Program
Astronomical observations of collisions in exoplanetary systems:
A potential snap-shot of the types of events that affected our own solar system.

Chair: David Kring

8:30 a.m. Su K. Y. L. *
Probing Terrestrial Planet Formation by Witnessing Large Impacts in Extreme Debris Disks [#2014]
I will review the properties of extreme debris disks, and how to use their variable disk emission to study the violent impact events and their aftermaths in young exoplanetary systems during the era of terrestrial planet formation.

9:00 a.m. Jackson A. P. * Su K. Y. L. Dong R. Rieke G. Gaspar A.
Observing Giant, Planet Forming Impacts in Exoplanetary Systems [#2019]
We describe the complex light curves expected for optically thick dust clouds produced by giant impacts in extrasolar systems experiencing ongoing planet formation and discuss the information we can extract about the forming planetary system.

9:20 a.m. Additional Discussion

Analyses of the rock record provide clues about the last phase of accretion and the latest portion of a subsequent period of bombardment.

Chair: Timmons Erickson

9:35 a.m. Walker R. J. * Bermingham K. R.
The Latest Views About Late Accretion [#2015]
Late accretion is a process that undoubtedly left its imprint on the terrestrial planets. The nature and timing of the process can be explored in new ways using geochemical signatures, dynamical modelling and the application of genetic isotopes.

9:55 a.m. Kring D. A. *
Evaluating the Relative Contributions of Asteroids and Comets to the Inner Solar System During the First Billion Years [#2013]
A review of lunar sample analyses and lunar impact crater size frequency distributions suggests asteroids were the dominant impactor on the Moon.

10:15 a.m. Coffee Break
Geochronologic and geological data are used to determine
the absolute and relative timing of impacts during the (lunar) basin-forming epoch.

Chair: Richard Walker

10:30 a.m. Swindle T. D. * Kring D. A.
*Asteroidal Constraints on the Early Bombardment History of the Inner Solar System [#2026]*
Radiometric ages of meteorites from six different groups record bombardment between 4400 and
3000 Ma ago. Most ages are between 3500 Ma and 4000 Ma, and with the exception of the
LL chondrites, ages between 4400 Ma and 4100 Ma are extremely rare.

10:50 a.m. Harrison T. M. * Hodges K. V. Boehnke P. Mercer C. M. Parisi A.
*Problematic Evidence of a Late Heavy Bombardment [#2031]*
Analytical artifacts coupled with the complexity of lunar melt rocks precludes K-Ar step-heating ages to
support the LHB concept and U-Pb accessory phase dating is biased to rocks making up a small fraction
of Moon’s surface requiring a new approach.

11:10 a.m. Boehnke P. * Harrison T. M.
*Illusory Late Heavy Bombardments [#2033]*
First principles modeling of $^{40}\text{Ar}^{39}\text{Ar}$ data does not require an inner solar system bombardment.

11:30 a.m. Zhang B. * Lin Y. Moser D. E. Shieh S. R. Bouvier A.
*Imbrium Zircon Age for Apollo 73155 Serenitatis Impact Melt Breccia: Implications for the Lunar
Bombardment History [#2021]*
Apollo 73155, 69 zircons have vermicular growths, a mean Pb-Pb age of 3928 ± 10 Ma and high U, Th
and Y contents, parallel to those from Imbrium SaU 169 and Apollo 12 high-Th samples. The Serenitatis
zircons point to a provenance of Imbrium basin.

11:50 a.m. Lunch
Geochronologic and geological data are used to determine the absolute and relative timing of impacts during the (lunar) basin-forming epoch.

Chairs: Richard Walker

1:20 p.m. Neumann G. A. * Deutsch A. N. Head J. W.
Ancient Buried Basins in Oceanus Procellarum from the First Billion Years [2030]
Additional buried basins on the lunar nearside addresses an apparent size-frequency difference between the two hemispheres but does not remedy a deficit in large basins relative to the main asteroid belt population statistics.

1:40 p.m. Meyer H. M. * Mahanti P. M. Joshi B. A. Robinson M. S.
Quantifying the Effect of Slope on Age Determination for a Single Geologic Event [2017]
We quantify the effect of topography on the retention of large craters (>1 km) and present the natural range of absolute model ages derived from crater density for a single, large-scale event, the formation of the Orientale basin.

2:00 p.m. Hartmann W. K. *
Collapse of the Terminal Cataclysm/LHB Paradigm: Solving the Problems [2009]
The widely accepted terminal cataclysm or late heavy bombardment (“LHB”) appears to be collapsing. We discuss origina of the idea, reasons for acceptance, reasons for questioning, and a solution to the major resulting problem.

2:20 p.m. Coffee Break

2:35 p.m. Additional Discussion

FIRST BILLION YEARS: DEVELOPING MODELS CONSISTENT WITH THE DATA
2:55 p.m. Agassiz/Fremont

Models of the processes that shaped bombardment have been developed that fit the data and may provide hints of future tests.

Chair: Heather Meyer

2:55 p.m. Marchi S. * Walker R. J. Canup R. M.
The Role of Large Collisions in Forming Early Compositional Heterogeneities on Mars [2011]
In this work we study the potential for large collisions to generate isotopic anomalies in the martian mantle and compare our results to HSE and $^{182}$W isotopic data from the shergottite-nakhlite-chassigny meteorites.

3:15 p.m. Morbidelli A. Nevorny D. Laurenz V. Marchi S. * Rubie D. Elkins-Tanton L. Wieczorek M. Jacobson S.
The Timeline of the Lunar Bombardment — Revisited [2005]
We show that lunar HSEs underestimate the material accreted by the Moon since its formation. In this case, planetesimals leftover from planet formation could explain the lunar crater record, without the need of any later impact spike.
3:35 p.m. Perera V.* Jackson A. P. Elkins-Tanton L. T. Asphaug E. Gabriel T. S. J.
*Cratering and Penetration of the Early Lunar Crust [#2024]*
We conducted a series of iSALE hydrocode simulations to model impacts onto the early Moon. This work will help identify the types of impacts that would have punctured the early crust and help us better understand the thermal evolution of the Moon.

3:55 p.m. Bottke W. F.* Nesvorny D. Roig F. Marchi S. Vokrouhlicky D.
*Evidence for Two Impacting Populations in the Early Bombardment of Mars and the Moon [#2008]*
Asteroids escaping the main belt — or leftover planetesimals — are needed to explain the majority of martian craters, but the ancient lunar farside may be most consistent with cometary impactors. Together, evidence for an early instability?

4:15 p.m. Minton D. A.* Fassett C. I. Hirabayashi M.
*New Insights into Crater Equilibrium Using the Cratered Terrain Evolution Model [#2029]*
We show how equilibrium is affected by the slope of the production function, the size of the area over which the new crater affects pre-existing topography, and the amount of degradation to the pre-existing landscape that each new crater generates.

4:35 p.m. Additional Discussion
Su K. Y. L. Jackson A. P. Dong R. Rieke G. H. Gaspar A.

*Short-Term Disk Flux Modulations due to the Orbital Evolution of Impact Produced Clouds of Dust in NGC2547-ID8* [#2025]

We showed that the short-term disk modulations observed in the ID8 system are due to the orbital evolution of an optically thick cloud produced by two impacts occurred at two different locations and time.

Perera V. Jackson A. P.

*Early Lunar Crust Healed Itself After Impacts Punctured Holes* [#2027]

Hydrocode simulations are appropriate for modeling impacts, yet they are unable to thermally evolve an impact site. Thus we developed a two-dimensional thermal model to quantify the thermal flux over time of a hole in the crust produced by an impact.

Koeppel A. H. D. Black B. A. Marchi S.

*Differentiation in Impact Melt Sheets as a Mechanism to Produce Evolved Magmas on Mars* [#2003]

In this study, we model the petrologic evolution of martian impact melt bodies. We identify signatures of differentiation in impact melt sheets on Mars and compare geochemical trends with measurements of martian igneous materials.

Schmieder M. Kring D. A.

*Found Locally in Arizona: Collisional Remnants of Planetesimal Affected by Impacts During the First Billion Years of Solar System History* [#2006]

The petrologic similarity between the Gold Basin meteorites from NW Arizona and stones recently found in SE Nevada suggests those meteorites are likely part of a single, much larger meteorite strewn field.

Meléndez-López A. Negrón-Mendoza A. Colín-García M. Ramos-Bernal S. Heredia A.

*Study of Glycine Adsorbed in a Clay Under Gamma Radiation in a Comet Cores Simulation* [#2036]

Our aim is to study the possible role of comets as carry agent of organic matter to the primitive Earth. We study the glycine adsorbed in a motmorillonite under gamma radiation at a low temperature, simulating a comet core.

Cruz-Castañeda J. Negrón-Mendoza A. Ramos-Bernal S. Colín García M. Heredia A.

*Stability of D-Ribose-Na\textsuperscript{+}Montmorillonite Suspension System Under Gamma Radiation Fields at pH 7 in a Comet Core Simulation: Implications in Chemical Evolution* [#2035]

Our aim is to study the role of clays as protector agent under $\gamma$ radiation and its implications in chemical evolution. We study the stability of D-Ribose-Clay suspensions in a core comet simulation under different irradiation dose and ratios clay.

Costello E. S. Ghent R. R. Lucey P. G. Tai Udovicic C. J. Mazrouei S. Bottke W. F.

*The Thickness of the Highlands Regolith: Implications for the Early Impact Flux at 1 AU* [#2022]

By coupling remote sensing analysis of the thickness of the highland regolith with a regolith formation model, we explore both the ancient impact flux.
The NASA Lunar Exploration Campaign will provide opportunities to test hypotheses.

**Chair:** Gregory Neumann

8:30 a.m. 
Cohen B. A. Runyon K. D.*
*Constraining Solar System Bombardment Using In Situ Radiometric Dating [#2020]*
A common framework of absolute ages of major planetary events across the solar system, including impact bombardment, volcanism, and windows of habitability, could be developed by including in situ dating on future planetary landers and rovers.

8:50 a.m. 
Runyon K. D. * Moriarty D. Denevi B. W. Greenhagen B. T. Jozwiak L. M. van der Bogert C. H.
*Characterization of Proposed Impact Melt Facies of the Moon’s Crisium Basin [#2007]*
Kipukas around Mare Crisium are likely exposed impact melt from Crisium’s formation. We describe their geomorphology, morphometry, composition, and geologic setting. Radiometrically dating these outcrops would help constrain lunar bombardment history.

9:10 a.m. 
Kring D. A.*
*Testing Bombardment Models with Human-Assisted Robotic Missions and Well-Trained Astronauts on the Lunar Surface [#2023]*
A forward-looking approach is outlined for collecting samples on the lunar surface that will resolve many of the issues involved with bombardment during the first billion years of solar system history.

9:30 a.m. Additional Discussion

**EARLY EARTH RECORD**

9:45 a.m. 
*Can “Granite” Clast from an Apollo 14 Breccia be a Fragment of a Terrestrial Meteorite? [#2032]*
Internally conflicting textural and chemical data obtained for “granite” clast from breccia 14321 indicates possible localised existence of Earth like conditions during magma crystallisation on the Moon or terrestrial origin of the clast.

10:05 a.m. 
Erickson T. M. * Cavosie A. C. Timms N. E. Reddy S. M. Nemchin A. A. Cox M. A. Schmieder M. Kring D. A.
*Where are the Shocked Grains in the Hadean Zircon Record? Insights on the Preservation of Shocked Zircon and Their U-Pb Systematics [#2034]*
The following will review the state of the knowledge on the microstructural development of shock features in zircon, their effects on the U-Pb systematics and the probability of shocked zircon surviving from the Hadean Earth.

10:25 a.m. Coffee Break
10:40 a.m. Zellner N. E. B. *  
*The First Billion Years: Impacts and Life on Earth (and Mars?) [#2016]*  
Multiple impact samples indicate a protracted bombardment in the first billion years of solar system history. However, despite the nature of the impact flux (i.e., cataclysmic, sawtooth, or other), life on Earth was early and widespread.

11:00 a.m. Schulz T. *  Ozdemir S.  Koeberl C.  
*Archean Spherule Layers: Windows into the Early Meteorite Bombardment of the Earth [#2010]*  
The discovery of up to 21 new spherule layers in the Barberton Greenstone Belt marks a great leap forward in deciphering Archean impact rates. We here present geochemical evidence for meteoritic components in these layers.

**LAST BILLION YEARS: IMPACTS IN THE SOLAR SYSTEM**  
11:20 a.m.  

**Chair:** David Minton

*What Happened 800 Million Years Ago on the Moon? [#2028]*  
The $^{39}$Ar/$^{39}$Ar-derived age distribution of “exotic” lunar glass spherules peaks at ~700–900 Ma, coinciding with a possible age of Copernicus Crater. We propose that this coincidence may be responsible for the excess of exotic glass spherules.

**BOMBARDMENT AND IMPLICATIONS FOR HABITABILITY**  
11:40 p.m.  

**Chair:** Nicolle Zellner

11:40 p.m. Boston P. *  
*Lowering the Boom on Life: From Origins to a Biosphere’s Future*

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**Tuesday, October 2, 2018**  
**FIRST BILLION YEARS: FINAL DISCUSSION**  
12:10 p.m.  

**Chair:** Kate Su  
**Chair:** David Kring

12:10 p.m. Final Conference Discussion
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B. Zhang, Y. Lin, D. E. Moser, S. R. Shieh, and A. Bouvier ................................................................. 2021
Introduction: Meyer and coauthors obtained U-Pb data for two zircon grains collected during the extraction of a felsic clast from lunar breccia sample 14321 and pointed out that these grains reside in the fragments that also contain grains of quartz and K-feldspar showing the same texture as that in the fused portions of the clast [1]. This led to the conclusion that two zircon grains originated from the partially melted “granite” clast described in section 14321,1027 by Warren and coauthors [2]. The clast has an estimated mass of 1.8 g and consists of 60% low-Ba alkali feldspar and 40% quartz with minor Fe-rich olivine and traces of ferrohedenbergite, ilmenite, and Fe-Ni-metal [2]. However, the clast is brecciated and also contains close to 30% of crystalline impact melt [2]. Unbrecciated areas are composed of quartz and alkali feldspar occurring as large (1.8 x 0.15 mm) inter-grown crystals [2]. Oxy CALCIOBETAfITE is also found as intergrowths with some of the primary feldspar grains [3]. The crystalline impact melt consists of intergrowths of silica and feldspar less than a few micrometers in size with a significant proportion of pyroxene and small blebs of Fe-metal (Figure 1).

Textural and chemical information available for this clast appears to be contradictory. Metallic Fe is indicative of low /O2 conditions and is ubiquitous in lunar samples. Bulk clast concentrations of Zn, Ge, Ga, Au, Ba, Ta, and REE, a single pyroxene analysis indicating relatively low Mn/Fe [2], and the highly radiogenic Pb isotope compositions of K-feldspar analyzed in one of the thin sections [4] also support the lunar origin of the clast. However, the presence of oxy CALCIOBETAfITE in the same felsite suggests this clast was formed under conditions that are not typically associated with the Moon [3]. It was found to contain significant amounts of Fe3+ and W6+, interpreted as an indication of its formation under relatively oxidizing conditions [3]. Additionally, a more recent investigation of the sample indicated that the REE pattern of oxy CALCIOBETAfITE shows a split to four consecutive curved segments that were referred to as tetrads and interpreted to reflect formation in the presence of water or F-rich fluid [5], which is also not fully consistent with a lunar environment that is considered to be relatively dry and reducing compared to terrestrial magmas.

These results in combination with the new trace elements data for zircon (REE and Ti) and quartz (Ti) from the “granite” clast, highlight controversy related to the interpretation of the origin of the clast.

Results: The REE analyses of two zircon grains in the thin section 14321,1613 show pronounced positive Ce/Ce* anomalies of 7.6 ± 2.3 and 17.5 ± 8.2 (2σ), (Figure 2A). Ti temperatures determined for two grains ranging from 763 ± 24 to 840 ± 28 °C (2σ) are lowest ever recorded among nearly 100 analysed lunar zircon grains (Figure 2B). The Hf, U, and Th concentrations in these two grains, range between 1.29-1.85 wt%, 298-986 μg/g, and 153-542 μg/g, respectively, and are the highest of any analyzed lunar zircon reported so far. In contrast, the total REE content is among the lowest in the lunar zircon population (Figure 2A).

Titanium concentrations of 214 ± 5 μg/g were determined in quartz from the clast in thin sections 14321, 1047 and 1029. Combining this Ti in quartz estimates with values obtained for Ti concentrations in the zircon grains allows calculation of P as well as T of clast formation, assuming that both zircon and quartz crystallised from the same felsic melt. An average crystallization P of 7.4 ± 1.2 kbar (2σ) is determined from the obtained data, which on the Moon indicates a formation depth of 135 ± 22 km (2σ).

Discussion: “Granite” clast shows clear evidence of textural and chemical dichotomy. Assemblage consisting of quartz, K-feldspar, zircon and oxy CALCIOBETAfITE appear to form as a result of magmatic crystallization, while presence of matrix in the samples indicates impact induced partial remelting of the clast. Additionally, the presence of Fe-metal and bulk concentrations of trace elements [2], in particular low concentrations of volatile metals in the clast, indicate an origin consistent
with crystallization conditions typically ascribed to lunar melts. A lunar origin for the clast is further supported by the highly radiogenic Pb isotopic composition of K-feldspar [4]. However, the chemical characteristics shown by the zircon grains are more compatible with the crystallization of felsic melts on the Earth. Oxycalciobetafite appears to support the zircon data, indicating a relatively oxidized, low-T, incompatible element and possibly water or F- rich melt. Another profound contradiction is evident when comparing the impact modelling results made as a part of our investigation of “granite’ clast, which limit the depth of excavation of the sample 14321 in the Imbrium ejecta to 30-70 km below the lunar surface, and P-T estimates obtained using Ti in quartz and zircon thermobarometry. The later suggests that formation depth of the felsite clast was 135 ± 22 km, significantly deeper than that indicated by the impact model.

This dichotomy explicitly implies a multi-stage petrogenesis, where part of the mineral assemblage represents the primary magmatic characteristics of the rock while other was acquired during secondary modification. Zircon is highly stable with the ability to resist chemical, thermal, and mechanical modifications, hence the zircon is anticipated to best reflect primary magmatic conditions of the crystallizing felsic melt. Quartz is almost as chemically and mechanically resistant as zircon and could be expected to survive the changes that accompanied the secondary processes that have affected the rock. Texturally, these minerals appear to constitute the unbrecciated part of the clast, also supporting preservation of their primary characteristics. In contrast, the Fe-metal and pyroxene that has a lunar Mn/Fe ratio appears to be confined to the crystalline shock melt and likely formed by post-crystallization heating during incorporation of the felsite into the breccia. K-feldspar shows a highly radiogenic Pb isotope composition, which is likely to have been acquired as a secondary component resulting from relatively low-temperature diffusion of mobile, highly radiogenic Pb during the breccia formation.

Previous studies have proposed that the felsite clast in 14321 formed by extensive fractional crystallization, which concentrated highly charged cations in the residual liquid, also resulting in an increasingly oxidized melt, perhaps due to the presence of a fluid [3, 5]. While these conditions do not appear to be common on the Moon, the data obtained for the felsite clast suggest that they could have existed locally within the lunar crust, possibly resulting from slow cooling and fractionation of an initially basaltic melt within the lower crust or near the crust-mantle boundary. If that is the case, after crystallization at a depth of 30-70 km, the felsite was excavated during the Imbrium impact and the characteristics indicative of more reducing conditions were introduced to the clast during incorporation of the clast into the host breccia at ca. 3.9 Ga. The only observation that cannot be explained directly by lunar origin of the sample is Quartz/Zircon Ti-based pressure estimate for the felsite formation that places it at 135 ± 22 km. This depth estimate is difficult to accept from the perspective of both crystallisation of the felsite, as it places it significantly below mantle-crust boundary where felsite likely cannot crystallize, and the predicted depth of excavation even when large, basin forming impacts are considered. An alternative interpretation that can reconcile all observations including the high P of crystallization, is a terrestrial origin for the felsite clast. On Earth, the depth corresponding to 7.4 ± 1.2 kbar is 20 ± 3 km and places the sample comfortably within the middle crust where it could have formed under oxidizing, low-T, fluid rich conditions common for terrestrial magmas. In this scenario, zircon, oxycalciobetafite and quartz would have preserved the record of primary crystallization. The subsequent history of the felsite would include excavation by a large impact from the depth of 20 ± 3 km and its delivery to the Moon as a terrestrial meteorite. This would have been followed by its incorporation into Imbrium impact ejecta and ultimately the 14321 breccia, which would have also introduced all of the secondary ‘lunar’ features.


![Figure 2: Trace elements in zircon from “granite” clast: A-REE (blue) zircon grains from [6 and 7]; red – zircon from saw cuts of the clast from breccia 14321); B- Ce/Ce* vs. crystallization T (lunar data from [6 and 7], terrestrial data from [8]).](image-url)
ILLUSORY LATE HEAVY BOMBARDMENTS. P. Boehnke and T. M. Harrison, 1,2,3
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Introduction: The Late Heavy Bombardment (LHB), a hypothesized impact spike at ~3.9 Ga, is one of the major scientific concepts to emerge from Apollo-era lunar exploration. While it was originally hypothesized based on analyses of Pb isotopes and Rb-Sr ages of a limited set of Apollo samples [1], a significant portion of the evidence now marshalled comes from histograms of 40Ar/39Ar “plateau” ages [2]. Despite the lack of erosion and plate tectonics, the lunar crust does not retain a perfect impact record due to protracted crust formation, lunar volcanism, and overprinting from subsequent impact events. Indeed, virtually all Apollo-era samples show 40Ar/39Ar age spectrum disturbances due to later re-heating events [3]. This provides evidence that partial 40Ar resetting is a significant feature of lunar 40Ar/39Ar analyses which could bias interpretation of bombardment histories due to “plateau” ages being misleadingly young. In order to examine the effects of partial resetting on the inference of bombardment histories from “plateau” ages, we combine chronologic information derived from the early heating steps of each 40Ar/39Ar analysis, as this represents a good approximation of the timing of the last reheating event, with a first-principles physical model of 40Ar* diffusion in Apollo samples. We use this modeling framework and data compilation to examine the uniqueness of inverting “plateau” age_histograms from synthetic impact histories as implemented in two models with differing assumptions about 40Ar loss during impacts.

Methods: Specifically, we compiled the initial step-heating ages for 267 analyzed Apollo samples and interpret them to represent the timing of the last re-heating event (termed Last Heating Ages; LHA). We utilize LHAs as our primary constraints on the shape of the impact history because they reveal the age of the least retentive portion of the sample and are therefore unlikely to be partially reset. Our diffusion model contains 1,000 synthetic ‘samples’ with initial ages drawn from the distribution of lunar 207Pb/206Pb zircon ages and trials random impact histories to recreate a distribution of LHAs. Since we have no prior information for the distribution of fractional loss of 40Ar in an impact, we utilize two separate models to examine the effect on the inferred impact history of a range of assumptions. In our first model run we utilize a uniform distribution of fractional loss and compile a histogram of “plateau” ages. In the second model we explicitly constrain it to create LHB at either 3.9 Ga or 4.1 Ga and instead fit a probability distribution of fractional loss for 40Ar in each impact event.

Results and Discussion: We find that both our models produce good fits to the LHA distribution. Model 1 is only constrained to the LHAs and produces a “plateau” histogram with a broad bombardment episode between 3 and 4 Ga. While Model 1 does not reproduce the LHB in detail, the fact that it produces an impact spike shows that “plateau” histograms are likely prone to suggesting late bombardments. Our model 2 was able to both fit the LHA distribution and create an apparent LHB at either 3.9 Ga or 4.1 Ga. Our results show that “plateau” histograms tend to yield age peaks, even in those cases where the input impact history did not contain such a spike. That is, monotonically declining impact histories yield apparent episodes that could be misinterpreted as an LHB-type events. Since not only Apollo samples show apparent impact spikes, but the H-chondrites and HED meteorites do as well [4], our findings have broader implications to impact histories for meteorite parent bodies.

The above interpretation is in fact a best case scenario as it is predicated on the assumption that meaningful thermochronological information can be obtained from 40Ar/39Ar age spectra of lunar samples. This is a highly optimistic assumption, however, as several analysis artifacts could entirely obscure the original signal.

