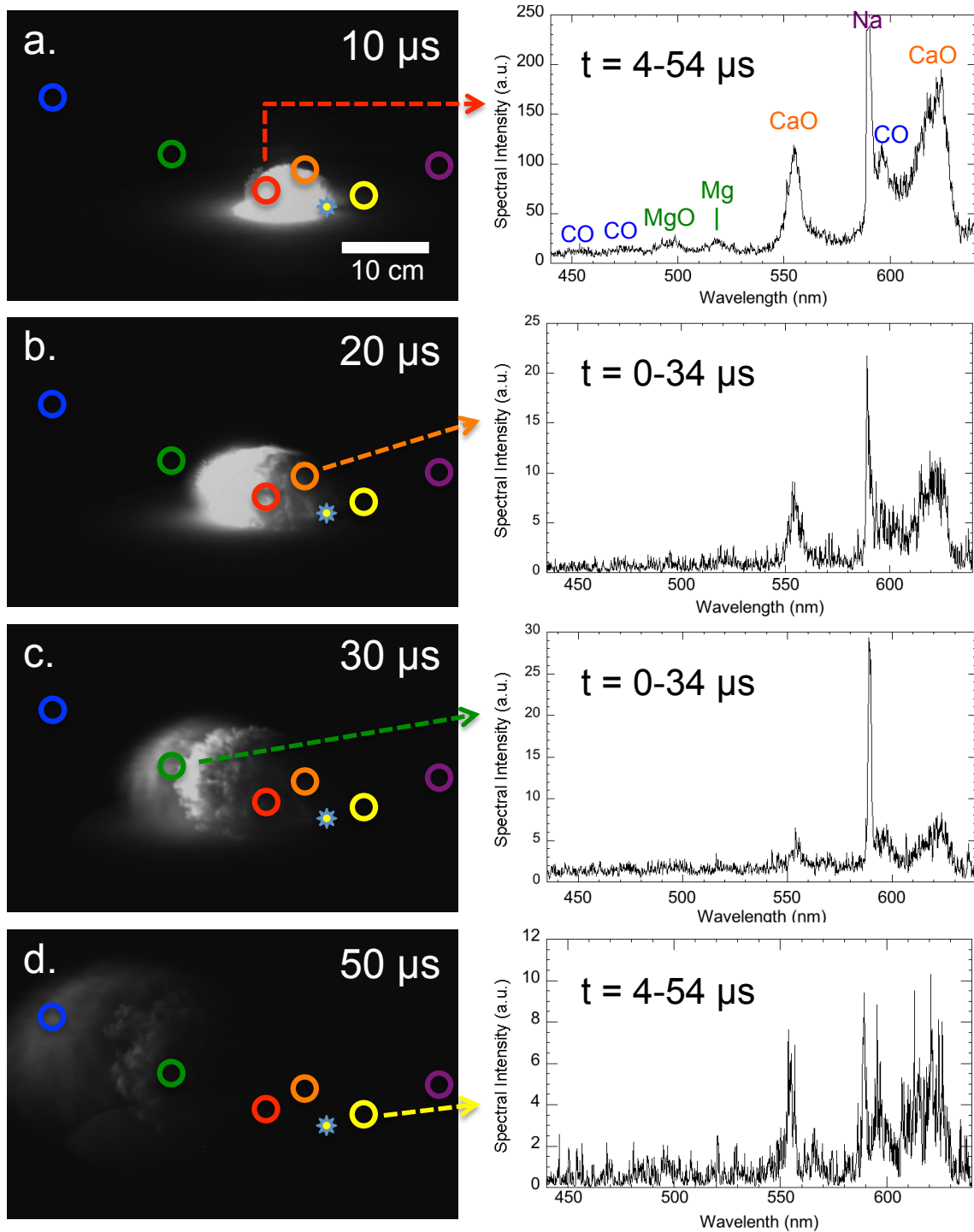


Figure 5.3

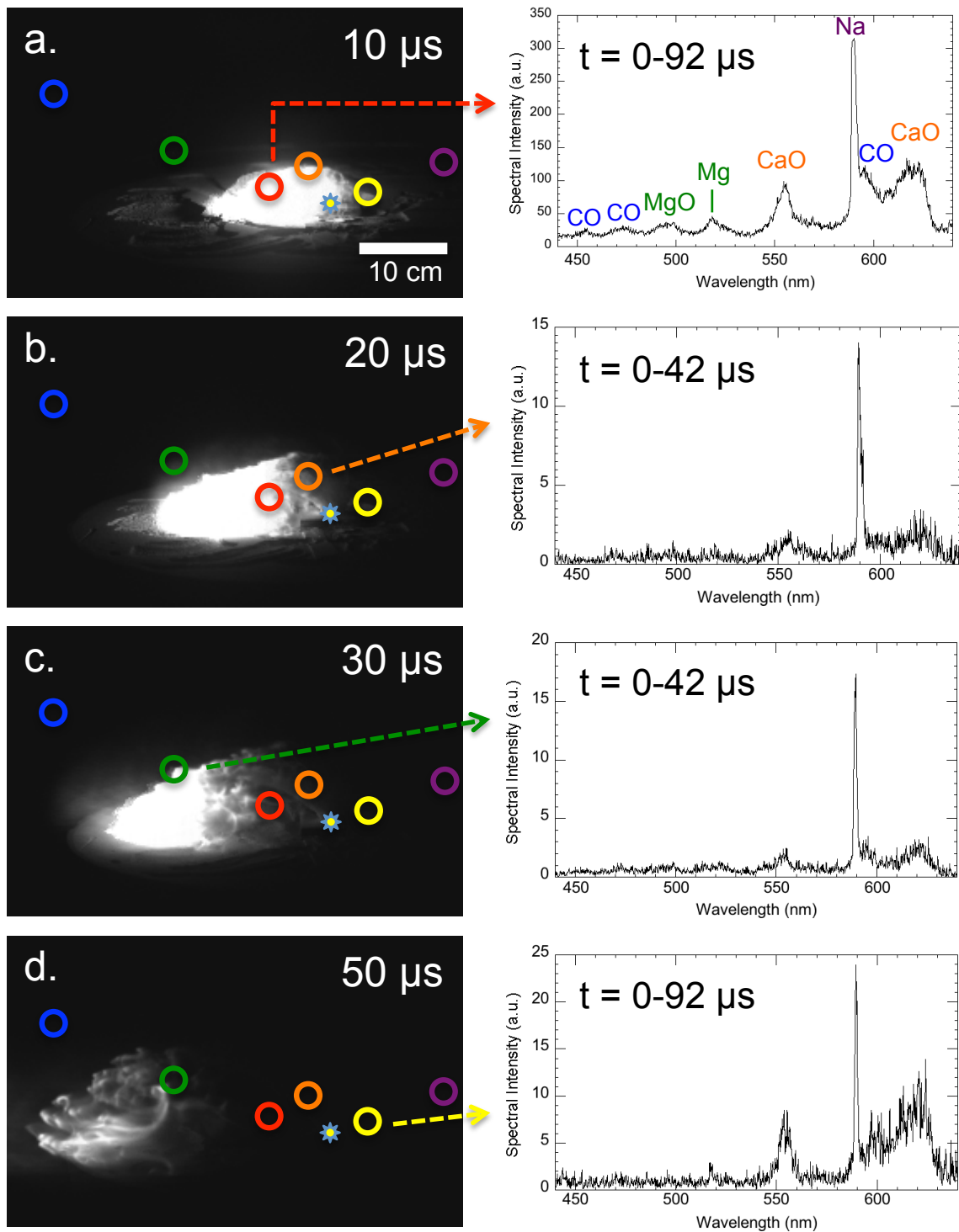


