LPI
SUMMER INTERN PROGRAM
IN PLANETARY SCIENCE

Papers Presented at the

August 11, 2011 — Houston, Texas
HIGHLIGHTS

Special Activities

June 6, 2011   Tour of the Stardust Lab and Lunar Curatorial Facility   JSC

July 22, 2011  Tour of the Meteorite Lab   JSC

August 4, 2011 NASA VIP Tour   
NASA Johnson Space Center and Sonny Carter Training Facility Site Visit   JSC/ SCTX

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AGENDA

8:00 a.m.  BREAKFAST

8:20 a.m.  Introductory Remarks by Drs. Paul Spudis and Dave Draper

8:40 a.m.  RACHEL BARNETT, University of New Mexico
(Advisors:  Jones J. H. and Draper D. S.)
Parental Liquid Composition of the Martian Meteorite NWA 6162 Shergottite

9:00 a.m.  KEVIN CANNON, Queen’s University
(Advisors:  Sutter B. and Ming D.)
Calcite and Magnesium Perchlorate Interaction in the Mars Phoenix Thermal and Evolved Gas Analyzer (TEGA):  An Inorganic Explanation for the Low-Temperature CO₂ Release Detected by TEGA.

9:20 a.m.  MATTIAS EK, University of Gothenburg
(Advisor:  Mittlefehldt D. W.)
In Situ Analysis of Orthopyroxene in Diogenites Using Laser Ablation ICP-MS.

9:40 a.m.  JULIA GORMAN, University of Rochester
(Advisors:  Gross J. and Treiman A.)
Spinel-Rich Lithologies on the Moon:  An Experimental Study of Possible Precursor Melt Compositions

10:00 a.m.  YIFAN WANG, Imperial College London
(Advisor:   Shebalin  J. V.)
Simulation of MHD Turbulence in a Spherical Domain

10:20 a.m.  BREAK

10:40 a.m.  ERICA JAWIN, Mt. Holyoke College
(Advisor:  Kiefer W. S.)
The Relationship Between Radar Scattering and Surface Roughness at Lunar Volcanic Domes

11:00 a.m.  KELLY NICKODEM, University of Notre Dame
Advisors:  Righter K. and Danielson L.)
Core-Mantle Partitioning of Volatile Elements and the Origin of Volatile Elements in the Earth

11:20 a.m.  SPENSER PANTONE, Weber State University
(Advisor:  Spudis P. D.)
Understanding the Scattering Properties of Lunar Geological Units Using Mini-RF and LROC Data

11:40 a.m.  KATHRYN POWELL, Rice University
(Advisors:  McGovern P. J. and Kramer G.)
Olivine Detections at the Rim of Crisium Basin with Moon Mineralogy Mapper

12:00 p.m.  LEE SAPER, Brown University
(Advisors:  Allen C. C. and Oehler D. Z.)
Exploration of Acidalia Mensa and Acidalia Planitia:  Probing Mud Volcanoes to Sample Buried Sediments and Search for Ancient and Extant Life
12:20 p.m.  **LILLIAN SCHAFFER, University of Houston**
(Advisor: Kring D. A.)
*Petrology of an Impact Melt Clast from Lunar Regolith Breccia 60016*

12:40 p.m.  **SAMANTHA JACOB, University of Hawai‘i at Manoa**
(Advisors: Mercer C. N. and Treiman A.)
*Tracking the Process of Volatile Release from the Lunar Highland Breccia Meteorite NWA 2996 Using Vesicle Size Distributions*

1:00 p.m.  **ADJOURN**
Lunch provided in the Great Room
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PARENTAL LIQUID COMPOSITION OF THE MARTIAN METEORITE NWA 6162 SHERGOTTITE. R. Gaylen Barnett1, John H. Jones2, David S. Draper3, and Loan H. Le1, 1University of New Mexico MSC04 2545, 1 University of New Mexico, Albuquerque, NM, 2Mail Code KR, NASA, Johnson Space Center, 2101 NASA Road 1, Houston, TX 77058, 3Jacobs Sverdrup Co., Houston, TX 77058.