We conclude that the assignment of apparent “plateau” ages bears an undesirably high degree of subjectivity. When compounded by inappropriately simplistic interpretations of histograms constructed from such “plateau” ages, impact spikes that are more apparent than real can emerge [5].

Motivating Problem. The sources of early bombardment for Mars and the Moon have long been debated [e.g., 1]. It is challenging to resolve this issue because (i) dynamical modelers are still struggling to understand the endgame of planet formation, which likely involved giant planet migration, and (ii) ancient crater records on Mars and the Moon, mostly examined using diameter $20 < D_{\text{crat}} < 150$ km craters, are not diagnostic enough by themselves to rule out different possibilities. Recently, however, new data for large $D_{\text{crat}} > 150$ km craters has become available via NASA missions [2-4]. Here we use this information, new numerical simulations of giant planet migration [5], and new crater scaling laws [6] to compare and contrast their early bombardment signatures.

Early Bombardment Sources. Plausible sources for early inner solar system bombardment are leftover planetesimals from terrestrial planet formation [e.g., 7] and the depletion of small body reservoirs (e.g., primordial Kuiper and asteroid belts) by giant planet migration as discussed in the Nice model [e.g., 5, 8]. The former would have been recorded as soon as planetary surfaces were stable. The latter could have hit early and/or late, depending on when the giant planets experienced an instability that reconfigured their orbits. The shape of the main belt size-frequency distribution (SFD) is shown in Fig. 1 [9]. In our model runs, most depletion comes from the inner main belt ($a < 2.5$ AU); the central main belt is less depleted and the outer main belt is left largely in place. From a probability standpoint, most Mars/Moon impactors come from the inner main belt, where the impact ratio is $\sim 10:1$, respectively, and $\sim 0.5\%$ strike Mars. Overall, the primordial main belt loses roughly a main belt’s mass (i.e., $\sim 8000 \times 2, \div 2$ bodies with $D_{\text{ast}} > 10$ km). The flux and SFD of leftover planetesimals is unknown.

Mars Bombardment. Using main belt numbers, we predict Mars was hit by $\sim 40 \times 2, \div 2$ $D_{\text{ast}} > 10$ km bodies, with the impacting SFD having the same shape as the inner/central main belt SFD from Fig. 1. To extend these SFDs to $D_{\text{ast}} < 3$ km, we assumed the main belt SFD followed the shape of the NEO SFD [10]; this is reasonable because most NEOs come from the main belt via the Yarkovsky effect [10] and the NEO SFD matches Vesta’s crater SFDs [6].

For Mars’s bombardment constraints, we first looked at $\sim 100 D_{\text{crat}} > 150$ km craters located predominately on its ancient southern highlands (Fig. 2) [2, 3]. They cover 40-50% of Mars and were identified after an extensive geologic mapping effort. Buried/equivocal structures were excluded. Also plotted are $20 < D_{\text{crat}} < 150$ km craters found near Hellas basin [11]. Hellas is perhaps the oldest largest basin to form after Mars’s surface was reset by the Borealis basin-forming event [12].

Intriguingly, the shapes of the main belt and Mars crater SFDs are the same (Figs. 1-2). Moreover, if we assume the crater scaling law function, $f$, is a simple ratio between crater and projectile diameters (i.e., $f = D_{\text{crat}} / D_{\text{ast}}$), and that $f \sim 24-30$, not only do the curves match one another but they also yield the observed...
Accordingly, main belt depletion produces <17% of the net ancient lunar bombardment. These impacts may explain the Moon’s younger basins/craters [15] (Fig. 3, green curve). Alternatively, if leftovers dominated lunar bombardment, the differences in Figs. 2 and 3 are more difficult to explain.

**A Comet Signature on the Ancient Moon?** The SFD in Fig. 3 looks similar to KBO/Trojan SFDs (i.e., cumulative slope $q \sim -2$ for $D_{\text{crat}} < 100$ km) [e.g., 16]. Their potential kinship may mean numerous comets struck the ancient Moon. We can test this idea using the Nice model [5]. It predicts $2.6 \times 10^5$ of the primordial disk struck the Moon for ~20 Myr after the giant planet instability. The number of $D_{\text{crat}} > 40$ km craters in Fig. 3 is ~3000, so a scaling law of $f \sim 24$ translates this number into $\sim 10^{11}$ comets with $D_{\text{crat}} > 1.67$ km. We can reproduce this value, provided there was once $5 \times 10^7$ comet/bodies in the disk (as predicted).

**Implications.** These results make more sense if the Nice model takes place close in time to the Moon-forming event (>4.5 Ga). Ancient Moon/Mars would have both been struck by comets, but Martian comet signatures would likely have been erased by the Borealis basin formation event ~4.43-4.48 Ga, with the timing inferred from Martian zircons [e.g., 3, 17]. These timescales are consistent with the ~20 Myr timespan of comet bombardment found in the Nice model. It also makes sense that the Ceres-sized Borealis impactor hit Mars during the planet formation era [3].

If the lunar farside is ancient enough to have a comet signature, the major mafic event at ~4.35 Ga inferred from lunar samples is likely describing an event within the PVT model, which can reproduce this value, provided there was once $5 \times 10^7$ bodies in the disk (as predicted).

**Number of D > 300 craters (Fig. 2) from $D_{\text{crat}} > 10$ km impactors (i.e., ~40 ×2, ~2). Our scaling law is based on empirical fits between the NEO SFD and terrestrial planet crater sizes formed over the last 3 Ga [6]. Accordingly, main belt depletion can plausibly reproduce Mars’s post-Borealis bombardment record.

If leftovers are important, possibly because many main belt impactors strike Mars prior to Borealis, they must match the numbers and SFDs discussed above.

**Lunar Bombardment.** The oldest lunar terrains (Pre-Nectarian; PN) are mostly located on the Moon’s farside [13-14]. Using the record of $D_{\text{crat}} > 150$ km craters identified in GRAIL data [3], we computed which regions that had the highest crater spatial densities. Collectively, these regions cover 70% of the farside and compare favorably to the PN terrains identified by [13-14]. The $D_{\text{crat}} > 150$ km craters located there are plotted in Fig. 3, with their numbers scaled to the entire lunar surface. Regions excluded include terrains near large young basins like Orientale, Hertzprung, and Moscoviene. A set of smaller PN craters ($20 < D_{\text{crat}} < 150$ km) are also plotted [15].

The two crater SFDs in Fig. 3 (blue curves) fit together and yield a cumulative slope of $q = -2.07$ for $40 < D_{\text{crat}} < 600$ km. This value is much shallower than the $q \sim -3$ slopes seen for $D_{\text{crat}} < 20$-30 km in Fig. 1. Mars shows no indication that it was hit by such crater SFDs in Fig. 3, and we argue the lunar craters plotted are not in saturation. We conclude that the lunar farside provides evidence for a different and likely older impactor population than that found on Mars.

**Interpretation.** If we assume main belt impactors produced Mars’s crater record (Fig. 2), a ~10:1 impact ratio for Mars/Moon yields ~30 $D_{\text{crat}} > 150$ km craters on the Moon younger than Borealis. This compares to ~180 predicted from the oldest PN terrains (Fig. 3). Accordingly, main belt depletion produces <17% (30/180) of the net ancient lunar bombardment. These impacts may explain the Moon’s younger basins/craters [15] (Fig. 3, green curve). Alternatively, if leftovers dominated lunar bombardment, the differences in Figs. 2 and 3 are more difficult to explain.

**REFERENCES:**

Introduction: The 92 km diameter crater Occator in Ceres’ northern hemisphere is perhaps the most enigmatic feature on the dwarf planet, and hosts the highest albedo features on the body [1]. These bright spots (named faculae) within Occator are dominantly composed of sodium rich carbonates [2], and exist largely within two regions of the crater. Cerealia facula is a dome of bright carbonate deposits that dominate the center of Occator [1-2]. This dome sits within a central pit similar in morphology and scale to those seen within craters on icy satellites [3]. Venalia faculae are a set of bright deposits within the northeast quadrant of Occator [1]. While spread out over a greater area and less voluminous, they have a similar composition and albedo as Cerealia facula [2].

The mechanism behind the formation of Ceres’ faculae, and their genetic relationship to the impact that formed Occator, is currently debated. When Occator formed, a significant amount of heat was deposited within the subsurface [4]. Because Ceres likely contains a significant fraction of water ice [5-6], this hot material would be capable of driving a temporary hydrothermal system, particularly in the center of the crater where the hottest material would have been concentrated [4]. Hydrothermal brines, upon interaction with the surface, could have sublimated, forming ballistically emplaced deposits of salts and carbonates within the center of the crater [7-8]. Alternatively, fractures induced by the Occator forming impact could have allowed for the transit to the surface of cryo-lavas from a pre-existing subsurface fluid reservoir unrelated to impact heat, both erupting onto the surface as well as forming an uplifted dome [9].

Although Occator a young, fresh crater [10-11], its morphology and composition still have important implications for ancient bombardment on the body. The presence of faculae within Occator strongly suggests that the crater’s morphology was modified post-impact by hydrothermal and/or cryovolcanic processes. These processes should have affected all craters on Ceres, and must be accounted for in order to understand the effects of the dwarf planet’s early bombardment history.

Detailed models of both Occator’s formation and evolution are required in order to understand the unique structure of the crater and what it implies about the anatomy and history of Ceres’s interior. In order to address this holistically, we directly couple models of Occator’s formation, a process dominated by impact shock and high strain rate flow, to hydrothermal simulations, which reproduce the long term evolution of fluid flow beneath the hot, fresh crater.

Methodology: We simulate the formation of Occator crater using the shock hydrodynamics code iSALE [12-14]. Our crater is formed when a main-belt like impactor traveling at typical Cerean impact velocities collides with a surface composed of a mixture of H2O ice and serpentine [4]. We implement a simple model that described how permeability within the subsurface develops as a function of porosity [15]. The results of our iSALE simulations describe the distribution of temperature, porosity, permeability, and water content at the end of the crater formation process and at the onset of hydrothermal circulation. The output from these simulations is then directly mapped into the hydrothermal evolution code HYDROTHERM, which describes water and heat transport in porous geologic media [16]. HYDROTHERM, which was developed to account for terrestrial hydrothermal systems, has been modified to account for the enthalpy of freezing, making it capable of modeling low temperature conditions such as those found at the surface of Ceres. In combination, these simulations allow for a detailed examination of both the formation and evolution of hydrothermal systems within the Cerean subsurface.

Results: The impact that formed Occator heated significant volumes of material above the eutectic of subsurface brines [7], allowing for effective fluid circulation immediately after crater formation. The total volume of hydrothermally viable material is dependent on the water-serpentinite ratio of the pre-impact surface, with higher water contents resulting in larger volumes of hot fluids. Hydrothermal circulation within the subsurface of the crater is largely constrained to a hot ‘plug’ in the center of the crater. Inward flow of fluids from depth leads to large amounts of effusion at the center of the crater. The hottest regions of the subsurface are largely composed of impactor material, which drives localized, relatively rapid fluid circulation. The lifetime of the system varies considerably depending on permeability, but has a maximum duration of a few hundred thousand years. Pressure, temperature, and water/rock ratio time-series for parcels of fluid within our simulations can be used to inform models of the chemical evolution of the brines that formed the faculae observed on Ceres.
Formation of Faculae and Occator’s Central Pit: Recent observations by the Dawn spacecraft’s 2nd extended mission can provide new insights into the crater’s formation, evolution, and the origin of Ceres’ faculae. Very high resolution images of Cerealia facula (Figure 1, [17]) seem to show that the bright carbonate deposits within the crater lie both on the rim of as well as in a dome at the bottom of the crater’s central pit. This may imply that the central pit of the crater formed either contemporaneously with or after the deposition of the faculae.

Central pits within craters on icy satellites and Mars are thought to form when H$_2$O impact melt, which is concentrated in a plug within the center of the crater, drains downward through impact induced fractures into the subsurface [18]. However, the presence of faculae deposits and our simulations of hydrothermal circulation imply that impact induced heating drove hydrothermal effusion of brines upwards onto the surface. Our results suggest that subsidence driven by the loss of material via hydrothermally driven brine effusion and sublimation lead to the collapse of Occator’s central pit. This brine effusion also can explain the formation of the faculae as evaporite deposits left behind after sublimation of the H$_2$O component of the brines. Under this model, Cerealia facula initially began to form before the collapse of the central pit, producing the portions of the bright deposits that currently exist on the rim. As the pit began to subside, much of the bright deposited material was subsumed downwards. Continued effusion of brines lead to the deposition and uplift of a central dome within the pit.

Our iSALE models imply that the hottest regions within the post-impact surface are composed of impactor material, consistent with previous models of impact heating [19]. Because our simulations are two dimensional and axisymmetric, most of this impactor material is concentrated within the center of our final crater’s subsurface. However, during oblique impacts, a significant portion of the impactor can travel into the subsurface in the downrange direction, potentially outside of the region of hydrothermally viable material which in our simulations is largely constrained to the crater center. Although not explicitly modeled, this off-axis hot material could potentially drive a secondary hydrothermal system with a shorter lifetime than the hydrothermal system within the center of the crater. Such a system could potentially form a secondary deposit of bright material on the crater floor such as Vinalia faculae. If this is the case, the location of Vinalia faculae implies that Occator was formed during an oblique collision with a projec-

tile impinging on the Cerean surface from the west or southwest.


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Figure 1: View of Occator crater with inset from Dawn’s 2nd extended mission showing bright material on the rim of the central pit [17].
Introduction: Our knowledge of absolute surface ages on other bodies, including Mars, Mercury, asteroids, and outer planet satellites, relies primarily on the crater calibration record for the Moon. Surface ages record two fundamental planetary processes: differentiation and volcanism via internal heating, and impact cratering. Despite the ubiquity of these processes, little headway has been made in creating a common framework of absolute ages for these events across the Solar System, including the Late Heavy Bombardment, planetary volcanism, and windows of habitable environments on the Earth, Mars, and beyond.

The leading, but contentious, model for Solar System impact history includes a pronounced increase in large impact events between 4.1 and 3.9 Ga [1-4]. This cataclysm would have bombarded an Earth and Mars that had oceans, continents, and atmospheres, and may have influenced the course of biologic evolution. Dynamical models to explain such a phenomenon are invoked to explain the arrangement of exoplanets around other stars. Evidence for a cataclysm, and constraints on the rest of the solar system impact flux, come largely from geochronology of samples derived from multiple samples from the Moon and asteroids.

The older end of the flux curve is bounded by the large, nearside lunar basins. However, the relationships between samples collected by the Apollo missions and the Imbrium, Serenitatis, and Nectaris basins have been called into question by new research using samples and orbital data. There is general agreement that Imbrium appears to be 3.92 ± 0.01 Ga, based on Apollo 12 and 14 KREEP-rich melt rocks [5-7]. At Apollo 17, where the mission objective was to sample and date the Serenitatis basin, new work has reinterpreted the Sculptured Hills deposits as having an Imbrium origin [8-10]. The aluminous Descartes breccias from Apollo 16 were originally interpreted as Nectaris ejecta, but new trace-element and age data show they are coeval with KREEP-rich melt rocks interpreted elsewhere as Imbrium ejecta [11]. These updated interpretations reopen the pre-Imbrian impact history to debate, which will only be solved by absolute chronology of samples definitively reset in lunar basins (e.g., SPA, Crisium, Nectaris, Orientale, Schrodinger).

Younger chronology is constrained by mare basalt flows and younger benchmark craters such as Copernicus, Tycho, Autolycus, and Aristillus [12]. Model ages of these craters generally agree with radiometric ages of the Apollo landing sites, but in some places can differ by a factor of 2-3, causing uncertainties in absolute age by up to 1 Gyr [13-15]. Newly-acquired U-Pb ages of samples from Aristillus are ~200 Myr younger than previously proposed ages and new Autolycus crater size-frequency distributions do not correspond to radiometric ages [16, 17]. Crater-density relationships imply that significantly older and younger basalts exist, expanding the active period of the Moon [13, 18]. Older basaltic clasts appear in lunar meteorites [19], but these lack known provenance. On the young end, controversial new results from the LRO mission suggest volcanism may have continued much longer than previously thought in the form of Irregular Mare Patches (IMPs). These areas exhibit a paucity of superposed impact craters that suggests they are younger than 100 Ma, or have physical properties that don’t support craters, and are instead billions of years old [20, 21]. Constraining the chronology of the post-basin era will require measuring radiometric ages of samples with well-established provenance, including young mare basalts and key stratigraphic craters.

Lunar meteorites provide another source of information about impact history from locations potentially far removed from Imbrium. Although relatively few crystallization ages have been determined for impact melt rock clasts, their distribution is broadly similar to that inferred from the Apollo samples and crater density studies of the lunar surface, with the oldest apparent ages around ~4.2 Ga, a peak at ~3.7 Ga, and a declining number of ages to ~2.5 Ga [22-24]. As many of the feldspathic lunar meteorites are regolith breccias, the longer tail to younger ages in the meteorite clasts compared to the Apollo melt rocks may reflect smaller or more localized impact events than the basins sampled by the Apollo sites.

Additional constraints come from impact ages of meteorites derived from asteroid parent bodies [25-27]. H-chondrites show a prominent group of reported 40Ar-39Ar ages between ~3.5 and 4.0 Ga, with the clast-poor, impact-melt rocks LAP 02240 and 031125 yielding especially well-defined plateau ages of 3.939 ± 0.062 Ga and 3.942 ± 0.023 Ga, respectively [26, 28]. Eucrites and howardites, believed to come from the asteroid 4 Vesta, have yielded numerous impact-caused Ar ages between 3.4 and 4.1 Ga, with groups of reliable ages between ~3.5 and 3.8-4.0 Ga, and few such ages between 4.1 and 4.5 Ga, but which may not be readily related to the crater density history revealed by missions like Dawn [25, 29-31]. Mesosiderites commonly yield Ar ages of 3.8 to 4.1 Ga [25], although the very slow cooling experienced by these meteorites following disruption and re-accretion of their parent body complicates the interpretation of these data.

Finally, the relative Martian chronology derived from stratigraphy is not yet tied to an absolute chronology. Confounding variables that contribute to the uncertainties associated with dating by crater density on Mars are the contributions of persistent volcanism.
could benefit from interpretations of their origin. But countless missions to Crisium or Nectaris, would add an enormous amount of data from all destinations where in situ precision (±100 Myr) can be accomplished at multiple locations aboard stationary landers or mobile rovers. The advantage of in situ dating for the Moon and Mars have been considered, developed, and proposed for the highest-priority lunar mission, sample return from the South Pole-Aitken Basin. However, these missions are acknowledged to be large and costly. The Decadal Survey specifically recommends developing in situ dating capability: “New capabilities for in situ age dating are of particular importance, as they can help to provide constraints on models of surface and interior evolution of all the terrestrial planets.” The NASA Office of the Chief Technologist also includes in situ dating in its Planetary Science Technology Needs, specifically recommending maturing age dating to TRL 6 by 2020 and noting its potential use in Discovery, New Frontiers, and future Mars missions.

Multiple groups are developing dedicated in situ dating instruments [36-40]. These instruments are on track to demonstrate TRL 6 readiness by 2020 and will need to be selected in the 2020’s and 2030’s for competed and directed flight missions to relevant destinations where in situ precision (±100 Myr) can provide meaningful constraints on geologic history. The capability for in situ geochronology on the Moon will open up the ability for this crucial measurement to be accomplished at multiple locations aboard stationary landers or mobile rovers. Discovery mission concepts and directed mission payloads that would take advantage of in situ dating for the Moon and Mars have been considered, developed, and proposed [41-44].

Multiple landing sites on the Moon, Mars, and asteroids such as Ceres and Vesta exist where these questions may be answered, whereas returning samples from all these places would be a daunting order. Though some questions will require the precision of laboratory measurements, many more are addressable with an accurate, though perhaps less precise, in situ age. For example, even a single mission to a key basin, such as Crisium or Nectaris, would add an enormous amount of clarity to understanding the cataclysm. A single mission to an IMP would resolve the wide disparity in interpretations of their origin. But countless missions could benefit from in situ dating capabilities, because geochronology is a fundamental measurement for the Moon and other planets. To develop a common framework for absolute ages across the solar system, in situ geochronology instruments should become a common tool, along with our imaging and compositional tools, for landers and rovers on the Moon, Mars, asteroids, and beyond.

**Introduction:** The thickness of the lunar regolith and megaregolith record both the age of the emplacement of the original surface upon which regolith develops, and the duration and intensity of its exposure to the impact flux. The age of lunar mare surfaces and the impact flux intensity are relatively well known. In contrast, the flux intensity and exposure time of the highlands remain relatively mysterious. Estimates of the impact flux prior to 1 Gyr differ, and uncertainty remains with respect to when the lunar crust presented a competent surface into which impacts could produce a recognizable regolith [1].

By coupling remote sensing analysis of the thickness of the highland regolith with a regolith formation model, we explore both the ancient impact flux, and the time when the highland crust consolidated. The thickness of the highlands regolith is typically taken as being 10-20 meters thick [e.g. 2], and described as being dominated by material of size <1cm, with only occasional and relatively rare surface and buried boulders. However, new insights gleaned from remote sensing suggest a finely comminuted regolith layer is much thicker - more on the order of 100 meters than 10. One such data set is derived from the abundance of craters with rocky ejecta. Rocky ejecta are produced from impacts that penetrate any thoroughly pulverized regolith and excavate more competent material from depth. In nighttime brightness temperature data from the Lunar Reconnaissance Orbiter (LRO) Diviner Lunar Radiometer Experiment, clasts ~1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components. The presence of surface clasts on the order of 1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components. The presence of surface clasts on the order of 1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components. The presence of surface clasts on the order of 1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components. The presence of surface clasts on the order of 1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components. The presence of surface clasts on the order of 1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components. The presence of surface clasts on the order of 1 m and larger in size remain warm relative to regolith with smaller particle sizes. To first order, we treat any given Diviner field of view as consisting of a mixture of these two components.

Studies of the optical maturity of small (1 - 5 km) highlands craters also suggest that the highlands regolith is hundreds of meters thick in places. Lucey et al. (2014) [5] isolated a population of optically fresh craters between 1 and 5 km using data from the Kaguya Spectral Profiler experiment, covering the lunar surface between 50N and 50S to inspect the spectral characteristics of the shallow crust. Adjusting for the relative area of highlands and maria, the frequency of craters in the 1-5 km size range exposing immature ejecta in the highlands is less than one third the frequency of such occurrences in the maria. As the depth of excavation of this 1-5 km crater size bin is roughly 100-500 meters, this data set also suggests the presence of an extensive, completely space-weathered layer in places hundreds of meters thick.

![Figure 1. Size frequency distribution of rocky craters in the mare and highlands. The abundance of rocky craters in the highlands is sharply lower at crater sizes below about 10 km.](image)

**Figure 1.** Size frequency distribution of rocky craters in the mare and highlands. The abundance of rocky craters in the highlands is sharply lower at crater sizes below about 10 km.

**Rocky Craters to 5 km:** We have previously measured the size frequency distribution of craters with rocky ejecta as measured by Diviner “rock abundance” measurements in both the highlands and maria for craters larger than 5 km (Fig. 1). The crater size frequency distributions of the two units are very similar at large crater sizes, but begin to diverge at crater sizes below 10 km, where the abundance of rocky craters in the highlands is sharply lower than that in the maria. This suggests that the population of small impactors, which in the maria sample a rocky substrate, are in the highlands beginning to sample in places a rock-free substrate. In other words, the impactor population that produces rocky ejecta in the maria frequently fails to penetrate a rock-poor surface...
layer in the highlands; this rock-poor layer we interpret to be the regolith. This experiment characterized craters larger than 5 km, so while considered with the optical maturity of 1-5 km highlands craters, this observation suggests that a rock-poor layer is locally up to 100 meters thick, assuming a ratio of crater diameter-to-excitation depth of 10.