Figure 5.4

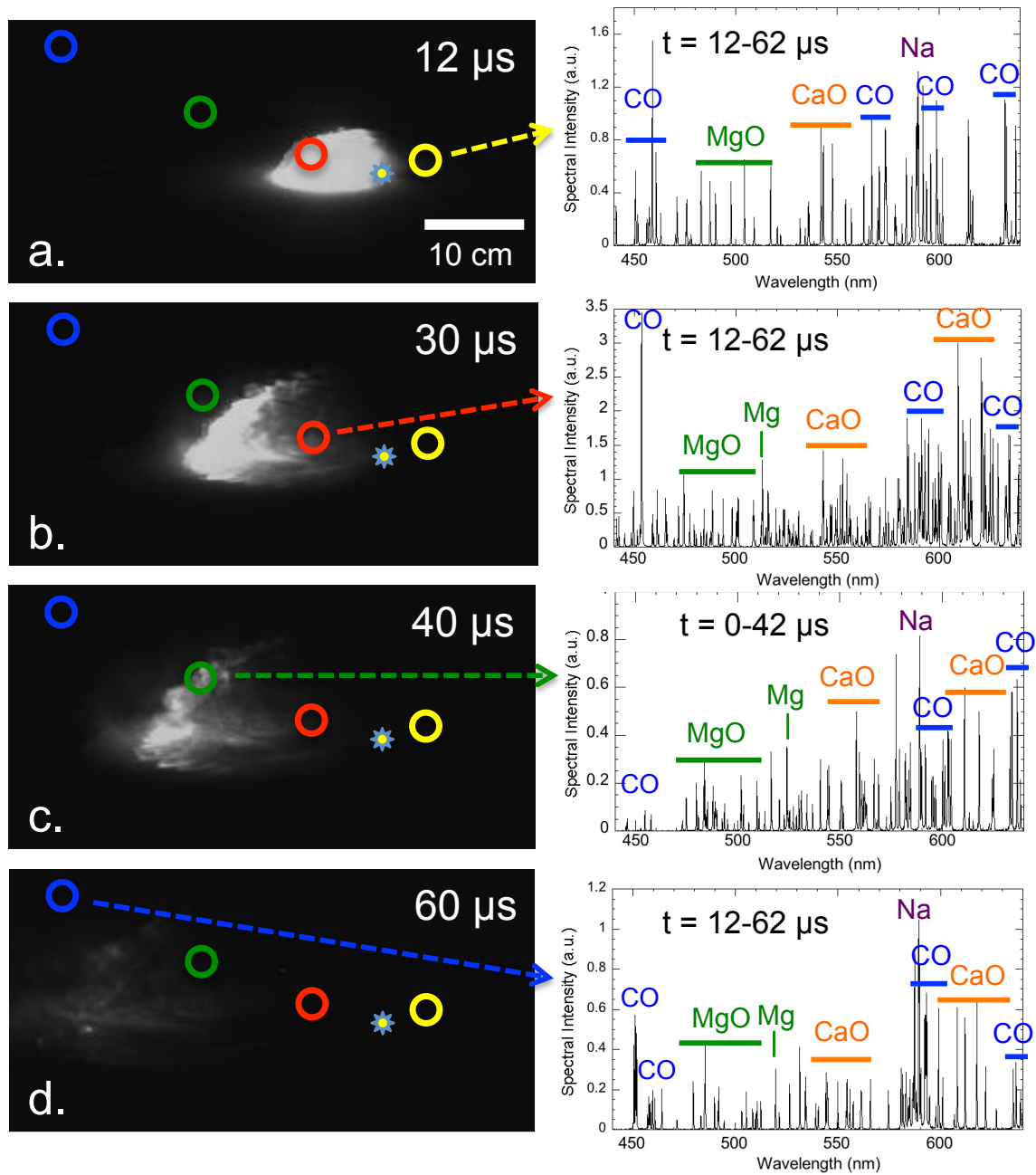