Introduction: The Martian meteorite North West Africa 6162 (NWA 6162) is a shergottite found in Morocco in 2010. Only one abstract is published about NWA 6162. It discusses the bulk composition of the meteorite and the composition of the olivine cores and rims. The meteorite bears olivine crystals with Mg-depleted rims of up to Fo65.1 and more Mg-rich cores of up to Fo84.1 [1]. It is similar both in appearance and composition to SaU 005. The parent magma from which the meteorite cooled is within the mantle of Mars.

Statement of Problem: Our objective is to determine if it represents a liquid or if it is a product of olivine accumulation. This accumulation would leave the parent melt more Mg-depleted and the olivine cumulates Mg-enriched. Understanding more about the origin of NWA 6162 will give insight about the composition of the Martian mantle. It may also make it possible to choose subsequent experiments to perform to constrain compositions of more primitive Martian liquids.

Approach: Experiments were conducted at JSC at 1 bar using gas-mixing furnaces over a range of temperatures with a controlled oxygen fugacity of IW (Iron Wüstite) +1 (i.e. approximate partial pressure of oxygen at one log unit above IW). The temperature ranged from 1550°C to 1150°C and the experiments were run at 50°C increments. Durations ranged from 1550°C to 1150°C and the experiments were run at 50°C increments. Durations ranged from one hour to 168 hours with longer durations at lower temperatures.

To make the starting material, oxides and carbonates were dehydrated in an oven with temperatures ranging from 350-1300°C. Then, all materials were transferred to a desiccator before water from the air could be absorbed in any noticeable amount. The materials were then weighed to desired proportions and ground in the agate grinder with acetone and again ground with acetone in an agate grinder (note that 5% of experimental glasses and observed smooth trends, suggesting systematic behavior of these elements (Figure 2). It can therefore be concluded that these experiments were acceptably close to equilibrium.

Results: Experimental results are summarized in Table 1. No crystals were present at 1550 or 1500°C. At 1450°C, in charge NWA 3 chromite and zoned olivine crystals appear. The olivine had cores of ~Fo88-Fo96 and rims of ~Fo65-Fo85. It was the only charge in...
which zoned olivine crystals were present. Every subsequent charge also contained olivine. At 1300°C pyroxene crystals appeared, and were also present in every charge run below this temperature.

The Mg content of the olivine crystals decreased as the temperature decreased (see Table 1). The closest we came to reaching the original forsterite composition was at 1300°C and below, where olivine ranged between Fo$_{69}$-Fo$_{75}$.

Because the first olivine crystals to form yielded a more forsteritic composition than those in the original meteorite we conclude that NWA 6162 is a product of accumulation.

**Discussion:** We plotted Mg content of olivine crystals vs. temperature and also the weight percent (wt%) modes of each mineral phase vs. temperature. The mode wt% of glass decreases by more than 10 wt% at each increment between 1450°C and 1300°C, but only decreases by 2 wt% between 1300°C and 1250°C (NWA 4). Likewise, olivine wt% displays a similar pattern, increasing by 8-13wt% between 1450°C and 1300°C and increasing only 2 wt% between 1300°C and 1250°C.

We offer two explanations for the similarities between NWA 4 and NWA 9. First, NWA 9 ran for 168 hours, the longest duration any of the experiments ran. It experienced significant sodium loss in the glass, decreasing from 0.6% in NWA 14 to 0.1% in NWA 9. This loss also accounts for the higher least squares value of NWA 9.

Second, we suspect there is a change in mineralogy between NWA 9 and NWA 4. It is possible that the pyroxene crystals are changing from orthopyroxene to pigeonite. The EMPA can not detect this change in structure, so further analysis using different methods would be necessary to confirm this hypothesis.

As previously stated, the olivine rims in the meteorite are ~Fo$_{65.1}$. None of these experiments produced olivine crystals with this Mg content. The amount of Mg in these crystals decreases with decreasing temperature, however NWA 10 proved difficult to analyze in the EMPA because of the small melt pockets, so lower temperature experiments may not be possible to analyze.

**Conclusion:** The data collected by the experimental results provide evidence that NWA 6162 is the product of olivine accumulation. It is desirable to run another charge at 1300°C for a duration fewer than 168 hours to test whether sodium loss accounts for the discontinuity described above. It is also planned to run experiments that attempt to produce olivine crystals with similar Fo values to that of the original meteorite. When this is accomplished it may be possible to constrain more closely the composition of NWA 6162’s parent magma.