The Model: We simulate regolith formation using an analytic model that describes the frequency a point at depth is sampled by impacts as a function of time. Gault et al. (1974) [6] presented a pioneering regolith mixing model predicated on the assumption that impact flux is a probabilistic process that obeys the Poisson distribution. In Gault et al. (1974) and our updated and expanded version of the model [7], regolith overturn is defined to occur when a point at depth has been influenced by an impact event (Fig. 2). The success rate of overturning events at depth is presented as a function of core input parameters: time, impact flux, and crater scaling. In previous work, we performed vital updates to the model, including reworking the model to accept crater production functions, and, most importantly, a treatment of secondary impacts. In previous work, we validated model calculations for Copernican-era overturn against reworking rates inferred from meter- to tens of meter-scale albedo features formed during the course of the LRO mission called splotches [8], vast, yet shallow surface density anomalies called cold spots ([e.g. 9]), crater rays [e.g. 10] and the depth distribution of surface-correlated materials in Apollo cores [11]. In previous work we showed that the inclusion of secondary impacts is necessary to model valid reworking rates and that reworking due to a primary impact flux alone is orders of magnitude too slow and shallow [7]. To specifically explore the production of regolith, we include here a further expansion of the model beyond overturn, and describe a zone of pulverization (Fig. 2). We explore the collisional activity in the first billion years by taking advantage of the analytic nature of the model, calculating impact flux from thickness of the highlands regolith.

Model Results: Preliminary calculations using the modern impact rate based on the shape of the Hartmann Production Function (HPF) [12] with secondaries included suggest that at a depth of 5 meters (the accepted average value of the mare regolith), there is a 50% probability that the mare surface has been overturned 10 times over the last 3.5 billion years. To achieve the same gardening efficiency and produce regolith that is 100 meters thick requires the moon have a competent surface to crater extending in time earlier than the solidification of the lunar crust assumed at 4.3 Gyrs or, more likely, a collisional history much more intense than that captured by the HPF prior to Orientale formation. To make 100 m of highland regolith in 4.3 Gyrs, the model calls for an impact flux prior to Orientale that is on the order of a hundred thousand times higher than the modern rate and two orders of magnitude than the impact rate captured by the HPF.

Discussion: Either we may be seeing the signature of a late heavy bombardment in the thickness of highlands regolith or our choice of input parameters is leading the model to under-predict regolith growth. In ongoing work we test and interpret this discrepancy in flux by exploring model input parameters such as the size distribution of secondaries, the changing shape of the crater production function during the first billion years, and by modeling the variability of regolith thickness and the growth of regolith depth from ejecta emplaced by large impacts which a) deposited thick ejecta deposits on the surface; and b) comminuted everything in its path during emplacement. We also plan to expand our remote sensing understanding of the thickness and variability of highlands regolith by including more small rocky craters in our crater counts (less than 5 km) and by performing preliminary quantitative investigations the optical maturity of highlands craters in the diagnostic 1-5 km bin from Lucey et al. (2014) [5].

STABILITY OF D-RIBOSE-Na+MONTMORILLONITE SUSPENSION SYSTEM UNDER GAMMA RADIATION FIELDS AT pH 7 IN A COMET CORE SIMULATION: IMPLICATIONS IN CHEMICAL EVOLUTION.

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Introduction: In chemical evolution, the stability of bio-organic compounds in the surrounding geological environment is as important as their syntheses, especially in the presence of an external energy source (e.g. ionizing radiation). Therefore, there must be a balance between the synthesis and the decomposition of these molecules to have them for other prebiotic processes [1]. Aldoses, in addition to their biochemical interest as energetic molecules or as structural molecules in biological systems, are also of paramount importance in the context of chemical evolution. The synthesis and preservation of aldoses under prebiotic conditions is a fundamental step for the abiotic formation of the nucleotides that make up the nucleic acids (e.g. RNA) [2]. A more plausible geological scenario should involve the participation of a multiphase system, formed by the presence solids/liquids or liquid/ gases interphases [3, 4]. Several solid surfaces may have been relevant in this context: sulfides, carbonates, and clays. In this work, we highlight the possible role of clay minerals due to their physicochemical properties, their broad geological distribution.

The primary objective of this work is focused in studying the stability of D-Ribose-Na+Montmorillonite suspensions at pH 7 under a high radiation field in a comet core simulation. To this end, the radiolysis of the system was carried out by exposing it to a different irradiation dose and ratios aldose-clay. The analysis of these systems was performed by UV spectroscopy and liquid chromatography (HPLC) and HPLC-coupled to a mass spectroscopy.

Our results indicate that the aldose-clay system are relatively stable under irradiation and the radiation-induced or thermal-induced reactions in these systems yield compounds of pre-biological importance.

References:


Introduction: Because the number of asteroids in the inner main asteroid belt (IMB) with absolute magnitudes H<16.5 is now effectively complete, the distributions of the sizes and the orbital elements of these asteroids must be devoid of observational selection effects. This allows us to state that the observed size-frequency distributions (SFDs) of the five major asteroid families in the IMB are distinctly different and deviate significantly from the linear log-log relation predicted by Dohnanyi [1]. One consequence of the existence of these differences is that the mean sizes of the family asteroids, taken as a whole, are correlated with their mean eccentricities and anti-correlated with their mean inclinations. While the latter observation has a simple explanation, we observe that the mean sizes of the non-family asteroids in the inner belt are also correlated with their mean orbital eccentricities and anti-correlated with their mean inclinations [2]. We conclude from this, and from the fact that the SFDs of the non-family and the family asteroids (again taken as a whole) are almost identical, that the non-family and the family asteroids in the IMB have a common origin. We estimate that ~85% of all the asteroids in the IMB with H<16.5 originate from the Flora, Vesta, Nysa, Polana and Eulalia families with the remaining ~15% originating from either the same families or, possibly, a few ghost families.

Implications: These new results imply that we must seek explanations for the differing characteristics of the various meteorites and near-Earth asteroids originating from the IMB in the evolutionary histories of a few, large, precursor bodies. Our findings also support the model that asteroids formed big through the gravitational collapse of material in a protoplanetary disk [3]. Finally, we present evidence (not explicitly stated in [2]), that the “non-halo” asteroids that now surround the major families in e-I space are deficient in small asteroids. This supports our observation [2] that the number of asteroids in a given family, and in a given size range, may peak for asteroid diameters ~1km, suggesting that small IMB asteroids, including future NEAs, are now being lost from the IMB at approximately the same rate as they are being created through the disruption of larger asteroids.

The above conclusions are based on observations of the present asteroid belt. However, if the initial asteroid belt consisted of a comparatively small number of large bodies formed in a near-coplanar, protoplanetary disk, then even after the excitation of the asteroid orbital eccentricities, the initial asteroid collision rate and the rate of supply of small bodies to the inner solar system would have been correspondingly low and would have increased to larger values only at a later date. This slow increase in the early asteroid collision rate could be the reason for the Late Heavy Bombardment.

WHERE ARE THE SHOCKED GRAINS IN THE HADEAN ZIRCON RECORD? INSIGHTS ON THE PRESERVATION OF SHOCKED ZIRCON AND THEIR U–Pb SYSTEMATICS. T. M. Erickson1,2,3, A. J. Cavosie2,3, N. E. Timms2,3, S. M. Reddy2,3, A. A. Nemchin2,3, M. A. Cox2,3, M. Schmieder1,4, D. A. Kring3,4, J. Jacobs – JETS, ARES division, NASA JSC, Houston, TX 77058, USA (timmons.m.erickson@nasa.gov), 2School of Earth and Planetary Sci., Curtin Uni., Perth, WA 6102, Australia, 3NASA Solar System Exploration and Research Virtual Institute 4Center for Lunar Science and Exploration, LPI, USRA, 3600 Bay Area Blvd., Houston TX 77058 USA

Introduction: While the earliest history of many planetary bodies within the inner Solar System is dominated by intense bombardment, this record is missing from Earth due to active tectonics and erosion. Whereas rocks from the earliest history of Earth are absent, mineral relics, such as ancient detrital zircon concentrated in sediments within the Jack Hills, Narryer, Illara and Maynard Hills greenstone belts of the Yilgarn Craton in Western Australia preserve a record of this time.

Shock in zircon: During shock deformation, resulting from hyper-velocity impact, zircon can be modified in crystallographically-controlled ways. This includes the development of planar and subplanar low-angle grain boundaries, the formation of mechanical {112} twins [2], transformation to the high pressure polymorph reidite [3], development of polycrystalline microtexture [4], and dissociation to its dioxide constituents SiO₂ and ZrO₂ [5].

Shock Effects on U–Pb systematics: Shock microstructures can also variably affect the U–Pb isotopic systematics of zircon (e.g., [6, 7, 8]) and, in some instances, be used to constrain the impact age. While shocked zircon containing planar microstructures including (100) deformation bands, {112} shock twins, and curvi-planar fractures can be partially reset such that the lower intercept Concordia age is consistent with the impact event [6], this record is often obscured by subsequent tectonothermal events [7, 8]. The U–Pb ages of granular zircon neoblasts [9, 10] are completely reset during impact event, whereas other polycrystalline textures may be partially to completely reset.

Longevity of shocked zircon: Although zircon is extremely refractory and can survive multiple tectono-thermal reworking events, erosion and deposition, shocked zircon is pervaded by defects and grain boundaries. Nevertheless, detrital shocked zircon derived from the Vredefort Dome in South Africa has been shown to survive long distances of transport within the Vaal River [7, 11], deposition in paleo fluvial terraces [12] and delivery to the Atlantic coast >2000 km from the crater [13]. Detrital shocked zircon grains have also been found in sediments from the highly tectonized Santa Fe impact structure [14] and within fluvial and glacio-fluvial sediments of the Sudbury impact structure [15]. While shocked zircon derived from the target bedrock of Vredefort Dome is found throughout the Vaal basin, the U–Pb systematics of the grains do not record an impact age [7, 9].

Hadean shocked zircon (or lack thereof): Despite thousands [this study] to tens of thousands of grains surveyed [16] no convincing shocked zircon grains have yet been reported. This may in part be due to the limitation of these studies to only surveying the Ernondoo Hill W74 discovery site in the Jack Hills. However, it suggests that the surviving Hadean zircon population is biased against shocked grains. Although shocked zircon grains with planar microstructures including twins and reidite lamellae can survive long distances of transport, it is still unclear whether polycrystalline aggregates can survive transport, sedimentation and diagenesis, or moderate metamorphic overprinting. Furthermore, while the volume of U–Pb shock-reset zircon from ancient craters may be large [17], these grains may be susceptible to subsequent age resetting, so caution should be used when interpreting U–Pb data from any future shocked grains discovered.


Figure 1. Scanning electron BSE image of Vredefort-derived shocked zircon from the Vaal River after [7].
INTRODUCTION

Tera et al.’s [1] concept of a lunar cataclysm at ~3.9 Ga grew from observations of element fractionation in samples of the heavily cratered lunar highlands. Specifically, U-Pb – and some Rb-Sr – data suggested apparent recrystallization ages between about 4.0 and 3.8 Ga. Although the parent/daughter behavior in these two geochronologic systems are quite different (i.e., both U and Sr are highly refractory, while Pb and Rb are variably volatile), both revealed evidence of some degree of isotopic disturbance, albeit in very different ways. Whole-rock Rb-Sr systems appeared largely intact, but the alkali-rich, interstitial groundmass, termed quintessence, gave younger ages. For example, Apollo 17 sample 76055, a recrystallized breccia, yielded an internal Rb-Sr isochron age of 4.49±0.05 Ga, but the coexisting quintessence indicated an age of 3.86±0.04 Ga assuming Rb-Sr isotopic equilibrium with the whole rock. Their U-Pb analyses of samples of fifteen Apollo 14, 15, 16, and 17 rocks scattered about a concordia array with an approximate intercept age of 3.9 Ga. The contrasting dating results obtained using the different chronometers represented something of a conundrum as Tera et al. (1974) suggested that U-Pb fractionation was due to Pb volatilization during metamorphism explained the young intercept age, but were forced to argue that Rb-Sr whole-rock systematics were not similarly affected. They acknowledged that the absence of a mechanism to explain this differential behavior “remains a basic problem”. Forty-four years later, this phenomenon is still unexplained.

The Late Heavy Bombardment (LHB) concept [2] was initially controversial [3], but ⁴⁰Ar/³⁹Ar step-heating ages of lunar impact melt rocks collected during the Apollo missions appeared to consistently yield ~3.9 Ga “plateau ages”, confirming the hypothesis. The LHB came to be regarded by many as one of the great scientific achievements of the Apollo explorations. Whole-rock ⁴⁰Ar/³⁹Ar ages of lunar impact melt breccias were interpreted as dating several of the large, nearside impact basins. These ages, in turn, have been used as the foundation for “crater count” chronologic calibration [4] that are used to date other planetary surfaces. Despite the influence that the lunar ⁴⁰Ar/³⁹Ar dataset has had on planetary research and the lack of an LHB-age spike in lunar meteorite ⁴⁰Ar/³⁹Ar ages (likely representing a more random spatial sampling of the lunar surface), there has been little in the way of critical examination of the underlying assumptions of these age interpretations over the past four decades.

PROBLEMATIC ⁴⁰Ar/³⁹Ar STEP-HEATING INTERPRETATIONS

The ⁴⁰Ar/³⁹Ar step-heating method has the potential to reveal intragrain isotope variations and is thus often used as a monitor for evaluating deep time thermal disturbances. However, this capability has several method-specific requirements that, if not met, preclude thermochronologic interpretations. Three such issues effectively rule out the use of virtually all lunar ⁴⁰Ar/³⁹Ar data as support for the LHB hypothesis: 1) the “plateau age” approach used is an aphysical concept for the thermally disturbed samples typical of most impact melt rocks brought back by the Apollo missions, 2) laboratory artifacts (e.g., recoil, size effects) destroy preserved diffusion information, or create false apparent age gradients; [5], and 3) the fact that different K-bearing phases in extraterrestrial samples have different activation energies for Ar diffusion requires thermal cycling during laboratory step-heating experiments to obtain meaningful thermal history information [6].

IN SITU GEOCHRONOLIC STUDIES

Advancements in mass spectrometer sensitivity and extraction system backgrounds now permit ⁴⁰Ar/³⁹Ar dating to be undertaken on small (10s of micron diameter) spots on thin sections using a laser to degas irradiated samples. In principal this approach has the potential to transcend the analytical challenge posed by the continuous impact reworking of the lunar regolith that produces fine-scale polygenetic breccias of multiple age and origins.

For example, Mercer et al. [7] undertook laser microprobe dating of two Apollo 17 impact melt breccias. While one sample yielded evidence of a single melting event at 3.83±0.02 Ga, data from the second yielded texturally correlated age peaks at 3.81±0.01, 3.66±0.02 and ca. 3.3 Ga. All dates are significantly younger than the age classically ascribed to Serenitatis [8, 9]. However, an earlier study of the same sample [10] yielded SIMS U-Pb dates for various accessory minerals that were interpreted as representing two generations of impact melts at 4.335±0.005 and 3.934±0.012 Ma. Thus, two different in situ dating methods appear to yield chronological data from the same sample that range over a billion years. The simplest explanation for this paradoxical result derives from the different retentive behaviors of Pb in accessory minerals and Ar in K-bearing minerals: the long and
complex thermal histories of most impact melt breccia samples means that multiple chronometers are more likely than not to yield discrepant results. The results for any single chronometer are unlikely to be definitive, but the application of multiple chronometers to the same samples can provide a rich understanding of the polygenetic nature of most impact melt breccias on the lunar surface.

Establishing a reliable, quantitative planetary impact chronology requires that all analyzed rocks are equally suitable for the application of specific chronometers. In the case of K-Ar dating, the broad presence of at least trace amounts of potassium in lunar rocks provides a somewhat level playing field (although rocks with a significant KREEP component will tend to have higher signal-to-noise). However, there is an underappreciated aspect of using accessory minerals for dating lunar events that limits their value in establishing global phenomena. The presence of these phases in an impact melt depends dominantly on its thermal state and chemistry, and not every target rock composition/impact condition couple can lead to formation of accessory phase crystals large enough for in situ chronologic investigation (>10 μm). This imposes an inherent bias in using them to establish impact chronologies, as the KREEP component is the likely source of the high levels of P and Zr needed to crystallize magmatic apatite and zircon. In general, the anorthositic highland terranes are low in incompatible elements, whereas the spatially limited Procellarum KREEP-rich terrane is characterized by high values [11]. Thus, impacts into such KREEP-rich rocks are more likely to produce dateable accessory minerals than virtually all other lithologies. Studies that exclusively rely on in situ accessory mineral U-Pb dating are inherently biased toward documenting impact events in KREEP-rich terranes, which make up only a small fraction of the lunar surface [11].

Other considerations: The longstanding assumption that lunar melt rocks had to have originated in large, basin-forming events has been challenged by Lunar Reconnaissance Orbiter Camera (LROC) images that document impact melt deposits in lunar highland craters as small as 170 m [12]. LROC images also have radically altered another longstanding assumption: about the distal distribution of impact debris from basalt-forming events. For example, a global map shows clusters of 'light plains' deposits [13] radiating outward more than 2000 km from the rim of the ~930 km diameter Orientale Basin, underscoring how poorly the spatial relationships between large basins and their surrounding deposits were understood when impact chronologies were being developed in the 1970s. Thus, the assumption that a specific lunar melt rock from a given landing site is representative of one of the basin-forming impacts is deeply flawed; a lunar grab sample could be from a basin-forming impact, a tiny, inconsequential impact of arbitrary age, or a mixture of those materials. And, of course, impacts are not the only source of heat early in lunar history capable of resetting K-Ar systematics. For example, juvenile basaltic magmatism began early on Moon and continued episodically for more than three billion years.

Lastly, we address a longstanding LHB meme for which we know of no basis. For example, Bottke and Norman [14] argued that because “pronounced clustering of lunar melt rock ages in the interval typically associated with the Terminal Cataclysm is apparent in diverse isotopic systems, including Rb-Sr, Sm-Nd, U-Pb, and $^{40}$Ar/$^{39}$Ar”, the “excellent agreement obtained by multiple techniques provides confidence that the ages are meaningful”. They did not, however, specifically point to an example of such intrasample concordancy and we are unaware of any such data.

Summary: The way forward begins with in situ investigations of petrologically well-documented, KREEP-rich, lunar melt samples using multiple radiogenic geochronometers (e.g., laserprobe $^{40}$Ar/$^{39}$Ar, SIMS U-Th-Pb). Such studies have not, to our knowledge, been undertaken and would provide ground truth information regarding the expectation of concordancy as evidence of an impact origin. If such concordance is gained, then a massive campaign focussed primarily on $^{40}$Ar/$^{39}$Ar laserprobe dating of Apollo samples and lunar meteorites might reveal unbiased age patterns that could be interpreted in terms of bombardment history. But until such measurements are made and the accumulating concerns of the classical interpretation of lunar geochronologic data mentioned above are addressed, the LHB concept should be viewed more as a speculation than proven concept.

References:
Introduction: “Terminal cataclysm” was originally defined simultaneously by two independent lunar sample analysis groups, ca. 1973, as a spike in impact rate, during which most of the prominent multi-ring impact basins formed on the moon (and perhaps other planets) during a brief (~150 Ma) interval, preceded by several hundred Ma with a low, possibly negligible, flux of large impactors [1,2]. The dynamicist, Wetherill, in the title of a 1975 paper [3], introduced an alternate term, “late heavy bombardment,” currently called “LHB,” but he specifically equated that term to the terminal cataclysm concept of Tera et al. [1], and he looked for dynamical explanations of low impact rates in the first few hundred Ma, followed by a sudden spike.

This concept of a cataclysmic impact spasm was strongly supported after impact melts were more clearly recognized; a 1990 study by Graham Ryder showed that impact melt rocks among Apollo samples showed a very sharp spike at about 3.85 to 4.0 Ga ago [4]. He explicitly defended a concept that I have called “Ryder’s Rule”: lack of impact melts before ~4 Ga ago proves lack of impacts during that time.

The “terminal cataclysm paradigm” influenced many subsequent interpretations of planetary history. For example, one of the main rationales for dynamists’ introduction of the “Nice model” of solar system evolution was that it could explain the impact spike at 3.9 Ga ago. (As made clear by the authors, this explanation worked only if certain giant planet migration effects, where timescale were free parameters in the model, were assumed to have happened at 3.9 Ga ago) [5].

The terminal cataclysm/LHB paradigm now appears to be collapsing because of contrary empirical evidence. For example, a Ryder-like spike in impact melts clasts is absent not only in asteroidal meteorites, and also in KREEP-poor lunar meteorites (which art thought to represent regions of the moon not sampled by the central front-side Apollo landings) [6,7]. In addition, increasing numbers of studies report lunar impact melt clasts in upland breccias, clustering at dates such as ~4.33 an ~4.22 Ga ago [8, 9].

As widely discussed in the 1960s, a famous book by Kuhn [10] pointed out that such “paradigm shifts” are important moments in science, and it is in that spirit that we attempt to revisit the current situation in our community. To take an example of “paradigm fallout” from our case: Even as the planetary science community pulls back from the terminal cataclysm paradigm, biologists and terrestrial geologists continue to cite the terminal cataclysm or LHB as a constraint on the origin of life, and on the interpretation of Hadean geology.

Misuse of terminology – and the consequences. It is important to note that the original definition of LHB [3] involved a short burst of impactors around 3.9 Ga ago. Recent years, however, have seen a tendency in our community to maintain use of the term LHB, even while subtly changing the definition to new various concepts completely contradictory to the original. An example comes from dynamical modelers, who until 2010 quoted the original LHB definition. By 2012, in an effort to reverse course and get rid of previous models’ spike at 3.9 Ga ago, the same group (largely) spoke of extending “the LHB,” not only backward in time to 4.1 Ga ago or more, but also forward to as recently as 2.5 Ga. The once formidable, cataclysmic LHB was now thus quietly reduced to a gentle swell, lasting as long as 1600 Ma --- but the same name was kept. The problem with this semantics approach is that it perpetuates the idea that the terminal cataclysm/LHB paradigm remains as viable as ever, thus misleading other scientific communities.

Solution to the problem of sample ages: Early critiques of the terminal cataclysm/LHB paradigm proposed that the reason for paucity of pre-4.0 lunar sample dates was not absence of impacts before 4.0 Ga ago, but the opposite --- an enormously high pre-4.0 impact rate. It was shown that this approach could also produce a spike in ages of recovered impact melt samples [3, 13, 14].

This idea was rightly criticized for not adequately explaining why lunar impact melt rocks from before 4.0 Ga ago were nearly absent, while numbers of primordial crustal rock fragments from well before 4.0 Ga were available in the lunar sample collection.

It is proposed here that the answer lies in the evolution of the megaregolith. Five key factors are involved: (1) The impact melt lens of a given early basin, formed at time T (say, 4.3 or 4.2 Ga ago), covers only a tiny fraction of the moon with impact melts of age T. (2) Such a newly formed impact melt lens extends to only modest depth D below the exposed floor of the new basin, and thus immediately begins to be rapidly pulverized by the proposed, pre-4.0 intense bombardment, forming breccia clasts. (3) The near-surface, impact melt depth, D, is assumed to be less than the modern depth of megaregolith that has been...
pulverized during the age of the moon. (4) At the base of the megaregolith lies a semi-infinite reservoir of intact or moderately fractured, early igneous lunar crust. Fragments of it can be ejected onto the surface by the largest post-mare impact craters, such as Tycho and Copernicus. (5) Boulder-sized rocks on the lunar surface during the last few Ga survive only some few hundred Ma [15, 16]. Putting these factors together, we see lunar samples collected by astronauts, landers, or rovers thus rely on recent impacts to eject subsurface rocks that, rather than being intact samples of the entire lunar history, reflect the history of lunar megaregolith processing. This model is supported by findings from GRAIL gravimetry, indicating lower density and higher porosity in the lunar near-surface layers, “to depths of at least a few kilometers” [17]. Early crustal materials from the semi-infinite global reservoir at depth are constantly replenished, but few intact impact melts from the lower-volume, finite-sized, earliest multi-ring basins survive.

Acknowledgment: Thanks to the International Space Science Institute, in Bern, for several invitations to serve as a visiting scientist, during which I have produced a lengthy, epistemological study of the paradigm shift regarding terminal cataclysm/LHB. I am currently pursuing options to publish this study, I hope in open access form that might be freely available to classrooms.