Figure 5.5

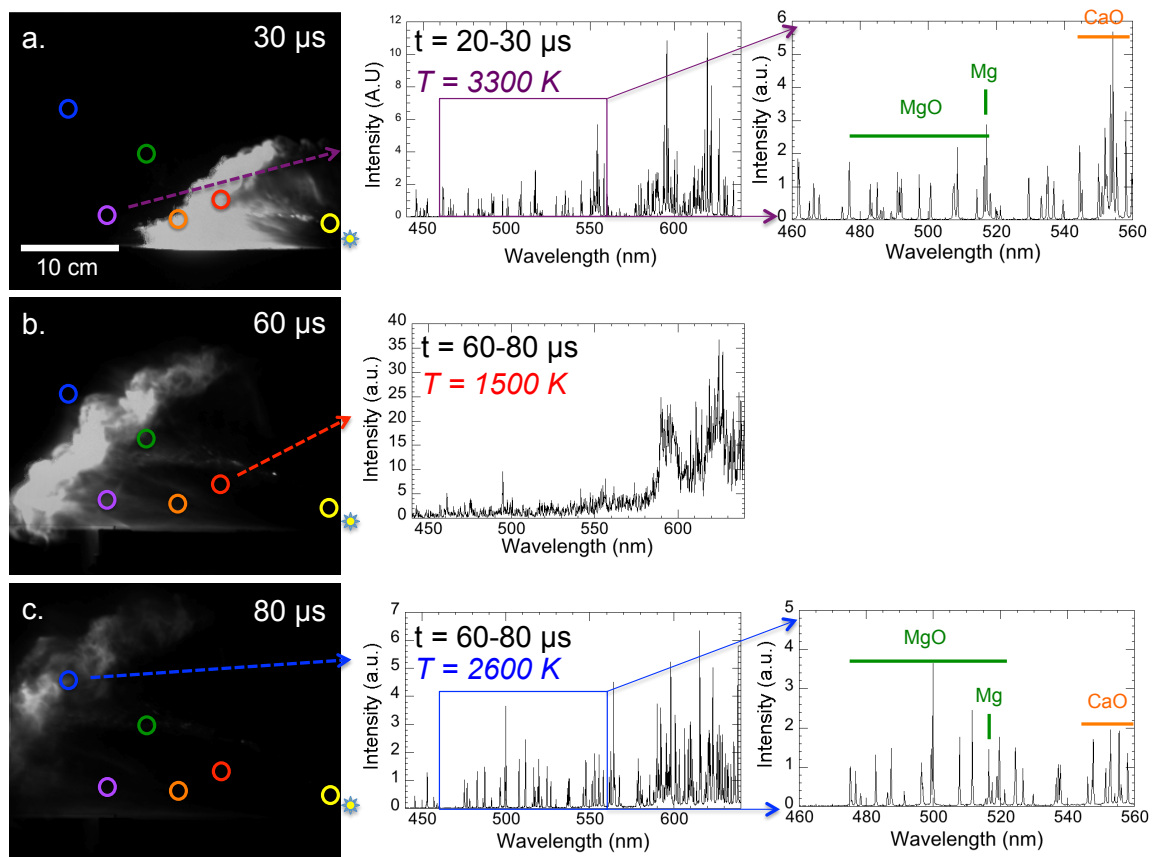


Figure 5.6