**Acknowledgments:** We would like to thank Anne Peslier and Kent Ross for their help in obtaining data from the EMPS and SEM, as well as Lisa Danielson and Tomohiro Usui for their aid in this project and Steven Sendejo for repairing furnaces. I would like to thank LPI for providing me with this opportunity.
Table 1: Run number, temperature, duration, phases present, Mg content in charge. Gl=glass, olv=olivine, chr=chromite, pyx=pyroxene. *experiment was air quenched

CALCITE AND MAGNESIUM PERCHLORATE INTERACTION IN THE MARS PHOENIX THERMAL AND EVOLVED GAS ANALYZER (TEGA): AN INORGANIC EXPLANATION FOR THE LOW-TEMPERATURE CO₂ RELEASE DETECTED BY TEGA. K. Cannon¹, B. Sutter², D. W. Ming³ and W. V. Boynton⁴ ¹Department of Geological Sciences and Geological Engineering, Queen’s University, Kingston, Ontario, K7L 3N6, Canada (kc19@queensu.ca), ²Jacobs-ESCG, Houston, TX (brad.sutter-2@nasa.gov), ³NASA Johnson Space Center, Houston, TX, ⁴Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ.

Introduction: Along with calcium carbonate the Phoenix Lander’s Thermal and Evolved Gas Analyzer also detected a lower temperature CO₂ release between 400°C and 680°C (Figure 1) in the Phoenix soil [1]. Possible explanations for the low-temperature CO₂ release include decomposition of Mg or Fe carbonates, release of CO₂ adsorbed to zeolites, or decomposition of organic compounds in the soil. The presence of calcium carbonate is supported by the alkaline nature of the soil (pH = 7.7), believed to represent the buffering capacity of the HCO₃⁻ = CO₃²⁻ system [2].

Figure 1. Plot of evolved CO₂ and oven temperature versus time from the TEGA Wicked Witch sample, showing a high-temperature (H) CO₂ release from CaCO₃ decomposition and an unidentified low-temperature release (L). Modified from [1].

The Wet Chemistry Laboratory (WCL) on Phoenix revealed 0.6 wt % perchlorate in the soil [3], a surprising finding with implications for the hydrology and astrobiology of Mars [e.g. 4,5]. Geochemical models using WCL data suggest the existence and possible dominance of Mg-perchlorates in Phoenix soils [6]. Ming et al. [7] demonstrated that perchlorates could have oxidized organics in TEGA, preventing their detection. The detection of Martian perchlorate prompted Navarro-Gonzáles et al. [8] to revaluate the Viking organic analysis, using Mars analog soils mixed with perchlorate. Their results suggested that both Viking landers in fact detected organic compounds mixed with perchlorate salts.

Magnesium Perchlorate: Magnesium perchlorate hexahydrate is stable at ambient Martian and terrestrial surface conditions [9]. Devlin and Herley [10] investigated the thermal decomposition of Mg(ClO₄)₂•6H₂O using differential scanning calorimetry (DSC) and X-ray diffraction (XRD). After dehydration, two endotherms at 410-478°C and at 473-533°C correspond to the breakdown of Mg(ClO₄)₂. They propose a two-stage process:

1) 2Mg(ClO₄)₂ = [MgO•Mg(ClO₄)] + Cl₂ + 3.5O₂

2) [Mg(MgClO₄)] = 2MgO + Cl₂ + 3.5O₂

In simulated TEGA conditions, Mg(ClO₄)₂•ₙH₂O phases break down at lower temperatures than carbonates, giving off Cl₂ and HCl gas [11]. The temperature range of this decomposition occurs at similar temperatures as the low-temperature CO₂ release detected by TEGA (Figure 1) If the chlorine gas produced mixes with moisture in the soil, then HCl gas can form and react with calcite to evolve CO₂:

3) CaCO₃ + 2HCl = CaCl₂ + CO₂ + H₂O

Other possible perchlorate species, NaClO₄ and KClO₄, are not likely involved in this reaction because their thermal decomposition does not evolve chlorine [12,13].

Calcite is present in higher molar amounts than perchlorate in the Phoenix soil; therefore it will not completely react and will decompose normally at higher temperatures, producing the second higher-temperature CO₂ release seen in TEGA data (Figure 1).