WHAT HAPPENED 800 MILLION YEARS AGO ON THE MOON? Ya-Huei Huang¹, David A. Minton¹, Jacob R. Elliott¹, and Nicolle E. B. Zellner². ¹Department of Earth, Atmospheric, and Planetary Sciences, Purdue University, West Lafayette, Indiana 47907 USA (huang474@purdue.edu). ²Department of Physics, Albion College, Albion, Michigan 49224 USA.

Introduction: Lunar impact glass spherules are common in lunar regolith and thought to be formed by relatively small impact craters. The ⁴⁰Ar/³⁹Ar-derived age distributions of glass spherules provide some insight to the bombardment history on the Moon. For example, several researchers reported an excess of impact glass spherules in Apollo regolith samples that formed during the last 500 Ma [1-3]. Although it may imply an increase in the recent impact flux, other physical processes, such as size-dependent argon diffusion process, the shallow sampling depths from which samples were collected can potentially bias the samples toward a young age [4,5].

While the ⁴⁰Ar/³⁹Ar-derived age of an impact glass spherule is set at the time of an impact event, the geochemical analysis of an impact glass spherule allows us to interpret the provenance location on the lunar surface. In particular, glass spherules are considered as “exotic” when their geochemistry indicates that they originated a large distances from the sampling location [6]. As ejecta debris produced by a large crater is capable of traveling a long distance, the ⁴⁰Ar/³⁹Ar-derived age distributions of exotic glass spherules are a potential impact record for large cratering events.

Considering only exotic glass spherules, as well as correction for a bias arising from diffusive loss of argon, the ⁴⁰Ar/³⁹Ar-derived age distributions of those exotic glass spherules from Apollo 14, 16, and 17 (Apollo 12 ropy glasses) regolith sample show a higher concentration in ages of 700-900 Ma [7]. The formation ages of 800 Ma-old exotic glass spherules coincide with the possible age for the formation of Copernicus Crater [8]. Together with the Copernicus Crater-forming event, it suggests that the Moon may have experienced a spike in the rate of impacts 800 Ma ago (Figure 1); however, an updated crater count for Copernicus Crater fits with lunar Neukum crater chronology using a constant impact rate assumption [9]. The coincidence between an excess of 800 Ma-old exotic glass spherules and the formation age of Copernicus Crater motivates us to consider other hypotheses.

Hypothesis: Here we investigate how the Copernicus Crater-forming debris affects an impact record that is interpreted from impact glass spherules. We pose the hypothesis that Copernicus Crater ejecta fragments can generate a sufficient number of exotic glass spherules as seen in Apollo regolith samples. Sesquinary (1.5-ary) impact craters are formed via escaped ejecta from the impact on a satellite that went into the orbit around the primary and later hit a satellite [10]. The impact velocity of hypothetical Copernicus Crater’s sesquinaries, just above lunar escaped velocity, is enough to produce melts that were subsequently ejected and formed melt droplets (precursor of glass spherules). If so, Copernicus Crater-forming sesquinaries are a potential candidate for a global impact event because it takes less than 10,000 years from geocentric orbit phase to re-impact the Moon [11].

Figure 1: The relative impact probability of exotic glass spherules from Apollo 14259, 64501, 66041, 71501 [7], and Apollo 12 ropy glasses [12]. The total number of exotic glass spherules is eight.

Yet, in order to make Copernicus Crater’s fragments consistent with observations, we must satisfy two main observational constraints: 1) the fraction of Copernicus Crater’s ejecta that re-impacts with sufficient velocity to generate glasses (secondaries and sesquinaries) is sufficient, and 2) the sizes of secondary and sesquinary impact craters can account for the exotic origin of impact glass spherules in our observed dataset (Figure 1). If both constraints are fulfilled, we are more comfortable with this explanation that the excess of 800 Ma-old exotic glass spherules can be accounted for by the formation of Copernicus Crater ejecta fragments.

Methods and Results: For the first constraint, we performed two independent simulations, impact cratering and orbital dynamics, to investigate the fraction of ejecta from Copernicus Crater that initially escapes the Moon and later re-impacts the Moon. The impact cratering simulation uses a hybrid approach of the SALES_2 and iSALE codes to estimate the total amount of escaped ejecta, as well as the size frequency distribution of fragments [13-15]; we were able to obtain a relation between ejection velocity and fragment size for Copernicus Crater. The impact conditions for
Copernicus Crater simulation uses 7 km diameter basaltic impactor with the impact velocity of 10 km/s to vertically impact the basaltic target material. The largest fragment size is ~120 m in diameter, and the mean fragment size is 4-5 m in diameter. The total number of fragments is $\sim 10^8$ for escaped fragments larger than 1 m in diameter (Figure 2).

Then, the orbital dynamic simulation using the Python and C++ based code, REBOUND [16], is to examine the fraction of escaped material at a given launch velocity that later comes back to the Moon. For each run with the same random seed and a given launch velocity (2.4 – 3.2 km/s), we distributed 100 test particles uniformly 1 km above the lunar surface with velocity radiating from the Moon. The position and velocity of massive bodies (Sun, Earth, and Moon) refer to the J2000 date from JPL HORIZON database. We found that for test particles launching at 2.4 km/s 5-6% of them came back to the Moon, and 1-2% of them hit Earth within 160 years. Beyond 2.4 km/s, less than 1% hit the Moon, and the fraction of test particles hitting Earth increases. Our calculations appear consistent with the previous estimates [11,17].

For the second constraint, we want to know the minimum distance from which part of the lunar surface the geochemical data of an exotic glass spherule from our observed data set is matched. Assuming that the geochemistry of glass spherules reflects the provenance of a target bedrock, we utilized the abundance maps of TiO$_2$, MgO, Al$_2$O$_3$, and K$_2$O interpreted by Lunar Prospector Gamma-Ray Spectrometer [18] and performed a linear least squares regression for each exotic glass spherule. We found that one Apollo 16 KREEPy glass spherule and one Apollo 14 highland-like glass spherule may come from >500 km away from their local sites, and one Apollo 17 glass spherule may have derived from >200 km from the site.

However, the largest sesquinary crater from SALES_2 simulation, as determined by the $\pi$-group scaling laws, is only a few kilometers in diameter, and the number of hundred meters sized fragment is very few. Considering our three exotic glass spherules were delivered from a hundred kilometers, it requires craters larger than just a few kilometers. The mean fragment sizes (~4-5 m in diameter) of escaped fragments from our SALES_2 run suggests that sesquinary impact craters with hundreds of meters of diameter are most likely to form globally, rather than kilometer-sized craters. We propose that Copernicus Crater’s sesquiniaries may have formed locally throughout the whole lunar surface, but those locally-forming sesquinary glass spherules in large distances can later be transported to a local site.

**Discussion:** Our proposed multiple-impacts origin of exotic glass spherules are supported by the existence of most exotic samples that include not only exotic glass spherules but also multiple generations of microbreccias within a single impact breccia. For example, Apollo 14315, an unusual regolith breccia, contains microbreccias, and its composition includes from highland basalts, low KREEP Fra Mauro, and KREEP components, demonstrating its interesting journey to the Apollo 14 landing site.

We also tested what probability of finding exotic glass spherules is if those exotic sesquinary-forming glass spherules were formed 500 km away from a local site. Initially, the exotic sesquinary-forming glass spherules were placed at >500 km away from the center of our grid space. The grid space uses the area of a lunar surface, 6000 km by 6000 km (1 pixel = 2 km). We then ran an 800 Ma-long impact bombardment simulation and calculated the probability of those exotic sesquinary-forming glass spherules at the center region. Interestingly, it is about 50% of probability within the center region of 10 km by 10 km.

![Figure 2](image)

**Figure 2:** Cumulative number of fragments higher than ejection velocity from our SALE_2 result. The x-axis is the ejection velocity of escaped fragments, and the y-axis is the cumulative number of escaped fragments higher than ejection velocity. The total number of escaped fragments is $\sim 10^8$ (>1 m).

**Reference:**
OBSERVING GIANT, PLANET FORMING IMPACTS IN EXOPLANETARY SYSTEMS.
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Introduction: Giant impacts – collisions between similarly-sized, gravity dominated bodies – are a key component of planet formation, representing the mechanism for the final assembly of terrestrial planets[1,2]. In our own inner solar system giant impacts have been proposed to explain Mercury’s large core fraction [3], the formation of the Moon [4], and the Martian hemispheric dichotomy (MHD)/Borealis basin impact [5]. Giant impacts have a wide range of possible outcomes, from efficient mergers to disruption[6], but the production of substantial quantities of small debris is universal. Averaged over a typical sequence of impacts the formation of an Earth mass planet is expected to result in the release of around a Mars mass of debris[7].

Dynamical evolution of the debris: Once released during the impact the debris will enter into orbit around the parent star. Initially the debris will be in a tight clump (Fig. 1a), however since the debris is launched with a range of velocities it will have a range of different orbital periods and so Keplerian shear will spread the initial clump out into an arc and then a spiral (Fig. 1b). The spiral will wind tighter and tighter until the individual windings merge into a smooth asymmetric disk (Fig. 1c). The pinch that can be seen at the right in Figs. 1b,c and the face-on view in Fig. 2 is the collision-point, this is the point at which the original giant impact occurred, and since all of the debris originates from this point all of the debris orbits must continue to pass through it, leading to the pinch observed which is a fixed point in space. Since all of the orbits pass through the disk mid-plane at the collision-point they also all share the same line of nodes, which leads to existence of a line on the other side of the star, the anti-collision line, along which all of the orbits again cross the disk mid-plane. This leads to the bow-tie like structure seen in Fig. 2 when the disk is viewed along the line of nodes. Finally over longer timescales differential precession smears out the collision-point, eventually producing an axisymmetric disk (fig. 1d).

Vapour production and optical thickness: For Mars-sized bodies the escape velocity is equal to the sound speed in silicates (~5 km/s) and thus we can expect shocks to be a ubiquitous feature of giant impacts. Shocks substantially heat material and some will be vaporised. This rock vapour subsequently

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig1.png}
\caption{Dynamical evolution of giant impact debris created at (1,0) at $t = 0$ from a progenitor on a circular orbit. Top left: appearance after 0.2 orbits. Top right: after 2 orbits. Bottom left: 200 orbits. Bottom right: 10 000 orbits. Precession due to a Jupiter-mass planet at 0.2 semimajor axis units is included. All images are normalized individually and have a Gaussian smoothing with FWHM 0.05 semimajor axis units applied.}
\end{figure}

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig2.png}
\caption{Detailed view of the asymmetric disc phase with logarithmic density map. Face-on view at top left, edge-on views at right and bottom.}
\end{figure}
recondenses into droplets with a characteristic size of mm to cm[8].

1% of an Earth mass in mm-size grains has a cross-sectional area of 0.8 AU². It is thus easy to see that the production of vapour condensates can easily lead to an optically thick debris cloud.

**Variability due to changing optical thickness:** As the initial clump of debris moves away from the collision-point it expands, and for an optically thick dust cloud this expansion increases the available absorbing/emitting area and the cloud increases in brightness. When the clump then approaches the anti-collision line and is funnelled down into a smaller volume the available absorbing/emitting area will decrease again. Overall this leads to two peaks and drops in brightness every orbit.

If we view the disk in a close to edge-on configuration then an additional effect comes into consideration from the disk ansae. When the clump of debris passes through the disk ansa the clump is oriented with its long axis along the line of sight, increasing the column of dust we are looking through and decreasing the emitting area we can see, resulting in a drop in the brightness of the disk. The disk ansae thus also lead to two peaks and drops in brightness every orbit. We then have two sets of peaks and drops with the relative phase between them determined by the angle between the collision-point and the disk ansa. This leads to complex behaviour that can be apparently bi-periodic depending on the viewing orientation, as shown in Fig. 3. The shape of the curve also depends on the quantity of dust as this influences how optically thick the cloud is, and on the velocity dispersion in the debris as this influences how rapidly the initial clump is sheared out.

By examining the light curve of a system we can gather a wealth of information about the forming planetary system, including the mass and orbital location of the bodies that collided. With spectra we can also access the composition of the forming planets.

**Example systems:** Our team have been monitoring a small sample of bright debris disc systems to look for variability resulting from ongoing planet formation. Two systems are particularly good candidates for the processes described here. P1121, a G9V star in the 80 Myr old open cluster M47, and ID8, a G6V star in the 35 Myr old open cluster NGC2547. More details on the observational program can be found in the abstract of Su et al.

**References:**

Differential in Impact Melt Sheets as a Mechanism to Produce Evolved Magmas on Mars. A. Koeppel1,2, B. Black2 and S. Marchi3, 1Northern Arizona University, Department of Physics and Astronomy, NAU Box 6010, Flagstaff, AZ 86011, ak2223@nau.edu; 2Department of Earth and Atmospheric Sciences, The City College of New York, New York City, NY 10017; 3Southwest Research Institute, Boulder, CO 80302.

Introduction: The discovery of evolved silica and alkali-rich igneous rocks on Mars (up to 67 wt % SiO2 and up to 14 wt % Na2O + K2O) attests to the diversity of igneous processes that have shaped the planet’s crust [1, 2, 3, 4]. These materials contrast with the basalts that comprise the bedrock and dust across most of Mars’ surface, indicating a degree of magmatic differentiation not typically seen in Mars’ effusive lavas [1]. As in situ rover measurements [3], high resolution orbital spectra [5, 6] and analysis of Martian meteorites [7] expose greater abundances of felsic and silica and alkali-rich material on Mars, it is important that new models for Mars’ crustal composition provide explanations for the origins of the observed heterogeneity.

Hypervelocity impacts are known to melt enormous volumes of rock and may generate the necessary conditions for magmatic differentiation. Petrological studies of impact structures on Earth and the Moon indicate that chemical differentiation is widespread in large impact melt sheets [8].

In this study, we investigate the petrologic evolution of Martian impact melt bodies using the thermodynamic phase equilibria calculator MELTS. We identify signatures of differentiation in impact melt sheets under a range of conditions and compare geochemical trends with measurements of Martian meteorites and evolved rocks from Gusev and Gale craters.

Methods:

Melt and crater scaling. We use the scaling equations outlined in Abramov et al. [9] to estimate volumes of impact melt produced by shock decompression melting and retained within the crater. Melt volumes are scaled for a range of representative impact sizes based on observed rim-to-rim diameters of well-known craters on Mars. This range includes: Gale (154 km), Schiaparelli (459 km), 700 km, Isidis (1500 km), Argyre (1800 km), Hellas (2300 km), and Utopia (3300 km).

In order to calculate melting depth, we assume an initially spherical geometry to the melt due to the radial propagation of the shock front (see Figure 1a) [10, 11, 12]. We place the top of this sphere at Mars’ surface. Following Cintala and Grieve [13], we assume the melt is well-mixed and that material crystallizing within the melt sheet is representative of the stratigraphy sampled by the spherical melt region. This is a basic approach to approximating initial melt volume and composition that does not account for other forms of melt production such as frictional melting or heating and decompression of near solidus material beneath the crater.

For estimating relative amounts of crust and mantle in the melt sheet, we use an average crust-mantle boundary depth of 50 km with a 2900 kg/m³ crust and 3500 kg/m³ mantle [14, 15].

Melt sheet thickness, and thus pressure at the melt sheet’s base, is calculated using a crater diameter-to-depth ratio from Tornabene et al. [16], derived from Mars Orbiter Laser Altimeter observations of complex craters. We let the apparent depth of a crater roughly represent the distance to the top of the melt sheet.

The initial composition of impact melts. For a primitive mantle composition, we use the Dreibus and Wänke [17] estimate of the parent body for SNC meteorites. For the crust composition, we use Taylor and McLennan’s average crust [18], which is derived from globally averaged soil and dust compositions.

We explore two extremes for possible water contents. “Dry” models are run with an anhydrous crust.
and 35 ppm H₂O mantle [17]. In the alternative, we adopt a “wet” crust concept from Ivanov and Pierazzo [19], and assume a wet altered Martian crust with 5.2 wt % H₂O, which is based on upper measurements for water contents in altered terrestrial oceanic basalt [20]. This could represent structurally bound water in hydrothermal minerals and pore fluids in the upper crust.

Measurements of mineral assemblages, trace element partitioning, and oxide compositions in Martian meteorites constrain oxygen fugacity (fO₂) in parental melts from very reducing conditions (near the Iron-Wüstite buffer) to somewhat oxidizing conditions (above the Fayalite-Magnetite-Quartz buffer) [21].

**Modes of differentiation.** MELTS simulations are run at two crystallization settings during cooling: equilibrium and fractional crystallization. If equilibrium crystallization behaves ideally, the solidified melt sheet composition will reflect the bulk composition of the liquidus magma. However, compositions along an equilibrium crystallization liquid line of descent may represent partial melts or melt from that point in the crystallization process that has become chemically isolated.

**Discussion:** Our modeling suggests that enrichment of SiO₂, Na₂O, K₂O, and Al₂O₃ as observed in some Martian rocks depleted in MgO (< 5 wt %), is achieved through the extended crystallization of olivine that occurs when feldspar and pyroxenes are suppressed, possibly due to dissolved water. For the initial compositions and melt sheet crystallization pressures we considered, fractional crystallization alone cannot explain the most alkali- and Al₂O₃-rich Martian samples. Liquid evolution during equilibrium crystallization provides a better fit to enrichment trends and may represent partial melting or segregated melts produced by impacts. Still, a potassium-rich initial melt composition is required to explain the potassium contents found in rocks such as the mugearite Jake_M in Gale crater [22].

Because exhumed Gale crater rocks sample ancient Martian crust, the alkaline differentiation trend in Gale crater has been interpreted as potential evidence for igneous processes that operated primarily in the Noachian [3], when impact processes were more numerous. Even if impacts were not responsible for the alkaline differentiation trends in Gale crater, impact melting and differentiation may have played an underappreciated role in shaping the petrology of the earliest Martian crust.

**References:**

EVALUATING THE RELATIVE CONTRIBUTIONS OF ASTEROIDS AND COMETS TO THE INNER SOLAR SYSTEM DURING THE FIRST BILLION YEARS. David A. Kring1,2, 1Center for Lunar Science and Exploration, Lunar and Planetary Institute, Universities Space Research Association, 3600 Bay Area Blvd., Houston TX 77058 (kring@lpi.usra.edu), 2NASA Solar System Exploration Research Virtual Institute.

Introduction: The ancient cratered highlands of the Moon provide the best and most accessible record of early bombardment in the Solar System. Samples from the highlands and, in particular, the margins of three nearside basins, reveal a surprising pattern. Their ages are dominantly ~3.9-4.0 Ga, which was interpreted to mean there was an increase in bombardment at that time (e.g., [1,2]) described as a lunar cataclysm. Strictly speaking, the samples only addressed the timing of the latter third of the basin-forming events. The timing of earlier events remains murky.

Testing the lunar impact cataclysm hypothesis is the highest priority science investigation when we access the lunar surface again [3]. The importance of bombardment to the evolution of the Moon and the need to better understand it is echoed in a recent document prepared at the behest of the NASA Associate Administrator for Science [4] and in a re-affirmation of a National Research Council report [3] by the Lunar Exploration Analysis Group [5].

While the magnitude and duration of the bombardment are immensely important, it is the implications those data have that really tug our interest. What caused the bombardment? What was the source (or sources) of the impacting material? What caused it to pummel the Moon and, by inference, the Earth and other terrestrial planetary surfaces? How did that bombardment affect the early evolution of those surfaces and, at least in the case of the Earth, the early evolution of life. Each of those issues is to be addressed in The First Billion Years: Bombardment topical conference. In this paper, I address the progress we have made in assessing the sources of the impacting material.

Evolution of Ideas: Because astronauts recovered samples of impact melt from the basin-forming epoch, the samples could be probed for chemical signatures of any projectile material entrained in them. The preferred tracers are siderophile elements, because the Moon’s native abundances were sequestered into the core and mantle. Thus, siderophile elements in impact melts produced in the lunar crust are largely derived from the impactors. Kring and Cohen [6] reviewed that data and found impact melts contained signatures of ordinary (OC) and/or enstatite chondrites (EC) and iron meteorites. Importantly, there did not seem to be any trace of CI or CM carbonaceous chondritic materials, which are our best proxies for comets. Thus, they concluded, “Comets were not important during this time.”

We then posed another question: Is there a geological fingerprint of the impacting projectiles? The answer is yes – in the form of the size frequency distribution of craters produced in the highlands. In a novel integration of two methods, Strom et al. [7] measured the size frequency distribution of observed craters and then used pi-scaling techniques to evaluate the size frequency distribution of projectiles that produced them. While the impact parameters for any individual impact can vary, the values for the entire population can reasonably be assumed average (e.g., with a 45 degree impact trajectory). The calculated size frequency distribution matched that of main belt asteroids and not that of comets and Kuiper Belt Objects known at that time. It is important to note that the population of the largest basins is small, which implies greater uncertainty among the sizes of the largest impactors. Nonetheless, based on that analysis, they wrote “Because the impact signature in the crater record in the inner Solar System is asteroidal, we conclude that either comets played a minor role or their impact record was erased by later-impacting asteroids.” Our unpublished analysis of the size frequency distribution indicated <15% of craters were produced by comets.

Attention then turned to the lunar regolith. Could fragments of the impactors have survived and be found as relics in ancient lunar soils? In ancient regolith samples, Joy et al. [8] detected relics with affinities to carbonaceous chondrites (CC) and properties distinct from comets. The lack of detectable comet fragments indicated <5 to 17% of the impactors could be comets, leading them to conclude “the impactor relics described here indicate that asteroids were the dominant objects hitting the Earth-Moon system at the end of the basin-forming epoch and that the flux of comets was small.” The study also showed the types of impactors diversified after the basin-forming epoch ended.

Those studies collectively suggested the relative contribution of asteroids and comets during the basin-forming epoch, or at least the latter part of the basin-forming epoch, was dominated by asteroids. Can we probe deeper in time?

An obvious target for that question is water, which had, in the midst of the studies above, been detected in lunar samples [9] and, thus, inferred to be present in the lunar interior. Water in asteroids and comets has sub-equal abundances and distinct isotopic compositions. Two questions followed. When did that water accrete to the Moon? What was the source of that wa-
Several teams addressed one or both of those questions. Saal et al. [10], Füri et al. [11], and Barnes et al. [12] determined that the water had affinities to asteroidal CC, although Greenwood et al. [13] preferred a cometary origin. Barnes et al. [12] went on to write “…we conclude that comets containing water enriched in deuterium contributed significantly <20% of the water in the Moon.” They also argued most of the water was delivered before the Moon’s crust solidified. If true, the asteroids delivered water to the Moon in its initial ~200 million years of evolution.

**Dynamical Insights:** The geological record of bombardment [7] has implications for the dynamical origin of the bombardment. Because the size frequency distribution of calculated projectile diameters matched that of main belt asteroids, the asteroid belt must have provided projectiles in a size-independent manner. That suggests gravitational resonances swept through the asteroid belt. That implied, in turn, that the orbits of Jupiter (and other giant planets) evolved.

In parallel with the geochemical, isotopic, and geological studies described above, several dynamical models for early Solar System evolution were being developed. Gomes et al. [14], for example, proposed an instability caused giant planets to migrate, which had the advantage of increasing the proportion of asteroids delivered to the inner Solar System over previous models [15]. The parameter space for those dynamical models continues to be explored (e.g., [16-18]), as discussed in separate papers at the conference.

**New Directions:** The work of [7] prompted a hypothesis: If resonances were sweeping inward, could we detect the early production of impactors from the outer belt, followed by impactors from the inner belt? Thus began an examination of the geochemical and isotopic fingerprints of impactors in lunar impact melts as a function of their age using new analytical techniques developed by Walker’s group [19-21] and others. Their data are broadly consistent with a diverse set of chondritic impactors and an additional contribution from a fractionated core-composition impactor, although Fischer-Gödde and Becker [22] suggest the diversity is, instead, a mixing trend between a CC component and a type IVA iron meteorite component. If the compositions of Liu et al. [21] reflect multiple impactors rather than mixing, then compositions change from CC affinities at 4.2 Ga to OC and EC affinities at 3.75 Ga, which might be reflecting sweeping of resonances as postulated by [7] from the outer to inner portions of the asteroid belt. In the midst of that sweep, impactors with iron meteorite affinities occur, which could have been scattered from the terrestrial zone and deposited in the midst of the asteroid belt before the resonances moved. Materials with geochemical and isotopic affinities to OC and EC, along with type IAB irons, are also important components in the local feeding zone of the accreting Earth [23].