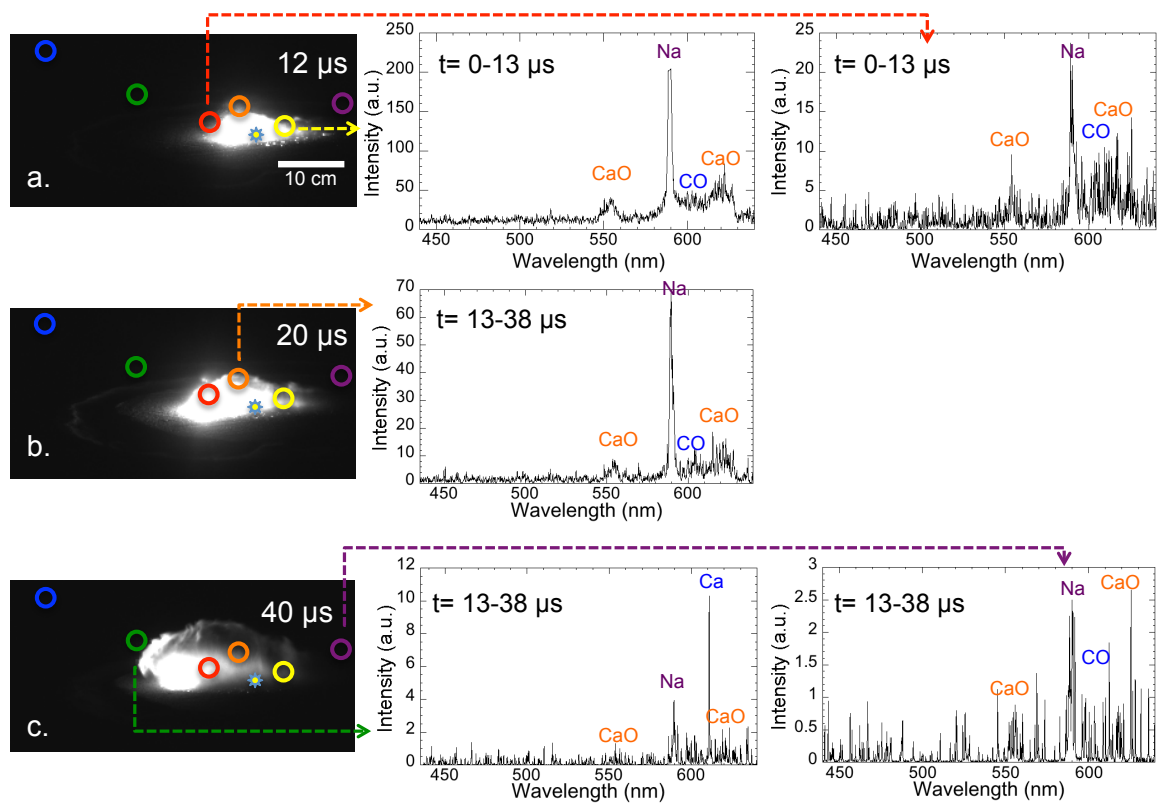


Figure 5.7

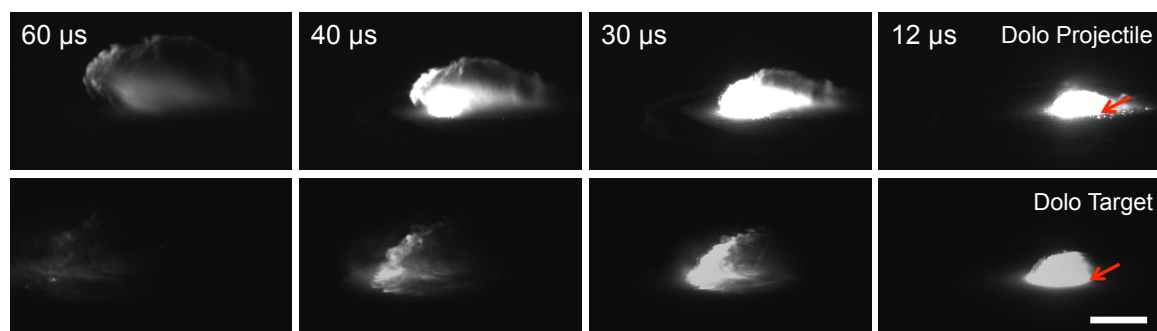