Here, we propose that magnesium perchlorate reacted with calcium carbonate in TEGA’s oven to produce the unexplained low-temperature CO₂ release. The objectives of this work are to (1) use laboratory DSC/mass spectrometry (MS) to simulate TEGA analysis where a series of calcite and Mg-perchlorate mixtures will be evaluated to determine if a reaction between the two could produce the low-temperature CO₂ release detected by TEGA and (2) utilize heated powder XRD to assess the mineralogical changes that occur during simulated
TEGA analysis of calcite and Mg-perchlorate mixtures.

**Materials and Methods:** Iceland spar calcite from Chihuahua, Mexico (Ward's), and reagent-grade magnesium perchlorate hexahydrate (Sigma-Aldrich) used in this work were crushed and sieved to 53-150 µm particle sizes for DSC experiments, and to <53 µm for better resolution in the powder XRD studies.

Simulated TEGA analyses were conducted on a Setaram Ligne 96 heat flux DSC coupled to a Pfeiffer Vacuum ThermoStar GDS 301 mass spectrometer. Experiments were carried out at 12 mbar with a flow rate of 1 sccm ultrapure N2 carrier gas. Samples were heated in an alumina crucible to 1350°C at a rate of 20°C/min. A reheat to 1350°C was used as a baseline and subtracted from the first temperature ramp.

The absolute amount (0.025 mmol; 2.5 mg) of calcite used in the DSC/MS experiments was similar to the absolute amount estimated by the TEGA analysis and was held fixed for all runs. Calcite was mixed with 0.0015, 0.01325, 0.025, 0.03675 and 0.0485 mmol Mg(ClO4)2•6H2O. The 0.0015 amount of perchlorate was similar to amounts estimated to be in the Phoenix soil [3]. All experiments were run in duplicate. A calibration curve for the amount of CO2 detected in the mass spectrometer was produced by heating isolated CaCO3 in the DSC and integrating the evolved CO2 using the program OriginPro (vers. 8.5.1, OriginLab).

Powder XRD was performed with an Anton Paar XRK900 heating stage on a PANalytical X'pert Pro diffractometer using CoKα (Fe filtered) radiation. 32.8 mg Mg(ClO4)2•6H2O was mixed with 167.2 mg CaCO3. This is the ratio they were detected in the Phoenix soil scaled up to the 200 mg that fits in the sample holder. The chamber was held at 12 mbar pressure with an N2 flow rate of 1 sccm, the same as the DSC/MS experiments. Scans were collected at 25°C intervals between 25°C and 875°C. The temperature was increased between each step at 20°C/m followed by a 30-minute equilibration period before the next scan. All X-ray scans were taken between 4° and 80° 2θ with an effective step size of 0.0167° and a 100 second counting time.

**Results:** Mass spectrometry results showed two CO2 releases when calcite was mixed with magnesium perchlorate in the Phoenix calculated amounts: a high temperature one consistent with the decomposition of CaCO3, and a low-temperature one (Figure 2). The low-temperature CO2 release occurred in the 375-600°C temperature range, which overlaps with the CO2 release detected by TEGA (400 - 680°C). The low-temperature CO2 release directly followed a spike in Cl2, HCl and O2 gas, with an accompanying endothermic phase transition. We interpret this as the decomposition of Mg(ClO4)2, and subsequent reaction of HCl breaking down calcite to release CO2 (3).

**Figure 2.** Plot of Mass 44 (CO2) and temperature versus time for Phoenix calculated amounts of calcite and Mg-perchlorate. The first peak is due to the interaction of calcite and HCl gas (Cf. Figure 1).

Increases of Mg(ClO4)2•6H2O relative to CaCO3 cause more CO2 to evolve in the low-temperature range and less at higher temperature where calcite normally decomposes (Figure 3). This indicates that as Mg-perchlorate increases, more HCl becomes available to decompose more calcite at lower temperature. This has the effect of leaving less calcite to decompose at its relatively higher thermal decomposition temperature.