**Conclusions:** There sometimes appears to be a tangle of evidence, in part because of the challenge to separate sources during accretion (both before and after the giant impact, both before and after core formation) and subsequent impacts (both before and after lunar crust formation). A comprehensive solution eludes us. While it will be productive to continue probing the existing collection of lunar samples and explore dynamical models that may have delivered the projectiles, it is clear that substantial, potentially transformative [4], progress can be made if/when we collect new samples from the Moon at specifically targeted sites and return them to Earth for detailed analyses. It will be enlightening, for example, to determine the age and nature of the projectiles that produced the oldest basin (South Pole-Aitken basin) and one of the youngest basins (Schrödinger and Orientale). As has been explored elsewhere (e.g., [24]), a mission to the Schrödinger basin has the potential to provide all of those answers.

**References:**

Testing Bombardment Models with Human-Assisted Robotic Missions and Well-Trained Astronauts on the Lunar Surface. David A. Kring1,2, 1Center for Lunar Science and Exploration, Lunar and Planetary Institute, Universities Space Research Association, 3600 Bay Area Blvd., Houston TX 77058 (kring@lpi.usra.edu), 2NASA Solar System Exploration Research Virtual Institute.

Introduction: The Apollo sample return missions revealed the Moon is a differentiated body that once harbored a magma ocean, that plagioclase crystallizing from that magma ocean buoyantly rose to form a solidified anorthositic crust, and that impacting objects, particularly early in Solar System history, dramatically modified that crust. Today we see the remnants of that period of bombardment in the form of immense basins, some of which are of order 1000 km in diameter.

The samples returned to Earth revealed a surprising pattern. A concentration of c. 3.9-4.0 Ga ages suggest there may have been an intense period of bombardment [1] or lunar cataclysm [2] that may be reflective of collisions throughout the entire inner Solar System [3,4]. Not only did the bombardment affect the geologic evolution of terrestrial planets, it may have also influenced the origin and evolution of life on the Earth, potentially on Mars, and potentially on water-rich bodies in the outer Solar System. Because the impact flux to the inner Solar System is both accessible and uniquely preserved on the Moon, additional samples to evaluate the impact flux are among the highest lunar science priorities [5].

Impact Cratering Science during the Apollo Era: The Apollo results emerged when the geologic community’s understanding of impact cratering processes was in its infancy. It was not understood, for example, the amount of target melting that could occur. Many photogeologically-observed features, such as melt flows in Tycho crater, were errantly interpreted to be post-impact volcanic products. We now understand that such deposits are likely a direct result of crustal melting produced by the intense kinetic energy of impacting objects.

Our interpretation of the samples was (and remains) challenging because they were collected around basins (Imbrium, Nectaris, and Serenitatis) that have been overprinted by younger impact events. Thus, when trying to evaluate the ages of complex breccias that may have incompletely reset radiometric systems and subsequently been overprinted by younger incompletely reset radiometric systems, interpretations of ages extracted from the samples can be difficult.

Collecting New Samples: Potential landing sites to test the lunar cataclysm hypothesis have been identified (e.g., [6]). Two of the most important sites are the Orientale and Schrödinger impact basins, the youngest and second youngest impact basins on the Moon. For that reason, the lithologies exposed there are nearly pristine. If collected, they will provide an opportunity to measure the age of each basin and, just as importantly, illuminate the nature of melt rocks and melt-bearing breccias produced by single basin-forming events. Once armed with that knowledge, we will be better able to assess the samples in the Apollo collection and samples collected from other, intermediate-age basins.

The second highest priority is to determine the age of the South Pole-Aitken (SPA) basin, the largest and oldest basin on the Moon. Missions to the SPA basin have been proposed (e.g., [7]). Potentially that age can also be determined with a mission to the Schrödinger basin, because it sits within the SPA basin and may have exposures of SPA impact melt within it [8].

Robotic and Human-assisted Robotic Missions: Landing sites, rover traverses, and sample stations have been identified within the Schrödinger basin for robotic mission that last from 14 days [9] to 3 years [10] in duration. These are compelling missions because they also address most of the other NRC objectives for lunar exploration [9,10] and several in situ resource utilization objectives. The shorter duration mission can be conducted entirely from Earth, assuming a communication relay to the farside exists, or with crew in the Orion vehicle. The longer duration mission can be conducted with crew in the Orion vehicle and within an orbiting Gateway [11].

Humans on the Lunar Surface: Landing sites, rover traverses, and sample locations for astronauts have been identified for simple sortie-type missions to the Schrödinger basin [12,13]. The Schrödinger basin is also an integral part of 5 human missions [11] recently outlined for the International Space Exploration Coordination Group, of which NASA is a member. In this scenario, two rovers are deployed at the first landing site. Crew then land, conduct up to a 42-day-long mission, collecting samples, before returning to Earth. The rovers are tele-robotically driven to the second landing site, where crew rendezvous with the vehicles. Crew conducts another 42-day-long mission, returning to Earth with a second suite of samples. The mission cadence is once per year and begins at the Malapert massif, followed by landings at Shackleton crater, the
Schrödinger basin, Antoniadi crater, and the center of the SPA basin.

Training: To complete those tasks, crew and the science staff supporting them in mission control will need to study analogue sites (Fig. 1) to learn about crater morphology, associated structural elements, the distribution of impact lithologies, and how to locate samples suitable for determining the ages of craters [14-16].

Astronauts and supporting science staff can also be taught to use complex craters and multi-ring basins as probes of the lunar interior. Normal faults in the modification zones of these craters expose subsurface lithologies and their stratigraphic relationships. Uplifted central peaks and peak rings expose even deeper levels in the Moon’s crust. Furthermore, clasts of subsurface lithologies are entrained in impact melt breccias deposited within the crater and beyond its rim. Thus, by combining observations of modification zones, central uplifts, and impact breccias, one can generate cross-sections of the lunar crust that may be kilometers to 10’s of kilometers deep. The volume of material beneath an impact site that is melted extends to an even deeper level than the material that is excavated. Thus, while collecting melt samples to determine the impact flux, crew will also be collecting samples of the lunar interior.

Large craters may have formidable crater walls, so some missions may be limited to the crater interior, while others may be limited to the crater ejecta blanket. Learning how to conduct radial sampling of an ejecta blanket to probe the subsurface stratigraphy exposed in the crater interior will be another key training objective at terrestrial craters.

In addition to basic geologic training, it will be essential to conduct mission-style simulations (Fig. 2). Good examples of the simulations were developed by the DRATS program, wherein crew in Lunar Electric Rovers conducted 3-, 14-, and 28-day long missions.

Figure 1. Kring training 2009 class of astronauts in Meteor Crater in a program that also involved studies in the nearby San Francisco Volcanic Field.

Figure 2. Mission simulations in the San Francisco Volcanic Field north of Flagstaff.

THE ROLE OF LARGE COLLISIONS IN FORMING EARLY COMPOSITIONAL HETEROGENEITIES ON MARS. S. Marchi¹, R. J. Walker², and R. M. Canup¹, ¹Southwest Research Institute, Boulder, CO, USA (contact: marchi@boulder.swri.edu), ²Department of Geology, University of MD, College Park, MD, USA.

Introduction: A common trait of the evolution of terrestrial planets (Earth and Mars, for which we have samples), the moon, and large planetesimals (such as Vesta, the parent body of the HED meteorites), is their protracted accretion period. Based on the concentration of highly siderophile elements (HSE, e.g. Pt and Au) in rocks derived from their crusts and mantles, it is estimated that these bodies have accreted ~ 0.1-2% of their final mass after the formation of their cores was concluded [1,2]. The roughly chondritic relative distribution of HSE has further suggested that this “late” accreted mass was delivered via collisions. For the Earth, HSE may have been delivered by large, differentiated planetesimals (>1000 km in diameter) [3]. Such large collisions could have resulted in substantial mantle compositional heterogeneities, potentially resulting in detectable W isotopic anomalies, such as those from Isua supracrustal rocks or Schapenburg komatiites [2].

In this work we investigate projectile mixing in large collisions on Mars with a suite of dedicated high resolution smoothed-particle hydrodynamics (SPH) simulations. In doing so, we study the potential for large collisions to generate isotopic anomalies in the martian mantle, and compare our results to HSE and 182W isotopic data from the shergottite-nakhlite-chassigny (SNC) meteorites.

Collisional mixing: To investigate mixing of projectile material into Mars, we performed SPH simulations with 0.5 to 1.2 x 10⁶ particles for two projectile masses, M = 0.003Mₘ and 0.03Mₘ, where Mₘ is Mars’ mass, with impact angles β = 0, 30, 45, and 60°, and impact velocities (v) of v/vₑ = 1.5, 2, 2.5, where vₑ is the system’s escape speed. These ranges of projectile mass and velocity encompass those inferred for the Borralis basin [4,5], and those seen in dynamical models of late accretion onto the terrestrial planets [6]. We consider projectiles with metallic cores comprising 30% of their mass, with cores resolved by 10³ to 10⁴ SPH particles. We use our SPH simulations to track the fate of the projectile’s core that contains its HSE budget (Fig. 1), and define three end states: merges with Mars’ core, remains suspended in the martian mantle, or ejected from the system. In this analysis, we closely follow the approach developed for the Earth [2].

Preliminary conclusions: We find that, for average impact angle and velocity conditions, about 10-20% of the projectile core is delivered to the martian mantle for M = 0.003Mₘ and 0.03Mₘ. This result implies that 1 to 4 collisions are capable of delivering the average concentration of HSE (Pt ~ 3 ppb) in SNC meteorites. Under these conditions, Mars is likely to develop large-scale impact-induced compositional heterogeneities. Within these domains, we predict a significant HSE concentration variation (Pt ~1-30 ppb), and well as W isotopic variations. To first order, these compositional variations are compatible with that observed in SNC meteorites. This reinforces the view that Mars’s mantle is heterogeneous, and raises a word of caution about using SNC data to derive global properties for that planet.

STUDY OF GLYCINE ADSORBED IN A CLAY UNDER GAMMA RADIATION IN A COMET CORES SIMULATION
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Introduction: John Oró, was the first scientific to proposed comets as key carriers of organic molecules to our early Earth [1]. Comets are formed mainly by ices (H2O, CO, CH3OH, CO2, etc.), organic matter, and silicates. It is proposed that these bodies could have brought water and organic material to the primitive Earth, this contribution might have been fundamental in the processes of chemical evolution (physical and chemical preamble prior to the appearance of life) [2].

Amino acids in a comet cores. In 2009, was announced that scientists had identified one of the fundamental building blocks of life in a comet for the first time: glycine [3]. Glycine is an essential molecular components of living organisms on Earth, which is commonly found in proteins.

Clays in a comet cores. There are clays in a comets cores [4]. John D. Bernal (1951) suggested, clays as concentrators of biological precursor molecules, as catalysis and clays might protect these molecules from high-energy radiation highlighting their relevance in the emergence of life, because these are on the list of minerals that would be expected to have been prevalent during the evolution of the early Earth, also in meteorites, and comets [5].

We present a detailed analysis of the study of glycine adsorbed in a motmorillonite under gamma radiation at a low temperature, simulating a comet core. The laboratory simulation consists of icy phases of prototype organic matter and a mineral in a physical mixture. This chemical system was irradiated with gamma radiation at 77 and 295 K. The icy phases are methanol, formic acid, and glycine in aqueous solution, in the presence of sodium montmorillonite as silicates surface.

We have been identified some radiolytic products of this mixture by Gas Chromatography (GC) and High Performance Liquid Chromatography-Electrospray Ionization-Mass Spectrometry (HPLC-ESI-MS). The protection role of the clay in the radiolysis of glycine was observed in this mixture. This result may be due to an energy transfer from the clay.

References:
**Introduction:** Crater density is the most common method of relative and absolute age determination on the Moon, providing a powerful tool for reconstructing lunar geologic history when combined with stratigraphic observations. The base assumptions for dating a surface using crater density are that (1) cratering is a random process and (2) the crater population accumulated over time (starting with no craters at time zero) is representative of the age of the unit [e.g., 1-4]. As such, significant effort has gone into understanding the factors influencing the formation and destruction of the crater population [e.g., 5-8].

Crater degradation is accelerated for craters formed on steep slopes due to downslope movement of materials (especially in thick regolith) and seismic shaking [5, 7-8]. Basilevsky [5] investigated degradation rates for craters ranging from a few meters to ~1–2 km in diameter, but the crater diameter for which slopes cease to significantly increase degradation rates was not well constrained. The morphology of small craters <1 km is controlled by target properties [9,10], making it difficult to predict the effect of slope alone on larger diameter craters (>1 km diameter). The effect of slope on the retention of craters >1 km, where gravity rather than target properties controls degradation and craters are less affected by regolith depth and seismic shaking [7-8], has not been quantified.

Prior to the arrival of the Lunar Reconnaissance Orbiter Camera (LROC) at the Moon, there was insufficient resolution in topography to investigate the effect of slope on craters >1 km in a robust manner. Building on a preliminary study [11], this work seeks to (1) determine if the trend of decreasing crater density with increasing slope observed by [5] holds true for craters >1 km and (2) quantify any effect of slope for craters of this size observable in overall crater retention.

**Study Area:** The Orientale basin constitutes a major stratigraphic marker, and its continuous ejecta, called the Hevelius Formation (HF), is areally extensive (>2 million km²). The hummocky terrain of the HF permits the study of the effect of slope on crater density. As a single geologic unit, age estimates should be consistent across the entire HF for craters >1 km; however, the large areal extent of the HF provides the opportunity to assess potential variation in absolute model ages (AMA) across a single geologic unit. Previous AMAs derived from counts on the HF returned a broad range of ages [3.64 Ga-3.8 Ga; e.g., 12-16].

**Data and Methods:** Starting with existing catalogs of all lunar craters 5-20 km [17] and >20 km [18], all craters with diameters >800 m were measured within the study area (Fig. 1). Craters were identified using the CraterTools extension for ArcGIS [19] from LROC Wide Angle Camera (WAC) mosaics (average 60° incidence) with opposite illumination thus minimizing lighting biases [20]. Secondary craters occurring in chains and clusters were excluded from the count area. The study area was defined using the HF boundaries as mapped by [21]. A slope map with a 3 km baseline (computed using a 3x3 kernel) was derived from the WAC GLD100 [22] topography, downsampled to 1 km to eliminate the contribution of slopes due to the craters themselves to the calculated slope values. Cumulative crater (all craters >1 km) density maps (e.g., Fig. 2) were created to identify areas of the highest and lowest density (HDZ, LDZ). Crater size–frequency distributions (CSFDs) were computed using pseudolog binning [23] and AMAs were derived for the HDZ and LDZ using Craterstats2 [24] and the chronology (CF) and production functions (PF) of [25].

Only craters >5 km are identifiable in the 3 km baseline slope map indicating that the 3 km smoothing...
is minimizes the contribution of the interior crater slopes for craters <5 km (9,356 craters in total for 1-5 km) and that the measured slope values are representative of the pre-existing terrain. Because the contribution from slopes on the crater interior and pre-existing slopes cannot be disentangled for craters that are visible in the slope map, craters >5 km were not used in the subsequent analysis of slope effects. The fractional crater density of craters up to 5 km in diameter was compared to estimates of the pre-existing slopes upon which the craters formed (Fig. 3).

Results: The HDZ, located at 14.2°S 241.4°E has an average slope value of 6.7°, comparable to previous estimates of average highland slopes [e.g., 26], and the LDZ, located at 37.5°S 263.0°E has an average slope of 9.1°. The HDZ yields an AMA of 3.7 Ga +/-0.03, consistent with the age expected for the Orientale basin from identical methods (e.g., 3.68 Ga [13,14]). As expected, the LDZ yields a younger AMA, in this case 3.5 Ga +0.08/-0.2. Based on these estimates, the ages could vary by 90 My up to 430 My (HDZ = 3.73 Ga, LDZ = 3.3 Ga) simply based on local slope. When the crater density is compared pixel-by-pixel to the underlying slopes, we see that steeper slopes display fewer small craters (Fig. 3), consistent with [5]. As the diameter increases, the distribution becomes wider, and its peak becomes lower in amplitude indicating that larger craters encompass a broader range of slope values and are less affected by steep slopes. This work is consistent with the following conclusions:

1. For craters 4–5 km in diameter, the effect of slope-induced degradation acceleration is minimized, and the shift in the slope distribution peak to shallow slopes for craters <5 km suggests that slopes still affect crater density up to at least 4 km in diameter.

2. Regions with lower average slope values yield higher AMAs than regions with higher average slopes, likely due to higher retention of countable craters.

3. Age estimates for the Orientale event can vary by up to 430 My, likely due to variations in topography and thus landform degradation acceleration. This effect requires verification over multiple baselines.


Fig. 2. Point density calculated in ArcGIS using a 25 km neighborhood to emphasize crater density variations for all diameters across the HF, red = high density, blue = low density.

Fig. 3. Crater density plotted against underlying slope.
NEW INSIGHTS INTO CRATER EQUILIBRIUM USING THE CRATERED TERRAIN EVOLUTION MODEL. D. A. Minton1, C. I. Fassett2, M. Hirabayashi1, 1Department of Earth, Atmospheric, and Planetary Sciences, Purdue University, West Lafayette, Indiana USA 47907 (daminton@purdue.edu), 2Marshall Space Flight Center, NASA Marshall Space Flight Center, 320 Sparkman Drive NW, Huntsville, AL 35805, 3Department of Aerospace Engineering, Auburn University, Auburn, Alabama USA 36849

Introduction: We have developed a new model for crater equilibrium for simple craters on the lunar mare. Using LROC observations of the Apollo 15 landing site as well as a numerical Monte Carlo code called the Cratered Terrain Evolution Model, we set constraints on the cratering processes that determine the diffusive erosion rate of lunar topography. Here we show how equilibrium is affected by the slope of the production function, the size of the area over which the new crater affects pre-existing topography, and the amount of degradation to the pre-existing landscape that each new crater generates.

Model for diffusive equilibrium of simple craters: Hirabayashi et al. [1] developed an analytical model for crater accumulation and degradation, which can be used to model craters in equilibrium. The crater production can be approximated as a power law, with its cumulative size frequency distribution (SFD) taking the form:

\[ n_{p>r} = X n_{p,1} r^{-\eta}, \]

where \( n_{p>r} \) is the cumulative number of produced craters per unit area, \( n_{p,1} \) is a coefficient that gives the cumulative number of craters larger than 1 m in radius per m\(^2\) of surface, and \( \eta \) is the log slope of the production distribution. The variable \( X \) is the same as that used in Hirabayashi et al. [1], and is a non-dimensional time factor used to scale the production function by the exposure age of the surface and \( n \) is the number density of craters in a size range. We use a similar model can define the accumulation and degradation of countable craters using a first-order linear differential equation:

\[ \frac{dn}{dr} = \frac{d\tilde{n}_p}{dr} - k \frac{dn}{dr}, \]

where \( dn \) is the differential number of countable craters, \( d\tilde{n}_p \) is the crater production rate, and \( k \) is a dimensionless degradation parameter that describes the fractional amount of degradation a crater experiences per unit time. In equilibrium, equation (2) is equal to 0.

The degradation parameter, \( k \), can be defined in terms topographic diffusion. In topographic diffusion, the classical diffusion model has been used subsequently to understand how landscapes on airless bodies, such as the Moon [2, 3] and Mercury [4], evolve under the erosive effects of impact cratering. The importance of diffusive downslope degradation by small impacts has also been recognized as being related to the observed empirical equilibrium of small craters [2, 3, 5-10]. As we develop our model it is important to distinguish between the two distinct roles that craters play (features vs. agents of degradation). To do this we adopt the notation system used by Hirabayashi et al. [1] in which we refer to countable craters that are being subjected to degradation have radius \( r \), while newly formed craters that contribute to degradation have the radius \( \tilde{r} \).

Any particular crater will play both roles in the evolution of the terrain.

In a surface that is undergoing classical downslope diffusion, topographic evolution takes the form

\[ \frac{\partial h}{\partial t} = \nabla^2 h. \tag{3} \]

A crater subjected to a diffusive process will degrade until it is no longer recognizable as a countable crater. Larger craters require more diffusive degradation to be rendered uncountable than smaller craters. We define the visibility function as the degradation state required to render a crater uncountable:

\[ K_p = K_{v,1} r^\gamma. \tag{4} \]

Constant degradation rate: If there is surface is subjected to a constant degradation rate, \( K_d \), then the cumulative size-frequency distribution of craters in equilibrium is:

\[ n_{eq,1} r^{-\beta} = K_{v,1} \frac{n_{p,1} \eta}{K_d} r^{-(\eta-\gamma)}. \tag{5} \]

In this model, the exponent (log slope) of the cumulative equilibrium SFD depends on the slope of the production function and the visibility function as \( \beta = \eta - \gamma \).

Crater size-dependent degradation: We now consider the case where each new crater contributes to the degradation of the old craters in a size-dependent way. We define the contribution to the degradation state of the surface by each new crater as the degradation function \( K_d(\tilde{r}) \).

We consider a simple model for a degradation function in which the amount of degradation each crater contributes is uniform over a circular region of area \( A_e = \pi \frac{\tilde{r}^2}{2} \). The amount of degradation is given by:
We find that the cumulative size-frequency distribution of craters in equilibrium for this model is:

\[ n_{eq,1}r^{-\beta} = \left( \frac{1}{\pi f_c^2} \right) \left( \frac{K_c}{K_{d,1}} \right)^{(\eta-2)/\psi} \left[ \frac{1 - (\eta - 2)/\psi}{\eta/(\eta - 2) - \gamma/\psi} \right] r^{-(2\xi + \eta)(\gamma - 1)} \]  

(7)

**Numerical simulations of equilibrium with CTEM:** The Cratered Terrain Evolution Model is a Monte Carlo landscape evolution code that models a three-dimensional surface that is subjected to impact cratering. It has previously been used to study the lunar highlands cratering record [10, 11] and impact transport of compositionally distinct surface materials [12]. Because it is a three-dimensional code, it can readily model topographic diffusion.

We implemented multiple degradation models into CTEM and confirmed our diffusion-based models for equilibrium.

**Figure 1. CTEM simulation of a constant degradation rate model.**

**Figure 2: CTEM simulation of a crater size-dependent degradation model.**

**References**

THE TIMELINE OF THE LUNAR BOMBARDMENT - REVISITED. A. Morbidelli1, D. Nesvorny2, V. Laurenc3, S. Marchi2, D.C. Rubie3, L. Elkins-Tanton4, M. Wieczork3 and S. Jacobson1,3,5 1Laboratoire Lagrange, Observatoire de la Côte d’Azur, Nice, France, 2Southwest Research Institute, Boulder, Co., 3Bayerisches Geoinstitut, Bayreuth, Germany, 4School of Earth and Space exploration, University of Arizona, Phoenix, Az, 5Department Of Earth And Planetary Sciences, Northwestern University

Introduction: The contrasting scenarios of terminal cataclysm and accretion tail have been proposed to explain the intense bombardment that the Moon suffered ~3.9Ga ago. Support to the cataclysmic hypothesis was provided in [1], under the traditional assumption that Highly Siderophile Elements (HSE) in the lunar mantle/crust track the amount of chondritic material that hit the Moon since its formation. However, a new result [2] shows that HSE sequestration into the core of a planetary body continues until 50-70% of the mantle has crystallized, due to the pervasive exsolution and segregation of FeS. If this is true also for the Moon, and the crystallization of its magma ocean occurred late [3], the lunar HSEs would severely underestimate the total mass accreted by our satellite since its formation.