Figure 5.8

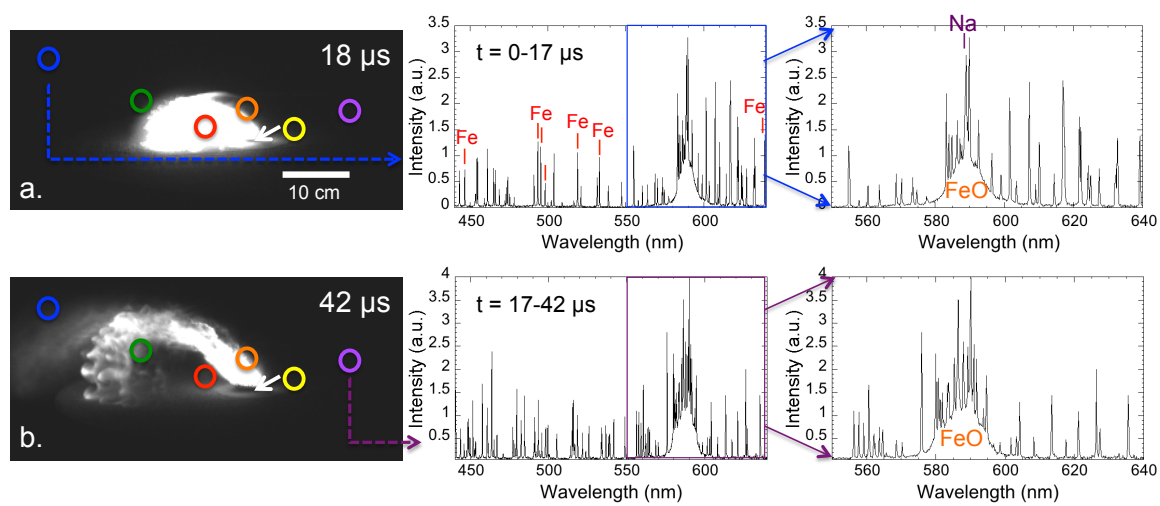


Figure 5.9