**Figure 3.** Plot of evolved CO2 at low-temperatures (orange) and high temperatures (red) versus amount of magnesium perchlorate hexahydrate (R² = 0.99 and 0.97, respectively).
The heated XRD results confirmed the reactions that produced the gas releases and thermodynamic transitions in the DSC/MS experiments. After dehydration, magnesium perchlorate decomposed to form MgO (periclase) (Figure 4). The evolved chlorine (or HCl) gas then attacked calcite, forming calcium chloride. The CaCl₂ decomposed to form exotic oxides of calcium and chlorine at higher temperatures, all of which eventually decomposed to form lime (CaO). It is unclear whether the Cl₂ by itself was able to react with CaCO₃ to make CaCl₂. In the DSC experiments, the temperature ramp was steep enough that there wasn't time for all the H₂O to be evacuated, allowing HCl to form. However, in the XRD experiments there was a period of hours between the dehydration of magnesium perchlorate and the formation of calcium chloride. We would not expect there to be enough water vapor in the sample chamber to produce HCl. Regardless of whether Cl₂ or HCl (or both) reacted with calcite, CO₂ would be released, which is what was observed in the mass spectrometry results.

Conclusions: The detection of HCl gas released from the thermal decomposition of magnesium perchlorate, and its reaction with calcite to produce calcium chloride and carbon dioxide is a clear demonstration of the thermal interaction between CaCO₃ and Mg(ClO₄)₂. This reaction occurs in the same temperature range as the unidentified low-temperature release of CO₂ observed in the TEGA and these results [14]. While it is true that perchlorate salts could hide the signature of organic compounds in the soil, there is no unequivocal evidence for organics from any mission to Mars thus far. Our findings have direct implications for the Curiosity rover, the centerpiece of the MSL mission. Curiosity contains the Sample Analysis at Mars (SAM) instrument package, which functions similarly to TEGA on Phoenix. The oven on SAM heats samples to 1000°C, and the evolved gases are analyzed by mass spectrometry. While it is unclear whether perchlorates or calcite exist at the Gale crater landing site, their combined presence would likely produce a similar signature to that found in TEGA, with two distinct CO₂ releases. Our work should caution any interpretations of similar findings as evidence for organic compounds on Mars.

IN SITU ANALYSIS OF ORTHOPYROXENE IN DIOGENITES USING LASER ABLATION ICP-MS.
M. Ek1, J. E. Quinn2, D. W. Mittlefehldt3, 1Department of Earth Sciences, University of Gothenburg, Sweden (mattias.ek@student.gu.se), 2Jacobs Engineering, ESGC, Houston, TX, USA, 3Astromaterials Research Office, NASA/Johnson Space Center, Houston, TX, USA.

Introduction: Howardites, eucrites and diogenites (HED) form a suite of igneous achondrite meteorites that are thought to have formed on the same asteroid, a strong candidate for which is 4 Vesta. Eucrites are basaltic in composition and consist primarily of pigeonite and plagioclase. They occur as both cumulate and non-cumulate textured meteorites. The diogenites consist primarily of orthopyroxene with varying amounts of olivine and accessory chromite and are cumulate rocks. More olivine-rich diogenites are sometimes referred to as olivine or harzburgitic diogenites. Howardites are brecciated mixtures of eucrites and/or diogenites and are thought to have formed on the HED parental body surface.

There are two major models for the formation of the HED parent body. One is the magma ocean model (e.g. [1]) in which the entire asteroid was one large continuously fractionated magma ocean with diogenites representing the lower crust and eucrites being upper crustal rocks. The second model hypothesizes that diogenites and eucrites were formed as a series of intrusions and/or extrusions of partial melts of a primitive proto-Vesta [2]. There are several variations of these end member models.

An important tool for understanding the formation of the HED meteorites is knowledge of the compositions of the melts from which they crystallized. There is no way to measure this directly for diogenites, but there are ways of extracting melt compositions from cumulate rocks that crystallized from them. One way is to use mineral compositions and partition coefficients to “reverse engineer” the parental melt.

One such recent study used diogenite- bulk rocks leached in acid to remove secondary phases and apatite [3]. These samples were analyzed by solution inductively coupled plasma mass spectrometer (ICP-MS) to look at Eu/Eu* anomalies in the rocks. (Eu* is the expected Eu content calculated from rare earth element (REE) diagrams.) These authors concluded that diogenites must have formed through contamination of their parent melts by partial melts of eucritic material, which would have the very low Eu/Eu* they infer for the diogenite parent melts. However, when interpreting bulk analyses, one has to be careful and take into account the possible effect of subsolidus reequilibration of the rocks [4]. One phenomena that could have a large effect on the composition of such samples is the difference in diffusive speed between Eu$^{2+}$ and the REE$^{3+}$ elements. Recent studies have shown that Eu$^{2+}$ has nearly twice the diffusive speed of the other REE in orthopyroxene [5]. This has the possible effect that due to reequilibration, the Eu/Eu* ratio of leached rocks would not be that of the original igneous cumulates.