HSE sequestration into the lunar core during LMO crystallization: In [4] we have re-examined the results of [2] for the case of the Moon. The lunar magma ocean (LMO) crystallizes bottom-up. Sulfur (S) remains in the liquid and therefore its concentration increases as the fraction of residual melt decreases. We have evaluated the Sulfur Concentration at Sulfide Saturation (SCSS) following [5], accounting for the decreasing pressure at the basis of the retracting magma ocean and its evolution in composition [3]. We show that eventually SCSS is exceeded and S exolves as FeS liquid. The HSEs partition strongly into FeS because they are chalcophile. The concentration of ilmenite-rich and FeS-rich high-density cumulates at the top of the lunar mantle then triggers a global mantle overturn. According to several models, e.g. [6] the heavy material can reach the core-mantle boundary. Some KREEP would also have been dragged to the basis of the mantle by this process. If the quantity of KREEP was large enough, this material could melt, thus accounting for the high-titanium basalts. Melting would decouple FeS from ilmenite and allow FeS to migrate into the core, with the associated HSEs. This scenario explains the presence of S in the lunar core, deduced by seismological observations, e.g. [7]. Moreover, if MO crystallization occurred on the Moon significantly later than on the Earth, the strong imbalance between the current HSEs concentrations in the Lunar and terrestrial mantles [8] would be the straightforward consequence of the two different HSE retention ages.

Implications for the Lunar bombardment: If the Lunar mantle HSEs track only the amount of material accreted since LMO crystallization and the latter happened 100-200My [3] after lunar formation, the population of leftover planetesimals that impacted the Moon throughout its early history was much higher than estimated up to now. In [4] we show that, in this case, the accretion-tail scenario can explain the lunar crater record. In this new scenario the Moon accretes in total \(5\times10^{5}\) Earth masses, but retains HSEs only starting from 4.35Gy ago (fig. 1).

ANCIENT BURIED BASINS IN OCEANUS PROCELLARUM FROM THE FIRST BILLION YEARS. G. A. Neumann1, A. N. Deutsch2, and J. W. Head2,1Solar System Exploration Division, Code 698, NASA Goddard Space Flight Center, Greenbelt, MD 20771, USA (gregory.a.neumann@nasa.gov), 2Department of Earth, Environmental and Planetary Sciences, Brown University, Providence, RI 02912, USA.

**Introduction:** Ancient, degraded basins have been identified from GRAIL Bouguer anomalies [1] and precise lunar topography. Better gravity and topography allow the existence of proposed basins to be confirmed or plausibly rejected on the basis of an anomaly produced primarily by crustal thinning and subsequent relaxation. The known relations among the diameters of basin rings and of the zone of thinning producing central Bouguer anomalies derived from well-preserved basins [2] allow inference of the approximate size of the main rim, even in some cases where no topographic rim is preserved. The impact parameters can thereby be compared in relation to the target properties and the flux of impactors.

Circular positive Bouguer gravity anomalies (PBGAs) and quasi-circular mass anomalies without discernible topographic signature can also be vestiges of impacts completely buried by subsequent mare volcanism [3]. However, the source of PBGAs is not necessarily due to impacts, given the likelihood of intrusive volcanism [4] or thorium-bearing silicic volcanism [5] giving rise to some of the anomalous gravitational signature in e.g., Compton-Belkovich or Aristarchus.

On the lunar nearside, especially in the Procellarum KREEP terrain, there are cryptic density anomalies that would increase the overall inventory of ancient impacts if their dimensions could be uniquely described. The mass anomalies proposed to be buried craters are only known by the extent of their prominent gravity gradient signatures [3] but are smaller than the 300+ km diameters traditionally ascribed to impact basins. We have focused on four PBGAs (Fig. 1) that are important in understanding the impact and volcanic/plutonic history of the Moon, in a region of elevated temperatures due to the Procellarum KREEP Terrane [6]. Analysis of the size of craters associated with PBGAs in anomalously thinner crust and elevated temperatures can affect the assessment of the early bombardment of the Moon since these predate the dominant 3.8–2.5 Ga pulse of mare volcanism. Forward modeling of the Northern Flamsteed PBGA, for example, incorporating the known depths of fresh lunar craters in mare regions [7], strongly suggests that mantle uplift from impact rebound or intrusive dikes controlled by impacts provides the bulk of the gravitational signal. This anomaly was earlier identified as quasi-circular crustal thickness anomaly CTA-24 [8], and was interpreted as a likely impact basin with a main ring diameter of 323 km. While our modeling based on much higher resolution data does not support quite as large a buried structure, the existence of additional basins in Oceanus Procellarum has long been proposed and may yet be determined from gravitational modeling.

![Fig. 1](https://example.com/fig1.png) Four 80–120 km diameter positive Bouguer gravity anomalies in Oceanus Procellarum. Modeling suggests they may arise from larger impacts in thin (13-16 km) crust.

**Conclusions:** The detection of additional buried basins on the lunar nearside may help reconcile the apparent difference in size-frequency distribution of impacts between the two hemispheres [1]. The suggested increment in the range of 150–300 km in diameter may remedy some of the discrepancy between the main asteroid belt population statistics and the lunar record, but does not address the implied deficit of very large basins [9].

EARLY LUNAR CRUST HEALED ITSELF AFTER IMPACTS PUNCTURED HOLES. V. Perera¹ and A. P. Jackson²,³. ¹Applied Physics Laboratory, Johns Hopkins University, Laurel, MD, USA. (viranga.perera@jhuapl.edu), ²Centre for Planetary Sciences, University of Toronto, Toronto, ON, Canada. and ³School of Earth and Space Exploration, Arizona State University, Tempe, AZ, USA.

Introduction: The early Moon likely had a unique rheological structure where an anorthositic crust floated atop a Lunar Magma Ocean (LMO) [e.g., 1]. The crust would have acted as a thermal blanket and slowed the cooling of the Moon, prolonging what would have taken $\sim 10^5$ yr to $\sim 10^7$ yr [2]. Recent work has shown that this tranquil cooling of the Moon would have been interrupted by a prolonged bombardment from debris released during the original Moon-forming giant impact [3]. Reimpacting debris may have either prolonged or expedited the cooling of the Moon depending on the degree to which the bombardment punctured holes into the lunar crust and delivered thermal energy [4]. While both the initial mass of the reimpacting debris population and the mass accretion rate as a function of time can be estimated based on Smoothed Particle Hydrodynamics (SPH) simulations [e.g., 5] and $N$-body simulations [3] respectively, the impact process itself is complex and requires hydrocode modeling to determine the consequence to the crust. In ongoing work using the iSALE hydrocode [6,7,8], we show the conditions under which impacts will puncture through the crust, expose hot magma, and increase the heat flux [9]. Yet, that increased heat flux is temporary since holes will be refilled by newly formed anorthosites. Notably, impact hydrocodes are generally not able to model the long term thermal evolution of a system after an impact. As such here we focus on the thermal evolution of holes generated in the early lunar crust to determine the implications for the overall cooling of the Moon.

Methods: We developed a two-dimensional thermal evolution code called iFill (impact Filling). iFill is similar to iMagma from [4]; however, while iMagma is a one-dimensional thermal evolution code that evolves the whole LMO, iFill is a two-dimensional thermal code that focuses on the local area of an impact site. iFill incrementally solidifies the local LMO and adds that material to both the solid mantle below and the flotation crust above. This models the fractional crystallization process after 80% of the initial LMO has crystallized, where denser material fall to the interior while less dense material float to the surface [e.g., 2]. Newly formed solids can either be placed at random horizontal locations or, more realistically, preferentially at horizontal locations with higher thermal flux.

iFill first reads in a two-dimensional iSALE material output that identifies the locations of the crust and magma material. We then start with a LMO depth that corresponds to the crustal thickness in the iSALE output. For instance, it is expected that when the lunar crust was $\sim 10$ km thick, the LMO depth was $\sim 100$ km. iFill then iteratively solidifies the LMO while calculating the time required to expel the heat of fusion and the energy released by secular cooling. At each step, the crustal thickness, conductive flux, and the surface temperature at each horizontal location is updated.

Similar to [4], we use the solidus temperature equation from [2] to estimate the temperature at the LMO-solid mantle boundary. Thus, we calculate the temperature within the LMO by following the adiabat from the LMO-solid mantle boundary with the assumption that the LMO is convecting through out the solidification process. The surface temperature is calculated self-consistently by equating the conductive flux through the crust to the radiative flux from the surface.

We set two stopping criteria for the iterations. First, similar to both [2] and [4], the iterations stop when only 1% of the original LMO remains. Second, the iterations stop if the LMO-solid mantle boundary and the crust-LMO boundary intersect each other. The second condition usually occurs when the remaining LMO is $\sim 1\%$.

Here we use the output of one of our iSALE simulations to demonstrate iFill. The iSALE simulation input parameters include an impactor diameter of 10 km, impacting speed of 4 km/s and a 10 km thick crust. In this simulation the impactor punctures through the crust, but crustal material resettles into the impact site leaving the crust thinned rather than directly exposing magma at the surface. Crustal thickness after the impact near the impact site is $\sim 3-5$ km.

Results: The top left plot in Figure 1 shows that the portion of the crust that has been thinned by an impact produces 5 times the thermal flux as compared to the unperturbed crust far from the impact site ($\sim 2.0$ vs. $\sim 0.4$ W/m²). Given the assumption that the rate of anorthosite formation under a given location is directly proportional to the thermal flux, we find that in $\sim 1$ Myr crust near the impact site thickens sufficiently so that fluxes near and far from the impact site are roughly equalized. This hole thus has an effective lifetime of
Figure 1: An example of a two-dimensional iFill thermal model. On the left is the half-space cross section of an iSALE output (bottom) and the corresponding flux and crustal thickness profiles (top). On the right is the thermally evolved iFill output showing the crust (in white) and the solid mantle (in dark cyan) have grown thicker while the quantity of magma (in red) has decreased. Lighter colors in the top panel shows the initial flux and crustal thickness profiles as references. Here solids are weighted towards horizontal positions with a higher thermal flux. The horizontally-averaged self-consistent surface temperature and the average temperature of the magma near the surface and the bottom are shown as $T_{\text{surf}}$ and $T_{\text{magma}}$ respectively.

Around 1 Myr. Run to completion, solidification of the local LMO takes $\sim 30$ Myr. This is similar to the solidification time of the global LMO with the effect of reimpacting debris removed [4] since this is equivalent to modelling the cooling of the global LMO with only a single hole.

Alongside the hole lifetime of around 1 Myr, it is notable that when run to completion (right hand panel of Figure 1) the increased cooling flux and magma crystallization rate beneath the hole location leads to a bulge in the solid mantle due to the correspondingly increased deposition rate. This would force the final LMO crystallization products, presumably including the ur-KREEP material, away from the hole site. While more work needs to be done examining whether this would be preserved through convection in the LMO and subsequently the solid mantle, it is possible that these deviations in the mantle might provide a way of identifying some of the final hole locations, despite the lack of surface topography.

CRATERING AND PENETRATION OF THE EARLY LUNAR CRUST. V. Perera¹, A. P. Jackson²,³, L. T. Elkins-Tanton⁴, E. Asphaug⁴ and T. S. J. Gabriel⁵, ³

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Introduction: Giant impacts are expected at the late stages of planet formation [e.g., 1]. There is substantial evidence for the occurrence of such impacts in our own solar system, in particular our Moon is believed to have originated in such an event [2,3]. Recent work has shown that typically at least a few percent of the impacting mass will have sufficient speed to escape the system [e.g., 4]. For the Moon-forming impact, at least $10^{23}$ kg ($\sim$ 1.3 lunar masses) of debris is estimated to have escaped onto heliocentric orbits, a substantial fraction of which subsequently reimpacted the Earth and the Moon over a period of $\sim$100 Myr [5]. At the same time the Moon would have been cooling from a mostly molten state, having formed from a high-energy impact event. Without impact bombardment, the Moon is thought to have solidified fairly slowly over around 10 Myr due to the early lunar crust functioning as a thermal blanket [e.g., 6]. Reimpacting debris however may have had a large influence on the cooling rate. They could have punctured the early crust, increased the thermal heat flux and thus expedited cooling. At the same time reimpacting debris would have carried substantial kinetic energy, which may have resulted in heating of the Moon. As such recent work found that reimpacting debris may have temporally altered the early lunar thermal evolution [7].

The initial mass of the reimpacting debris population and the mass accretion rate as a function of time are estimated based on Smoothed Particle Hydrodynamics (SPH) simulations [e.g., 4] and N-body simulations [5] respectively. Using those results recent work utilized a simple prescription to convert the mass accretion rate into an area of holes produced on the crust [7]. This simplification was used since the degree to which holes are produced depends on a number of parameters pertaining to both the impactor (e.g., size and velocity) and the crust (e.g., thickness and mechanical strength). Here we seek to improve the previous work by modeling the impact process using a hydrocode. This will help constrain the conditions under which holes are produced by reimpacting debris and in turn improve our understanding of the overall thermal evolution of the Moon. Our work can also provide insights for other bodies on which impacts strike a solid surface underlain by liquid, such as the icy moons of the giant planets [e.g., 8].

Methods: Here we use the iSALE hydrocode, which is tailored for impact simulations and has been tested against laboratory experiments [9,10,11]. We start each simulation with an impactor just above the surface of a crust, which itself is atop a liquid layer to represent molten magma. Since ANEOS equations of state for lunar crustal and mantle material are not available, we represent the lunar crust with ANEOS granite (e.g., [12]) and the lunar mantle (along with impactors) with ANEOS dunite (e.g., [13]). This choice of materials has been used in previous studies, and essentially here satisfies two important conditions: that the crust has a lower density than the magma to ensure stable buoyancy, and that the crustal material has a higher melting temperature than magma beneath.

We have run over 150 simulations varying the impactor diameter, impact velocity and crustal thickness.
We simulate impactors with diameters ranging from 100 m to 30 km and crustal thicknesses ranging from 1 to 40 km. Since the debris originates from near-Earth orbits impacts are concentrated at low velocities, and so we focus on impact velocities from 3 to 9 km/s with our highest impact velocities being 15 km/s. As we are primarily interested in the penetration of the crust and the boundary between cratering and penetration, we do not simulate the smallest impactors striking the thinnest crust.

**Results:** In Figure 1 we illustrate the four types of cratering outcomes identified by our simulations: (1). ‘Classical cratering’ where fracturing is localized in a hemispherical volume, (2). ‘Full depth fracturing’ where the crust is fractured through to its base and as a result the fracture volume is approximately cylindrical, (3). ‘Partial penetration’ where the impactor penetrates through the crust but a significant amount of crustal material flows back to cover the hole and (4). ‘Complete penetration’ where the magma remains directly exposed after the crust has relaxed (see Figure 2 for an example iSALE simulation). Figure 3 shows that impactors are able to penetrate the crust when their kinetic energy is sufficiently high to do the necessary mechanical work in removing crustal material. Since some of the kinetic energy goes into breaking and heating crustal material, more kinetic energy is required for penetration than merely work required to move crustal material.

**Conclusions:** We find that impact kinetic energy is a very good predictor of the size of hole produced by an impact, with the hole opening efficiency being around 5% by our measure of the work done in opening a hole. This contrasts with the case for craters where scaling relations are more complex. Hole production is much more closely tied to the transient cavity, whose size is also known to correlate well with impact energy. The onset of partial penetration occurs when the transient cavity breaches the base of the crust. This also leads to a minimum hole diameter that is approximately equal to the crust depth.

**References:**  
CHARACTERIZATION OF PROPOSED IMPACT MELT FACIES OF THE MOON’S CRISIUM BASIN.
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Introduction: Understanding the impact history of the Moon has important consequences for Earth’s earliest history, including the time around the emergence of life ~4 Ga [e.g., 1], as well as for the overall bombardment history of the inner Solar System. The Nectaris and Crisium basins, particularly, are important anchor points in understanding this history. Spudis and Sliz [2] mapped 10 locations of putative impact melt outcrops (high-standing kipukas embayed by mare basalt flows) around the Mare Crisium’s periphery. If these outcrops are impact melt from Crisium, they would provide sampling opportunities to radiometrically date and derive an absolute age for the Crisium forming impact event, allowing better calibration of crater-derived ages [3].

Here, we present high-resolution mapping (~1:50,000) of the proposed Crisium impact melt sites identified by Spudis and Sliz [2], identify lithologies, describe relevant regional geology, and further assess their likely origin(s). Specifically, we hypothesize that if these kipukas are outcrops of Crisium impact melt, their lithologies should reflect a more feldspathic composition than the embaying mare basalts since the target rock was likely highlands material.

Methods: Our data sets include global Lunar Reconnaissance Orbiter Camera (LROC) Wide Angle Camera (WAC) low- and high-incidence angle monochrome basemaps (100 m/px); low- and high-incidence angle LROC Narrow Angle Camera (NAC) images (~1 m/px); the combined 59 m/px global Digital Elevation Model (DEM) from the Lunar Orbiter Laser Altimeter (LOLA) and Kaguya Terrain Camera (TC) [4]; Moon Mineralogy Mapper (M3) spectra; and Diviner Lunar Radiometer data. Using NAC images, the DEM, and the derived slope information, we mapped the boundaries of kipukas in Mare Crisium at a scale of 1:50,000 in ArcMap, with finer mapping (1:15,000) provided in places of low topographic contrast. We measured fracture dimensions using the LOLA/Kaguya DEM.

Results and Discussion: Kipuka Mapping. Examples of our kipuka mapping, refined from the work of Spudis and Sliz, are shown in Figure 1. As described by Spudis and Sliz [2], the kipukas display a range of morphologies, from domed and fractured to more subdued and hilly. Below, we focus on the Western Crisium Kipuka (WCK; box A, Figure 1) and northern Crisium kipukas (box B).

Fracture Morphometry. The WCK exhibits fractures reminiscent of features in the Maunder Formation [2], the melt sheet of Orientale Basin [5]. However, fracturing is not unique to melt sheets, and floor fractured craters (FFCs) can display fractured morphologies [e.g., 6], hypothesized to originate in response to subcropping dikes, sills, and laccoliths [e.g., 7] that cause doming and fracturing of the crater floor. Photic measurements of the WCK fractures, and fractures in two FFCs showed identical slopes of around 15°. In contrast, the Orientale and Imbrium fractured melt sheets have higher slopes (~32-39°), and often beyond the ~33-34° angle of repose for many granular media. The higher fracture slopes in melt sheets may be consistent with formation in response to normal faulting [8,9], or could simply reflect fractures that are younger and less degraded than those observed at Crisium.

Composition: Reflectance. We examined high-resolution NAC images with low incidence angles to evaluate the reflectance properties of the WCK and northern kipukas. Spudis and Sliz [2] report that the kipukas, especially the WCK, have a depressed FeO content relative to the rest of Mare Crisium basalts (~8.3 vs. >15 wt% for the WCK), interpreted as evidence favoring a Crisium impact melt interpretation. Photometrically normalized high-sun imaging shows the WCK to be overprinted by highland ejecta from Proclus Crater to the west, suggesting the depressed FeO values could be, at least in part, a result of Proclus contamination (Figure 3). Impact craters superposed on both the WCK and surroundings also have low-reflectance ejecta (“dark halo craters”; Box A in Figure 1; Figure 2), indicating a likely excavation of more mafic material from beneath Proclus ejecta and/or from below a thin surface unit native to the WCK (Figure 2). Reflectance values on the low-reflectance ejecta are 0.08, similar to surrounding mare, compared to >0.1 for the rest of the WCK.

Figure 1. Refined mapping of the Crisium kipukas outlined in red. Part of Box A is shown in Figure 2.
**Composition: Spectra.** The western and northern kipukas stands out from surrounding mare basalts [2] in terms of both the abundance of pyroxene and pyroxene composition as inferred from 1 and 2 µm absorption features observed in M' data. Both regions are more mafic than local massifs but more feldspathic than mare basalt with approximately a noritic anorthosite composition, which is consistent with what is expected for Crisium impact melt. In the case of the WCK, these spectra do not appear to show a large influence from Proclus ejecta. A ~3.5 km D diameter crater superposed on the WCK’s northeast edge has a reflectance spectrum indicating a pyroxene composition similar to the noritic central peak of nearby Yerkes Crater, but is less mafic with a lower abundance of pyroxene. The most “pristine” WCK materials are probably those associated with the 3.5 km D crater: the ejecta are similar to non-mare materials observed in surrounding massifs and craters.

**Composition: Thermal IR.** We have begun to investigate Diviner Radiometer 8 µm data to locate the Christensen Feature position to further discriminate lithologies in the manner of Greenhagen et al. [10].

**Discussion and Conclusions:** The morphology and noritic anorthosite composition of the WCK and northern kipukas are consistent with an impact melt origin, concordant with the work of Spudis and Sliz [2]. At the WCK, we propose an impact melt origin followed by intrusive magmatic uplift in a manner similar to FFCs [6,7], embayment by mare basalt, overprinting by Proclus ejecta, and finally excavation of intrusive material by small impacts. The northern kipukas appear to also be Crisium impact melt but with only minor contamination via impact gardening.

It is unclear whether Luna or Apollo (A17) samples contain Crisium material [3]; therefore, sample return or in situ radiometric dating of kipuka material will be needed to definitively determine the basin’s age. This underscores the need for a rigorous lunar surface exploration program to tightly constrain the impact history of the Moon, and, by extension, that of Earth. A future lander could be sent to the Crisium kipukas to perform in situ age dating [1] or return samples; geologic field and lab work could provide further context and understanding of their origin.

![Figure 2. 1:50,000 mapping of the Western Crisium Kipuka (WCK; boxed as “A” in Fig. 1). Embayment and other topography contrast defines the mapped contact with a ±100 m uncertainty buffer. Low reflectance ejecta craters (LRECs, green +), possibly indicate excavated intrusive mafic material. INSET: Zoom of the northernmost LRECs shows a strong reflectance contrast between surface/subsurface material. Base image: NASA/GSFC/Arizona State University.](image1)

![Figure 3. High-Sun photometrically normalized WAC mosaic showing highlands ejecta from Proclus Crater overprinting the WCK (centered at 15.0°N, 50.3°E) and other western kipukas, outlined in red. Base image: NASA/GSFC/Arizona State University.](image2)

**References:**

**Introduction:** Similar to the H- and LL-chondrites [1–3], L-chondrites record a long and complex history of accretion, differentiation, and bombardment in the inner Solar System. This history begins with the formation of the L-chondrite parent planetesimal originally some 260–280 km in diameter [4], overlapping in time with early large impact events between ~4.52 Ga and ~4.40 Ga [3,5], and followed by impacts between ~3.80–3.48 Ma [3] and around ~800 Ma [6]. Many L-chondrites show evidence for a major impact-induced disruption event in the asteroid belt at ~470 Ma [7,8], after which asteroid fragments were sent into Earth-crossing orbits, producing L-chondritic meteorite falls for at least ~5 Myr during the Ordovician [9]. Additional impact events on L-chondrite asteroids occurred post-disruption, between ~350 Ma and ~50 Ma [3].

An example of a prehistoric L-chondrite fall is presented by the Pleistocene (~15 kyr-old) ‘fossil’ Gold Basin strewn field in the eastern Mojave desert of NW Arizona [10], only ~250 km WNW of Flagstaff and the 2018 Bombardment conference of the The First Billion Years series. The strewn field, originally discovered by the late Professor Jim Kriegh and until recently thought to stretch over ~22 km from North to South (~225 km²), consists of several thousand fluvially and alluvially reworked stones [10] from an L4–6 chondrite breccia with at least one igneous clast [10,11]. The strewn field, thus, provides a unique reservoir of meteorite samples that can be used to study the record of accretion, differentiation, and impact events in the inner Solar System.

In this petrologic study, we provide evidence for a significantly larger Gold Basin strewn field that extends well into southeastern Nevada north of the Colorado River, thereby doubling the size of the strewn field. Implications for atmospheric and physical constraints on the Gold Basin meteorite fall, presumably an airburst explosion similar to the LL-chondritic Chelyabinsk event in Russia in 2015 [12], will be briefly discussed.

**Samples and Analytical Methods:** The present study would not have been possible without the dedicated field work and passion of meteorite hunters Joe Franske and Larry Atkins, who discovered and collected meteorites in SE Nevada and provided samples for analysis, along with locations of meteorite finds. Polished thin-sections were made from three randomly chosen individual ordinary chondrite stones. The thin-section samples, DKLPI–205, DKLPI–206 (Fig. 1A), and DKLPI–207, were analyzed using optical microscopy at the LPI and a CAMECA SX 100 electron microprobe at the NASA Johnson Space Center, Houston.

**Results:** Optical microscopy reveals that all three chondrite samples contain well-defined chondrules and part of olivine-phyric angular clast (upper left) in breccia (DKLPI–205; plane-polarized light).

**Sample DKLPI–205:** This chondrite sample contains a light olivine-phyric angular clast ~7 mm in size (Fig. 1B) and sub-mm-wide Fe-oxide alteration veins. Olivine is Fa24–27, predominantly Fa24–25 (n=13), with FeO/MnO (wt%)=43–50; Fe/Mn [AFU]=43–50. Low-Ca pyroxene is Wo1.1–2.5En74–79Fs24–20, predominantly Wo1.1En77–79Fs21–20 (n=12), with FeO/MnO=24–31; Fe/Mn=24–32. Kamacite has ~6.5 wt% Ni and 0.77–0.84 wt% Co (n=4). Sulfide is troilite (n=2).

**Sample DKLPI–206:** Olivine is Fa24–29, predominantly Fa24–25 (n=12), with FeO/MnO=43–53; Fe/Mn=43–52 (one Fa29 grain has FeO/MnO=56). Low-Ca pyroxene is Wo1.0En78–80Fs21–19. Low-Ca pyroxene is Wo1.0En78–80Fs21–19, predominantly Wo1.0En78–79Fs21–20 (n=8), with FeO/MnO=25–29; Fe/Mn=24–29. High-Ca pyroxene is Wo34–46 En46–54Fs19–21 (n=2), with FeO/MnO=22–42; Fe/Mn=21–44. Kamacite has ~6.4 wt% Ni and 0.77–0.84 wt% Co (n=7). A small grain of taenite analyzed has ~40 wt% Ni and 0.16 wt% Co. Sulfide is troilite (n=5).

**Sample DKLPI–207:** This sample contains some thin Fe-oxide alteration veins. Olivine is Fa24–26, predominantly Fa24–25 (n=10), with FeO/MnO=43–53; Fe/Mn=42–51. Low-Ca pyroxene is Wo1.2En78–80 Fs21–20, predominantly Wo1.2En78–79Fs21–20 (n=7), with
FeO/MnO = 26–33; Fe/Mn = 24–32. Kamacite has \( \sim 6.6 \) wt% Ni and 0.73–0.80 wt% Co (\( n = 5 \)). Taenite has \( \sim 35–37 \) wt% Ni and 0.20–0.23 wt% Co (\( n = 3 \)). Sulfide is troilite (\( n = 3 \)).

**Interpretation and Discussion:** The petrologic characteristics and geochemical composition of olivine and pyroxene (including their Fe/Mn) and Fe,Ni-metal of samples DKLPI–205, –206, and –207 (Fig. 1) indicate these samples are L4 chondrites. Within analytical uncertainty and natural variability, they are identical to those of stones previously recovered from the Gold Basin L4–6 chondrite breccia strewn field in NW Arizona [10,11], some 20 km farther south. According to the entry for Gold Basin in the *Meteoritical Bulletin* [15], olivine is Fas4, low-Ca pyroxene is Wo1 Fs20, and kamacite has 0.72±0.09 wt% Co [10,15]. Two additional fragments of the Nevada stones collected by L. Atkins, analyzed at UCLA, suggested those fragments are L6 chondrites (S4, W1) with olivine Fas3,9u0.2 (\( n = 15 \)) and low-Ca pyroxene Wo1,6u0.2 (\( n = 12 \)) [16]. We note that individual stones within the strewn field may show variations in their petrologic type (typically ranging from L4 to L6) and their shock inventory, and as expected for the fall of a chondritic breccia in a rugged (and, during Pleistocene time, non-arid) paleolandscape, their degree of weathering is also variable (W1–W4) [10].

Based on the striking petrologic similarity between the Gold Basin meteorites from Arizona and the stones recovered from SE Nevada, all of those meteorite finds may be part of the same meteorite strewn field. If true, this would make the Gold Basin strewn field one of the largest meteorite strewn fields on Earth behind Chelyabinsk [12]. The new Nevada finds occur ~16–27 km north of the northern margin of the previously mapped Gold Basin samples (Fig. 2) [10] and are, thus, within the 25 km limit usually required to assign a separate meteorite name. To test the pairing of the Nevada stones with Gold Basin, one should determine if their cosmogenic age is the same as that of Gold Basin [11].

Smaller stones in the southern portion of the strewn field in Arizona, with masses ≤1.5 kg [10], and significantly larger masses of up to 15 kg found in SE Nevada [16], suggest an oblique trajectory of the incoming meteoroid from the South to the North, in an airburst event that was probably similar to Chelyabinsk; however, the Gold Basin meteoroid, some 6–8 m in pre-atmospheric diameter [11], would have been less than half the diameter of the Chelyabinsk meteoroid [12]. Because all of the previously analyzed Gold Basin L-chondrites from Arizona were derived from a pre-atmospheric depth of ≤2 m within the meteoroid as constrained by cosmogenic nuclides [11], we hypothesize that the largest fragments of the fall, found recently farther north in Nevada, might carry a cosmogenic nuclide signature suggestive of greater shielding, representing portions of the ‘missing’ core of the Gold Basin meteoroid [11].

The new estimate of the extent of the Gold Basin strewn field and mass distribution therein can be used to help reconstruct the impact event from atmospheric entry to disintegration of the asteroid, the terminal burst, and the fall of decelerated meteorite fragments that produced the strewn field ~15 kyr ago.


Fig. 2: The Gold Basin, AZ, strewn field as previously mapped (overlay map with small dark spots for individual finds [10]) and northern portion of the strewn field in SE Nevada (bold spots) recently discovered by Larry Atkins [16]. Meteorites from both areas are petrologically indistinguishable L4–6 chondrite breccias.
Archean spherule layers: Windows into the early meteorite bombardment of the Earth

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Introduction: Impacts by large extraterrestrial object have undoubtedly occurred on Earth during the extended post-accretion bombardment phase, but the evidence of these early impact events has long since been obliterated. Discrete beds of post-Hadean spherule deposits, most likely representing distal impact ejecta, provide the only accessible traces of the early Archean impact record. Such spherules can be preserved over extremely long periods of time and are rarely observed associated with post-Archean impact events. The oldest spherule layers identified so far range in age from 3.4 to 3.2 Ga and are found in the Pilbara craton in Australia and the Barberton Greenstone Belt (BGB) in South Africa. These spherule layers form excellent time-stratigraphic markers and some of them are probably correlated [1]. In some of these spherule beds, the contents of the highly siderophile element (HSE) are anomalously high, up to chondritic or even slightly suprachondritic concentrations, and have previously been interpreted as a result of secondary enrichments [2, 3]. However, chondritic Cr isotope signatures in most of the analyzed spherule layers provided convincing evidence for an impact origin of these layers [4, 5]. Four, possibly up to eight [6-8], such Archean spherule layers were identified so far. The corresponding impactor sizes have been speculated to reach of up to ~60 km in diameter [9]. Such impacts would release huge amounts of energy and would have been capable to cause major changes in the Earth’s atmosphere, hydrosphere, and (if already existing) biosphere, maybe also initiating large-scale tectonic rearrangements and certainly causing huge tsunamis. Recently up to 21 new spherule layers in two drill cores from the Barberton Greenstone Belt (Barb5 and CT3) were discovered. They represent an unknown, but smaller number of impact events and provide excellent opportunities to gain more insights into the Archean impact record. Here we review the evidence for chondritic impactors in some of these newly discovered impact spherule layers [10-12] using highly siderophile element (HSE) abundances and 187Os/188Os isotope signatures.

Results and Discussion: Chondrite-like HSE and 187Os signatures in Barb-5 spherule layers [11], contrast up to four times chondritic peak HSE concentrations in CT3 spherules [12]. Initial (back-calculated to an age of 3.2 Ga) 187Os/188Os ratios always plot above or on the chondritic 187Os isotope evolution line for spherule layers from both cores. Based on in-situ HSE measurements of Ni-rich chromium spinels, which are the suspected carrier phases of the HSE in Barb-5 spherule layers [13], and data for whole rock analyses of Barb-5 spherule-groundmass assemblages [11], modal abundances of up to 1 wt.% of spinels would be necessary to explain chondritic concentrations in the spherule layers. However, this is not observed.

Conclusions: The following observations can be made from the current dataset for both drill cores: (i) spherule layers and shales and cherts from the Fig Tree group, intercalating these layers, both exhibit HSE concentrations ranging from sub- to slightly suprachondritic abundances, (ii) peak concentrations of up to four times the chondritic value [10] represent the highest abundances ever reported for Archean spherule layers so far, (iii) rough trends toward nearer to chondritic highly siderophile interelement ratios with increasing total Ir content suggest varying meteoritic admixtures to the samples, (iv) samples from both drill cores define rough linear mixing trends connecting non-impact related Fig Tree sediments with chondrites in HS interelement diagrams, (v) initial 187Os isotope signatures further strengthen the proposition of chondritic impactors.

The data unambiguously confirm enormous (up to 100%) amounts of meteoritic HSE admixtures to the spherule samples. The existence of suprachondritic HSE abundances might best be explained as resulting from (i) fractionation processes within the impact plume, (ii) syn- or post impact hydrothermal activity, or (iii) post-impact mechanical enrichment of HSE carrier phases, or a combination of these processes. Post-Archean impact ejecta and deposits typically exhibit orders of magnitude lower meteoritic admixtures compared to Archean spherule deposits. The extreme magnitudes of meteoritic admixtures might, thus, be a general feature that seems to be related to the enormous impactor sizes suggested for the Archean but, potentially, may also indicate impactor compositions that are not represented in the present-day meteorite collections.
Extreme Debris Disks and Terrestrial Planet Formation: Planet formation is ubiquitous, thousands of exoplanets have been detected, indicating that many young planetary systems build terrestrial planets. The most dramatic phases of this process are thought to be oligarchic and chaotic growth, roughly up to ages of 200 million years, when violent collisions occur between bodies of sizes up to proto-planets. Such events should be marked by the production of huge amounts of debris, including clouds of dust, as has been observed in some of the extreme debris disks. I will review the properties of extreme debris disks, and how to use their variable disk emission to study the violent impact events and their aftermaths in young exoplanetary systems during the era of terrestrial planet formation.

Introduction: Planetary systems form and evolve in circumstellar disks. After an initial short-lived (≤10 Myr) and optically thick protoplanetary stage, optically thin debris disks emerge, sustained by the fragmentation of colliding planetesimals and detectable through their infrared emission and some in optical scattered light. Some disk structures (bright disk halo composed of small dust [1], clump and disk asymmetry [2,3,4]) inferred from the resolved images of debris disks have been linked to the aftermath of giant impacts; however, the dust in the majority of debris disks is believed to be the product of steady state collisional grinding of planetesimal belts that are analogous to the Kuiper Belt in the Solar Systems [5,6,7] (termed traditional debris disks). The bright, variable emission by the dust produced in the aftermaths of large impacts in extreme debris disks provides a better opportunity to study the violent events during the era of terrestrial planet formation. Each large impact produces an observable signal due to the production of huge clouds of dust and silica vapor (Fig. 1, [8]). A few % of stars in the appropriate age range for rocky planet formation have extreme debris disks indicative of high rates of collisional activity that should accompany active plant growth [9,10]. Table 1 summarizes the major difference between the two classes of debris disks.

Indicators of Extreme Rates of Planetesimal Impacts: Two lines of evidence show that large-scale impacts of asteroids and/or planetary embryos are ongoing in these extreme debris disks. Firstly, they show very prominent solid-state features in the mid-infrared from sub-μm amorphous/crystalline grains (Fig. 2). Because of the short lifetimes for small grains in debris disks, strong solid-state features point to an elevated level of dust-producing collisions. Furthermore, some extreme systems show a distinct 9 μm (SiO2) feature from freshly condensed silica smoke [11, 12, 13]. The presence of silica in the HD 172555 system indicates hypervelocity (>10 km/s) impacts [12, 14]. In addition to silica grains condensing from vapor, annealing at high temperatures (>1200 K, a localized condition resulting from large-scale impacts) can also transform magnesium-rich silicate material into both crystalline and amorphous forms of silica and crystalline forsterite (Mg2SiO4) [15]. The collision between two large, gravity-dominated bodies is a violent event where powerful shocks vaporize a large fraction of the impactor [16]. Therefore, thermodynamic alteration of minerals is expected, and the detailed characteristics of these crystals, inferred from their mid-infrared spectra provide insights into their formation conditions in young exoplanetary systems [17].

Secondly, the disk emission output in these extreme systems varies on timescales of months to years [18, 22] with one rare example (TYC 8241 2652 1) that the disk almost disappeared over a period of a year [21], in stark contrast to stable disk emission in traditional debris disks. The most thoroughly studied example of disk variability is ID8, a 35-Myr old system in NGC 2547. We found that the disk variation in 2013 required a large impact that produced an optically thick cloud of glassy silica condensed from impact-produced vapor. These condensates were then ground down by a collisional cascade over a nominal timescale of a year, consistent with the gradual flux decline in 2013 [19]. The behavior since then is not predicted by the impact event documented by [19], and indicates multiple, violent impacts producing more debris and possibly at a different radial zone [20], suggesting that the exoplanetary system around this star is undergoing vigorous evolution.

Multi-year Spitzer Monitoring Program and Future JWST Follow-ups: To investigate the incidence, nature, and evolution of these impacts, we were awarded a multi-year Spitzer program to monitor a dozen of extreme debris disks. Our program reveals a rich variety of disk variability indicative of major ongoing episodes of planetesimal impacts, including (1) dramatically rising and falling behavior in ID8, V488 Per, and RZ PsC, (2) a gradual long-term (multi-year)
decay trend on top of stochastic short-term behavior in P1121, HD 113766, HD 23514 and HD15407A. Particularly, the short-term disk flux modulation observed in the 2013 and 2014 ID8 light curves is directly linked to the orbital evolution of the two impact produced clouds [20]. A critical aspect of understanding the disk variability and its implications for terrestrial planet formation is to obtain high-quality mid-infrared spectra to determine the mineralogy of the collision products and to search for variations as the impact products evolve. The combination of the warm Spitzer and JWST observations will document the collisional history and patterns of behavior over a 10-15 year baseline, providing unique insights into the process of planet building in terrestrial zones.

Table 1 – Properties of Two Types of Debris Disks.

<table>
<thead>
<tr>
<th>Traditional Debris Disks</th>
<th>Extreme Debris Disks</th>
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<tr>
<td>No variability with the disk emission stable on long ~10^2 yr to ~Myr timescale.</td>
<td>Disk variability on monthly to yearly timescale.</td>
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<tr>
<td>Infrared fractional luminosity, ( f_d \leq 10^{-4} ), dominated by cold ( T_d &lt; 80 \text{ K} ) dust and only ( \leq 1% ) with a hot component within 1 au.</td>
<td>Bright and hot excesses with ( f_d \geq 10^{-2} ), ( T_d \geq 400 \text{ K} ), and prominent solid-state features, suggesting abundant small grains.</td>
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<tr>
<td>Around all stages of stellar evolution from (~10 \text{ Myr} ) to (~10 \text{ Gyr} ).</td>
<td>All around stars from (~10 \text{ to } ~200 \text{ Myr} ) with one exception BD+20 307.</td>
</tr>
<tr>
<td>Dust is sustained by collisional cascades of (<del>10-100 \text{ km size bodies down to blowout size } (</del>\mu \text{m}) ); timescale for disk variability is on the order of ( 10^2 ) to ( 10^6 \text{ yr} ).</td>
<td>The large amounts of small grains are related to stochastic, large impacts of big asteroids; the short-term disk variability is consistent with the aftermaths of such large impacts.</td>
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Figure 1: The grey histogram is the fraction of stars that have large 24 \( \mu \text{m} \) excesses (left y-axis) vs. age [23]. Various symbols represent the measured 24 \( \mu \text{m} \) excesses relative to stellar photospheres (R24, right y-axis). The red star symbols are the extreme debris disks monitored by our Spitzer program. The colored lines are the expected dust levels from three selective simulations during the giant impact phase from [8]. Each of the spikes represents a giant impact, injecting new debris into the system.

Figure 2: Spitzer mid-infrared spectra of extreme debris disks. The top three systems have silica-rich dust that can be identified by the feature peaking near 9 \( \mu \text{m} \) (dark grey, vertical lines), while the others have complex features due to forsterite (light grey lines). Mineralogical comparison between the dust composition in extreme debris disks and that of meteorites suggests that the material detected in these disks is experienced high temperature events [17].
SHORT-TERM DISK FLUX MODULATIONS DUE TO THE ORBITAL EVOLUTION OF IMPACT PRODUCED CLOUDS OF DUST IN NGC2547-ID8. Kate Y. L. Su, Alan P. Jackson, Ruobing Dong, George H. Rieke, Andras Gaspar, Steward Observatory, University of Arizona (933 N Cherry Avenue, Tucson, AZ 85721, ksu@as.arizona.edu), Centre for Planetary Sciences, University of Toronto, Toronto, ON, Canada.

NGC2547-ID8 – A 35 Myr-old Exoplanetary System: ID8 is a solar-like (G6V) star in the 35 Myr-old open cluster NGC 2547 [1] at a distance of 360 pc [2]. The star shows a weak (0.01 mag) modulation of 5 days in the optical due to spots on the stellar surface, suggesting its rotation axis is unlikely to be pole-on from our line of sight [3, 4]. The spectral energy distribution of the ID8 system shows prominent solid-state features, suggesting the presence of abundant small silicate-like grains. Olofsson and colleagues [5] presented a detailed SED study by simultaneously determining the dust composition and disk properties. They found that ~10% of the small dust in the ID8 system is in the form of crystalline silicates with ~2/3 of them belonged to the Fe-rich crystalline grains. The location of the debris is estimated to be 0.3–0.64 au, and is dominated by sub-μm size grains in a steep power-law size distribution with a total dust mass of 2.4×10⁻⁶ Earth mass.

Expected Short-Term Evolution of an Impact Produced Cloud of Dust: Jackson and colleagues [6] presented a detailed description on the dynamics of debris released by a giant impact. In their dynamical calculations, there are two spatially fixed locations for the evolution of impact produced debris: the collision-point and the anti-collision point. The collision-point is where the impact occurred, which is a fixed point in space through which the orbits of all of the fragments must pass since they originated from there. The anti-collision line is a line exactly on the other side of the orbit from the collision-point through which all of the orbits of fragments must also pass (the alignment of the ascending/descending nodes). For simplicity, here we refer them as collision and anti-collision points. The two opposite points in the orbit of an impact produced cloud naturally explain the bi-periodicity because the extension of the cloud reaches minimum at these two locations. For a system viewed at nearly edge-on geometry like ID8, the projected area of the cloud also reaches minimum as it passes the disk ansae, naturally explaining another bi-periodicity [3]. The combination of disk ansae and collision and anti-collision points predicts an intermixed nature of periodicity without invoking an eccentric orbit. The detailed evolution of the expected disk light curves, their dependency on geometry, impact condition and orbital eccentricity are further discussed by the presentation of Jackson et al. in this conference [7].

3-D Radiative Transfer Calculations: The simple geometric, dynamical model presented in [7] qualitatively describes the expected modulations in the cross section of an impact produced debris cloud. We further carried out 3-D radiative transfer calculations and confirmed the expected flux modulation during its orbital motion. Fig. 1 shows the flux modulation of an optically thick cloud at different orbital phases from two viewing angles: face-on (an inclination of 0°) and close to edge-on (an inclination angle of 85°). The initial point of the orbital phase is defined at the collision point (phase of 0.0). The orbital phase of 1.0 is at the same point but after one orbit of evolution, and the orbital phase of 1.5 is its corresponding anti-collision point. For the face-on case, the disk flux reaches local minimum when the cloud passes the collision and anti-collision points. For theinclined case, the collision point is set exactly between the disk ansae behind the star, i.e., the disk ansae are at the orbital phases of 0.25 and 0.75 after the impact. The flux of the cloud drops whenever it passes the collision and anti-collision points and disk ansae.

Application to the Modulations in ID8: Fig. 2 shows the ID8 flattened (after subtraction of the long-term flux upward and downward trends) disk light curve observed in 2013. By associating the big dips with possible collision and anti-collision points, we identified the true orbital period of the cloud is 108 days. An orbital period of 108 days suggests that the impact occurred at a distance of 0.44 au from the star within the expected debris location. By examining the nearby, secondary dips around the identified collision and anti-collision points, we further identified that the disk ansae are likely at the orbital phases of 0.2 and 0.7, i.e., the cloud reached the disk ansae 21.6 days after passing the collision and anti-collision points. The phase difference between the disk ansa and collision point suggests that the angle between the collision point and the disk ansa is about ~70°. Given the flat disk light curve observed in 2012 (prior to the 2013 event) and the orbital period of 108 days, we also identified that the 2013 impact occurred at BMJD 56227 (2012 Oct 26).

Fig. 3 shows the ID8 flattened disk light curve in 2014, which is very different from the one observed in 2013. Since there is one single period (10.4 days) identified (rather than two intermixed periods), the disk ansae are exactly at the half way between the collision and anti-collision points, suggesting a true orbital period of
41.6 days. This orbital period suggests that the cloud responsible for the 2014 modulation was at an orbital distance of 0.24 au, implying that the modulations in the 2013 and 2014 light curve are caused by two different impact events. We also folded 2015–2017 light curves in phase space with the 2014 one using the same period of 41.6 days to identify additional modulations that might be produced by the same cloud. The flux dips in the 2015 light curve (Fig. 4) are likely associated with the orbital phases of 6.5, 7.0 and 8.0 for the 2014 event. The overall modulation behavior is consistent with the expected evolution – the impact produced clump in the disk lasts for ~10 orbits and the disk flux modulation is strong and observable during this clump phase [6, 7].

Summary: Using the simple geometric, dynamical model presented in [7], we successfully explain the short-term disk modulation observed in the ID8 system. Our analysis suggests that the modulations are caused by two impacts occurred at two different locations and time. The 2013 event occurred in 2012 Oct 22 at 0.44 au (an orbital period of 108 days) while the 2014 event likely occurred in early 2014 at 0.24 au (an orbital period of 41.6 days). The terrestrial zone in the ID8 system is undergoing vigorous evolution.

ASTEROIDAL CONSTRAINTS ON THE EARLY BOMBARDMENT HISTORY OF THE INNER SOLAR SYSTEM. Timothy D. Swindle1,2 and David A. Kring2,3, 1Lunar and Planetary Laboratory, University of Arizona, Tucson AZ 85721-0092 USA, tswindle@lpl.arizona.edu, 2Lunar and Planetary Institute, USRA, 3600 Bay Area Blvd., Houston TX 77058 USA, 3NASA Solar System Exploration Research Virtual Institute.

Summary: Radiometric ages of meteorites from six different groups record bombardment between 4400 and 3000 Ma ago. Most ages are between 3500 Ma and 4000 Ma, and with the exception of the LL chondrites, ages between 4400 Ma and 4100 Ma are extremely rare. This is most consistent with an increase in bombardment rate and/or an increase in impact velocities at ~4100 Ma.

Introduction: Every attempt to determine the history of the flux of impactors to the inner Solar System relies on a dataset that is biased or flawed in some way. The lunar rocks we have are biased by the influence of the Imbrium Basin, which occurred near the center of the KREEP-rich region with the rocks best for dating by many techniques. We don’t have enough old rocks from Mars to provide a meaningful sample, and while we have high-resolution imaging that makes it possible to determine crater densities, turning Martian crater densities into accurate ages requires a better knowledge of the relative cratering rate between Mars and the Moon than we have.

We have rocks from at least five Main Belt asteroids that have radiometric chronologies (most, but not all, determined by the 40Ar-39Ar technique) that appear to have been affected by impacts between ~4400 Ma and ~3000 Ma. The drawback to these is that they come from unknown (with one likely exception) asteroids with histories and/or source locations that make it possible for them to provide meteorites to Earth, but they do provide a set of dates with a different biases than the other sets we have to work with, so we will focus on them. The five sets of meteorites we will discuss are the H, L, and LL chondrites, the HED meteorites, and the silicates in IIE iron meteorites. The data used will be from two reviews [1,2] and references therein. Except where noted, we are not aware of any more recent data that change the fundamental conclusions.

It is worth comparing these to the lunar record. Early results from lunar samples were interpreted as evidence for a “terminal lunar cataclysm” a relatively short period of intense bombardment of the Moon at ~3.9 Ga. As time has passed, it has become clear that the most extreme version of this hypothesis, a period of a few tens of Ma in which all the major basins were formed and virtually all lunar impact melt rocks were formed, is an oversimplification at best. Both lunar meteorites and glass spherules collected by the Apollo program show ages extending to ages of 3.5 Ga or more recently, and evidence for impact melts at ages of 4.2 Ga or more have been found. It is clear that the impact cratering rate on the Moon was much larger between 3.5 and 4.0 Ga than it is at present, but whether that represents the tail end of a monotonically decreasing flux, a period of increased bombardment after a relative lull in cratering, or some combination (e.g., a “sawtooth” pattern) is not at all clear.

One of the clear advantages to using asteroidal meteorites is that they have definitely not been affected by Imbrium. On the other hand, the events recorded among the meteorites are largely simple crater-forming events (otherwise, the PB would have been disrupted), and are not the basin-size impact events that define the intense period of lunar bombardment. Thus, one anticipates they would produce a broader range of ages (4.1 to 3.2 Ga) vs. the lunar basins (4.1-3.8?). Clearly, smaller events would have occurred on the Moon later than 3.5 as the period basin-forming impacts waned.

Asteroidal Meteorite Data

We will discuss the data by meteorite group, where each group presumably represents a single asteroid parent body or similar bodies that accreted together in the same part of the Main Belt. Age data are displayed as relative probability plots, where each meteorite’s measured age is used to construct a Gaussian curve that matches the age and standard deviation, and has the same area under it as every other meteorite. The summed value is equivalent to a histogram.

HED meteorites. Bogard [1] compiled Ar-Ar ages from 46 eucrite samples from various HED meteorites, presumably representative of the asteroid 4 Vesta. The data (Fig. 1) show a strong peak at ages >4400 Ma, a significant peak at ~3700-3800 Ma (data from nine samples overlap), and other possible peaks between 3400 Ma and 4100 Ma. There is far less evidence for any ages 4100-4400 Ma than before or after that period. Note that the old peak comes from unbrecicated eucrites, and may represent cooling or metamorphic ages, not ages of impacts. For the ordinary chondrites discussed below, the samples plotted are only those with petrographic evidence for strong shock events – there are many relatively old ages that could be shock ages, but could also be metamorphic ages.

H-chondrites. The distribution for H chondrites is similar to that for eucrites, for the older meteorites. In particular, there is a cluster of ages at ages >4400 Ma,
a dearth of ages between 4400 and 4100 Ma, then seven meteorites with ages of 3500-4100 Ma. There is one meteorite with an age between 4400 Ma and 4100 Ma, although it has rather large uncertainties. In addition, there are some meteorites on either side of that range whose uncertainties make them compatible with that. However, the data are most consistent with early bombardment (>4400 Ma), followed by a lull until ~4100 Ma, followed by an increase in thermal events associated with cratering. The H chondrites show a considerable amount of evidence for ages <1000 Ma, something that is largely absent in the eucrite ages, despite suggestions of relatively recent large impact events on Vesta (but see [3]).

L chondrites. The data for L chondrites is dominated by the effects of an event at 470 Ma [1,2,4,5]. Among shocked L chondrites with evidence for older ages, there are five with ages ≥4400 Ma and three with ages of 3000-3800 Ma.

Silicates in IIE irons. Silicates from five IIE irons give Ar-Ar ages of 4400 Ma or greater, but those from Watson, Kodaikanal, and Netschaëvo give concordant Ar-Ar, U-Pb and Rb-Sr ages of 3700 Ma [1].

LL chondrites. The data for LL chondrites is completely different from the other four asteroidal meteorite groups discussed. In the case of the LL chondrites, six have ages of 4100-4400 Ma, with one possible older age, and two ages of ~3900 Ma. The dominance of ages 4100-4400 Ma over those 3500-4100 Ma is what would be expected if the impactor flux was steadily declining, although more ages ≥4400 Ma would be likely in that scenario. On the other hand, the LL chondrites are rich in breccias, and in clast-rich and rapidly cooled impact melts, which could suggest different types of impact events or a different location within the asteroid belt for the LL-chondrite parent body [6].

Discussion: Thus, with the exception of evidence from the LL chondrite parent body, asteroidal meteorites appear to record an accretional interval of cratering events circa 4.4-4.5 Ga, followed by a lull, before the belt was dynamically excited and cratering events increased from 4.0 to 3.5 Ga.


Fig. 1. Relative probability plot for 46 eucrite samples, from [1].

Fig. 2. Relative probability plots for ordinary chondrites that have petrographic evidence of strong shock events [2].
THE LATEST VIEWS ABOUT LATE ACCRETION. Richard J. Walker and Katherine R. Bermingham, Department of Geology, University of Maryland, College Park, MD 20742; rjwalker@umd.edu

Introduction: “Late accretion” is a term that is widely used in the planetary community, but has come to mean different things to different investigators. It is a concept that was created in the 1960’s to explain the apparent overabundance of highly siderophile elements (HSE) in Earth’s mantle relative to what would be expected from metal-silicate partitioning known at the time. It refers to accretion that occurs subsequent to the cessation of core segregation. In the case of Earth, the total mass of late accreted materials would need to equal or exceed ~0.5 wt.% of Earth’s mass to account for the HSE present in the mantle [1-2]. The idea is that as long as metal is extracted from the silicate shell of a planetary body into a growing core, HSE abundances would remain extremely low as a result of the strong preference for these elements to enter metal compared to silicate. Thus, additional accretion would be needed to account for the abundances of these elements. In addition to supplying HSE to the mantle, numerous studies have also invoked late accretion as a means to deliver volatile-rich and organic-rich material to Earth’s mantle as well as the mantle of the Moon [3]. Some have envisioned this process as a consisting of a gentle rain of carbonaceous chondrite material that formed a “late veneer”. Some studies of HSE in mantle-derived komatiites on Earth have interpreted changes in projected source concentrations with time to indicate a downward mixing of this putative veneer over ~1 to 2 billion years [4].

The concept of late accretion has strengthened somewhat over the years as new geochemical data and dynamical models have provided compelling evidence for the process. For example, Os isotope and relative abundances of most HSE projected for the bulk silicate Earth (BSE) appear to be within the range of chondritic meteorites. These characteristics are difficult to explain by indigenous processes. Nevertheless, some experimental studies of metal-silicate partitioning have discovered conditions (normally at high temperatures and pressures) at which the HSE become less highly siderophile and might conceivably explain the abundances of at least some HSE in the mantle by metal-silicate segregation at the base of a deep terrestrial magma ocean [5]. Given the rapidly growing evidence for and against late accretion, it is valuable to re-assess the concept.

Evidence For and Against Late Accretion: Although frequently forgotten, Os isotopes continue to be the strongest evidence for late accretion. They leave little wiggle room with respect to the precisely chondritic relative abundances of Re-Pt-Os in the bulk silicate Earth (BSE). Conversely, Ru/Ir and Pd/Ir ratios estimated for the BSE are not perfect matches to chondritic meteorites [6], and may indicate that either our ability to estimate BSE ratios from upper mantle materials is not as good as we wish to believe, or that additional processes may have been involved, such as the generation of a Hadean sulphide matte [7]. Alternatively, these ratios may also mean that late accretion was not the process that dominated their abundances in the mantle.

Other bodies provide some supporting evidence for late accretion as a major process in the planetary accretion process. Comparative Earth-Moon mantle abundances of HSE, together with the ~25 ppm difference in 182W/184W between the silicate shells of the two bodies provides evidence for stochastic late accretion, whereby late accretion to Earth was dominated by a small number of Pluto mass bodies [8]. This conclusion, coupled with recent dynamical modeling of impactors of that size, suggests that late accreted materials may have been forcefully injected into the mantle with mantle and cores separating, and should not be thought of as having formed a “veneer” [9]. Further, Os isotopes and HSE abundances of mantle-derived rocks from Mars are quite similar to terrestrial compositions and suggest a similar late accretionary process was responsible for establishing the abundances of HSE in the mantle of another, albeit much less massive planet.

New Concepts Injected Into Old Ideas: Warren [10] divided planetary materials into carbonaceous chondrite (CC) and non-carbonaceous chondrite (NC) groups based on mass independent genetic isotopic signatures of lithophile elements such as O, Ti and Cr. The mass independent isotopic compositions of these elements vary among meteorite groups as a consequence of either their parent bodies incorporating varying proportions of presolar materials (Ti, Cr), or by incorporating variable proportions of materials that had been isotopically modified by photochemical processes in the nebula (O). It has been presumed by some that the CC group, consisting of carbonaceous chondrites, as well as some iron meteorite groups, formed in the outer solar system, perhaps beyond the orbit of the proto-Jupiter [11].

One of the most exciting advances in understanding late stages of planetary accretion during the past 10
years comes from the application of isotopic genetic tracers based on siderophile elements, especially the HSE Ru and the moderately siderophile element Mo. As with the lithophile element genetic tracers, Ru and Mo can be used to distinguish between NC and CC groups. As originally noted by Dauphas et al. [12], and referred to as the “cosmic correlation”, Ru and Mo isotopic data for most meteorites plot along a linear trend, with carbonaceous chondrites and CC irons plotting at one end, and data for NC meteorites extending along the linear trend, terminating at the terrestrial composition. Data for our work on iron meteorites are shown in Fig. 1.

Figure 1. The Mo–Ru cosmic correlation defined by $\mu^{97}\text{Mo}$ and $\mu^{100}\text{Ru}$ for iron meteorite groups and individual meteorites. All data were corrected for cosmic ray exposure, and all data were also collected from the same ~ 2cm portions of the meteorites. The circles denote NC groups and the triangles denote CC groups, defined by Mo isotopic compositions. The solid black line is the linear regression through the data and the error envelop is indicated by the gray dotted lines. From Bermingham et al. [13].

Two important findings are revealed from these data. First, the Ru isotopic composition of the BSE is at the opposite end of the trend from CC group meteorites. This means that the dominant component of the putative late accretion could not have been carbonaceous chondrite-like material. Thus, unless there are as yet unidentified water/organic-rich NC group meteorites, late accretion could not have been the process that delivered much of Earth’s water and organic molecules.

Second, again as originally noted by Dauphas et al. [12], the fact that the Ru-Mo isotopic compositions of Earth lie along the cosmic correlation means that the major final building blocks of Earth did not substantially change from the final 10-20% of accretion that provided the Mo to the mantle. Otherwise the BSE would not plot along the now well-defined trend.

It will be important to explore the genetic signatures of materials added to the terrestrial planets at the tail end of the late accretion process, which some have proposed to have been a terminal cataclysm or terminal bombardment. Siderophile element isotopic studies of lunar impact melt rocks and soils will ultimately reveal whether the genetic characteristics of the latest stage of major accretion remained the same as during the preceding 10-20% of accretion.

Introduction: Scientists have questioned the characteristics of the lunar impact flux since before the first Apollo samples were brought to Earth. Baldwin [1,2] and Hartmann [3,4] proposed that the lunar impact flux was roughly 200× the average post-mare extrusion rate with a peak rate that was most likely higher. After Apollo, a new interpretation based on U-Pb ages of lunar samples emerged when Tera et al. [5; and others using different radiometric systems] proposed that the Moon had experienced episodic impacts, culminating in the formation of Imbrium, Crisium, and Orientale (and perhaps other basins), within 30 million years (My) of each other, around 3.9 billion years ago (Ga); this event was called the “terminal lunar cataclysm” [5,6]. In an extreme interpretation of Apollo and Luna samples that showed a similar age, Ryder [7] proposed that the Moon experienced only light bombardment in its first 600 Ma and that all of the large near-side basins formed in a narrow window of time centered around 3.85 Ga. Further work by Dalrymple and Ryder [8,9], as well as investigations by others, appeared to support this interpretation.

In the past several years, however, orbital data from lunar spacecraft, along with new and improved analyses of lunar samples, indicate that the lunar nearside observations and samples may have biased our interpretation of Apollo and Luna impact data. In particular, Imbrium samples may have contaminated all (or most) of the nearside sites from which samples were collected [e.g., 10]. Dynamical models that once provided support for the “cataclysm” [e.g., the Nice Model; 11,12] have been modified to take into account these new orbital and sample data [e.g., 13,14].

Impact Frustration: How the impact flux inferred from the Moon may have affected conditions on the early Earth has been an ongoing concern as well. In the view of Maher and Stevenson [15] and Sleep et al. [16], a “cataclysmic” bombardment would have sterilized the Earth, leading to an “impact frustration” that made it difficult for early life to form and evolve on Earth; if the Moon’s impact rate can be applied to Mars, any life on Mars would most likely be affected, too. This proposed impact sterilization would thus constrain the time interval for the formation of life, leading to bottlenecks from which a last universal common ancestor may have emerged. However, if interpretations of ages of lunar impactites were meant to improve our understanding of the lunar impact flux in the context of Solar System bombardment, then analyses of lunar meteorites, meteorites from asteroids, and other lunar samples should, at some level, provide similar evidence (Figure 1).

Extraterrestrial Samples: A variety of lunar impact samples have been investigated, both compositionally and with radiometric systems. Meteorites and Apollo and Luna samples provide unique information about the lunar impact flux, across a range of impactor sizes. While Apollo samples have been shown to be influenced by Imbrium [e.g., 17,18,19], there does exist evidence for very old (i.e., >3.9 Ga) impacts [e.g., 20,21,22]. These old ages are additionally found in lunar meteorites [e.g., 23,24] and meteorites from asteroids [e.g., 25,26,27]. Lunar impact glasses [e.g., 28] and paleoregoliths [29] also show formation and closure (respectively) ages ≥3.9 Gy old.

Orbital Data: High-resolution images and data from the LRO, SELENE, and GRAIL spacecraft have allowed for an improved investigation of the lunar surface. In particular, ancient basins that are buried or otherwise obstructed from view have been “uncovered” by altimeter [e.g., 30,31] and gravity measurements [32]. As a result, numerous very old basins (>300 km diameter) have been identified, providing evidence for ancient impacts early in Solar System history.

Terrestrial Samples: Terrestrial impact samples older than ~3.5 Ga are elusive, though many impact-affected zircons [e.g., 33] and Archean spherule beds [e.g., 34,35,36] provide evidence for impacts. However, little evidence to support a “cataclysm” on Earth has been found.

Effect of Impacts on Early Life: Taking together terrestrial data and lunar sample and orbital data, dynamical models favoring a protracted “sawtooth” bombardment, during which the impact flux was never exceptionally high, have been proposed [e.g., 13,37]. Evidence from all of these data sets should describe the nature of the impact history of the Moon, and its presumptive application to Earth. Consequently, what we learn from these data affects our understanding of Earth’s habitability and the conditions that existed when life first appeared on a young Earth. While “impact frustration” and a sterilized Earth may no longer be important considerations, given the changing view of the impact flux from “cataclysm” to “sawtooth”, impacts did affect Earth, with heating perhaps only locally severe [e.g., 38,39]. Additionally, any existing hydrosphere and/or atmosphere would be affected.

Thus, it is important to consider the terrestrial biological evidence that has come to light in the past several years. For example, Bell et al. [40] reported putative biogenic C in a 4.1 Gy-old zircon, 200 My after
Evidence for the appearance of liquid water [e.g., 41,42] and about 100 My before the oldest continental rock surfaced on the early Earth [43]. Biogenic C has been reported in multiple other samples, including those from the Sagglec area in Canada [3.95 Ga; 44] and Akilia Island [3.85 Ga; 45] and the Isua Greenstone Belt in Greenland [3.8 Ga; 46]; together this evidence pushes back the proposed appearance for life to >3.7 Ga [47]. Even more extraordinary, Dodd et al. [48] proposed 3.77-Gy-old “tubelike” fossils in a hydrothermal system, and 3.7 Gy microbial structures similar to stromatolites have also been found [49]. The earliest signs of life on land indicate an appearance ~3.48 Ga [50], shortly after the onset of plate tectonics [3.5 Ga; 51] and before the oldest terrestrial impact craters [3.470 Ga; 52]. Even with large impacts occurring until ~3.23 Ga [34], fossil evidence shows that life on Earth was established by 3.43 Ga [53].

**Conclusion:** The first billion years of Solar System history was an exciting time for terrestrial biology, in spite of the impacts (Figure 1). A protracted “sawtooth” model appears to satisfy the lunar impact data, lunar orbital data, terrestrial impact record, and terrestrial biological evidence; a “ecataclysm” does not [e.g., 54,55,56]. How the Moon’s impact flux could (or should) be applied to Mars, and the subsequent investigation of Mars’ habitability, is largely dependent on our understanding of the Solar System’s dynamical evolution and analyses of martian orbital data and samples. However, evidence for past and current water on Mars, and the detection of organic molecules [e.g., 57], make for intriguing possibilities. In any case, the information we have learned, using the Moon’s impact flux as a proxy for Earth’s, has made this an exciting time in planetary science.


![Figure 1](image-url) Figure 1. The impact flux in the first billion years of the Solar System, as represented by lunar impact samples and meteorites, is shown, with the influence of Imbrium samples eliminated. Terrestrial biological activity at specific times (as described in the text) is marked by *. Figure modified from [54].
Imbrium Zircon Age for Apollo 73155 Serenitatis Impact Melt Breccia: Implications for the Lunar Bombardment History
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Introduction: The Serenitatis basin is thought to be one of the oldest basins on the near side of the Moon [1]. The Apollo 17 landing site is located within the southeast area of Serenitatis basin. This area is of special interest as a site for ancient crustal rocks which were thought to have avoided the influence of Imbrium-basin-forming event [2]. Chemically and petrologically distinctive from Apollo 14 breccias (representative of Imbrium ejecta) [3], Apollo 17 samples are regarded as a single large impact source from Serenitatis basin [4]. Previous geochronological studies of Apollo 17 impact melt breccias (IMBs) by Ar-Ar chronometer place the age of Serenitatis basin formation at ~3.9 Ga [5]. This age for Serenitatis basin was interpreted as a spike in the impact flux during the so-called period of late heavy bombardment (LHB).

However, it has been argued that the predominant ~3.9 Ga ages of lunar near-side basins (including Serenitatis basin) may be due to Imbrium contamination [6]. Recent Ar-Ar studies of Apollo 16 [7] and Apollo 17 [8] highland IMBs suggested that they may actually contain considerable Imbrium components. Here, we present petrological, geochemical, and U-Pb geochronological results for Apollo 17 highland IMB 73155,69 thin polished section.

Fig. 1. A representative ~50 μm zircon grain (backscattered electron image) in Apollo 73155 IMB.

Petrology and Mineralogy: Apollo 73155 is an impact melt breccia with fine-grained aphanitic matrix (~90 vol.%). It was sampled on the Light Mantle deposit at the foot of South Massif (Station 2A) in the Taurus-Littrow Valley, southeastern Serenitatis basin. The matrix consists of olivine, pyroxene and plagioclase. The Apollo 73155,69 has pigeonite to augite pyroxenes, Ca-rich plagioclase and Mg-rich olivine. The grain size of zircons ranges from a few microns to 50 μm. Most zircon grains are anhedral and have vermicular intergrowths with plagioclase and pyroxene grains (Fig. 1). This unusual texture of lunar zircons has been identified in the Imbrium-originated samples of lunar meteorite Sayh al Uhaymir (SaU) 169 and Apollo 12 high-Th IMBs [9].

Results and Discussion: Nanoscale secondary ionization spectrometry (NanoSIMS) analyses of the 73155 zircons have a weighted mean Pb-Pb age of 3928 ± 10 Ma (N=10, 2σ; Fig. 2). This age is consistent with the Ar-Ar age of 73155 matrix at 3937 ± 16 Ma [5]. The vermicular zircons of Apollo 12, Apollo 73155 and SaU 169 are thought to have crystallized in-situ and simultaneously within impact melts due to their small sizes (<20 μm) and enclosed inclusions (matrix minerals) [9-11]. This scenario is supported by the presence of high-Zr contents of the matrix inclusions within 73155 zircons. The vermicular intergrowths of these zircons with matrix minerals, and the overlapping ages of zircons and matrices, indicate that Apollo 73155 IMB and its zircons formed contemporaneously, and coeval with the Imbrium event.

Fig. 2. Weighted mean and 2SD Pb-Pb age of 10 73155 zircon spots.

The zircon grains found in Apollo 73155 have comparable texture, ages and rare-earth-element (REE) concentrations to those found in Imbrium-originated SaU 169 and Apollo 12 high-Th IMBs. The 73155 zircon Pb-Pb average age is similar to Imbrium meteorite SaU 169 (3920 ± 13 Ma, 2σ) and Apollo 12 IMBs (3914 ± 7 Ma, 2σ) within errors. Besides, Apollo 73155 zircons have
extraordinarily high contents of U, Th and Y. The 73155 zircons (12 spots on 10 grains) have high contents of Y, ranging from 3279 to 6347 ppm; U and Th concentrations vary between 131–457 ppm and 254–302 ppm, respectively. These high Y, U and Th concentrations are similar to SaU 169 (U=161–294 ppm, Th=35–288 ppm, Y=1024–10318 ppm) and Apollo 12 (U=60–519 ppm, Th=51–425 ppm) samples [9, 11]. The Th contents and Th/U ratios of 73155 zircons drift distinctively from those of other Apollo 17 zircons (Fig. 3), which have a mean Th content of 49.9 ± 37.7 ppm (N=133, 1SD) and a mean Th/U ratio of 0.49 ± 0.13 [12-15]. In summary, Apollo 73155 zircons with high-U and Th contents, ~3.92 Ga ages and Imbrium-like textures should have a similar source as SaU 169 and Apollo 12 high-Th samples. The lunar meteorite SaU 169 and Apollo 12 high-Th IMB were related to the Lalande impact crater within Imbrium basin [9, 16]. Therefore, the zircons in 73155 IMB may originate from the Lalande crater, and afterwards they were transported to its recovery site and recrystallized therein.

The aphanitic Apollo 73217,52 has impact ages of baddeleyite at 3929 ± 10 Ma, apatite at 3936 ± 17 Ma and polycrystalline zircon at 3934 ± 12 Ma [14], as well as the phosphate ages of Apollo 17 Station 2 and 6 at 3922 ± 5 Ma and 3930 ± 5 Ma [17], are indistinguishable from the Imbrium ages indicated by those of Apollo 73155, SaU 559, Apollo 12, Apollo 14 IMBs [18-20]. These clustered U-Pb impact ages at 3.92–3.94 Ga might be the most precise estimate of the Imbrium event [21].

Fig. 3. Th and U concentration distribution of zircons in 73155, Imbrium IMBs (SaU 169 [9, 11] and Apollo 12 [9]) and other Apollo 17 samples [12-15].

Apollo 17 impact melts have been related to the production of Serenitatis ejecta because of mutual proximity [1]. Aforementioned in-situ U-Pb ages and Ar-Ar studies suggest that Imbrium-related materials can be identified in the Apollo 17 IMB suite. Moreover, topographical and crater-counting data from the Lunar Orbiter Laser Altimeter (LOLA) and the Lunar Reconnaissance Orbiter Camera (LROC) provide evidence that Serenitatis basin may be in fact older than Nectaris basin. From LOLA data, the crater density of Serenitatis basin (N(20) = 298 ± 60 km⁻²), which is three times higher than that of Crisium (113 ± 11 km⁻²), indicates that Taurus Mountains region predates Nectaris (135 ± 14 km⁻²) [22]. From LROC images, the Sculptured Hills, a unit in the Taurus Mountains between Serenitatis and Crisium, is not directly related to Serenitatis ejecta but instead to Imbrium ejecta [23].

Conclusions: The Apollo 73155 zircon grains have textural characteristics (vermicular intergrowths with matrix), U-Pb age (3928 ± 10 Ma) and trace-element contents comparable to those found in SaU 169 and Apollo 12 high-Th IMBs from Lalande crater area, Imbrium basin. Apollo 73155 zircons point thus towards an Imbrium component within Apollo 17 IMBs (Serenitatis samples). Together with updated topographical datasets from the LROC and LOLA, we suggest that Serenitatis basin was contaminated by Imbrium ejecta and is older than Nectaris basin. These results and observations support other evidences that a “Terminal Cataclysm” scenario for the lunar surface at ~3.9 Ga may have been overestimated because of the transportation of Imbrium-related impactites to other large basins.