We are evaluating laser ablation (LA) ICP-MS to measure in-situ concentrations of trace elements in orthopyroxenes. This technique has the advantage over whole rock methods in that we are able to look at possible variations both within a single grain and between several different grains. This gives us the opportunity to evaluate reequilibration effects in diogenites, and possibly to “look through” them to develop an understanding of their magmatic history.

Method: Backscatter electron image mosaics of thick sections (~150 µm) were collected using a JEOL 5940LV scanning electron microscope (SEM) at NASA Johnson space center (JSC). These images were used to plan elemental analyses of the samples. Major and minor element compositions of orthopyroxene and olivine grains were determined using a Cameca SX100 electron microprobe at JSC using a 20 kv accelerating voltage and 40 na beam current. Mineral standards were used for calibration and subsequently run as unknowns to ensure a good calibration.

In situ trace element compositions were analysed using a New Wave Research 193 nm solid state Nd:YAG laser ablation unit connected to a Thermo Scientific Element-XR magnetic sector ICP-MS at JSC. Analyses were carried out using a 100 micron
spot size, 10Hz repetition rate and 100% laser output. All elements were measured in low resolution mode and Ca43 was used as an internal standard to calculate absolute concentrations. The USGS standard glasses BIR1g, BVHO2g, and BCR2g were measured at least six times during each analytical run and were used to calculate calibration curves. The precision is estimated to be better than ±10% for all elements apart from Sr, La and Eu; due to their low concentrations their precision is estimated to be better than ±20%.

**Results:** Six diogenite thick sections from the US Antarctic meteorite collection were studied. Compositions of orthopyroxene and olivine grains were determined for core compositions and for zoning against adjacent minerals. These results were used to select targets for in situ analyses of trace elements. Brief petrological descriptions and analytical results follow.

**SAN 03473,13** is a brecciated diogenite containing small amounts of accessory silica and metal phases. Orthopyroxene has an Mg# of 76.8-77.4 and a Fe/Mn ratio of 26-28. The minor element variations are large compared to the other diogenites and the distribution is scattered and does not follow linear trends (Fig. 1) as is commonly observed in diogenites [6]. Trace elements show a similar scattering (Fig. 2). Most orthopyroxenes have a Eu/Eu* value of 0.18-0.23 with another possible cluster of 0.35-0.40.

**GRA 98108,30** is an unbrecciated diogenite with a high olivine content (~18%) and is classified as a harzburgitic diogenite. It has an orthopyroxene Mg# of 76.7. The minor and trace element variations are small and clustered (Fig. 1 & 2). It has a low Eu/Eu* ratio of 0.04-0.12, together with a strong LREE depletion.

**GRO 95555,47** is an unbrecciated diogenite containing small amounts of silica in close association with a metal phase (Fig. 3). Orthopyroxene has an Mg# of 75.5. Minor element compositions show significant linear variations (Fig. 1), with higher concentrations closer to the silica inclusions (Fig. 3). The trace elements show a similar trend. The Eu/Eu* ratio is 0.4-0.7 and the REE concentrations are similar to those from [3, 7].

**MIL 07001,40** is an unbrecciated harzburgitic diogenite. Orthopyroxene has an Mg# of 76.5. Olivine occurs as small rounded grains poikilitically included in orthopyroxene. There are small variations in the minor and trace element concentrations in orthopyroxene that follow a linear trend (Fig. 1 & 2). The REE concentrations and Eu/Eu* ratios of 0.5-0.7 are similar to those from [3].

**MET 00425,23** is a brecciated diogenite containing silica in contact with orthopyroxene. There are two populations of orthopyroxene based on the Mg#. One has an Mg# number of 83.5, and a low narrow range of minor and trace elements concentrations (Fig. 1 & 2). The other has a Mg# of ~81 with a large span of minor and trace element concentrations that follow linear trends (Fig. 1 & 2). The second population appears to be associated with orthopyroxene in contact with silica. The orthopyroxene Eu/Eu* ratios vary between 0.3-0.7.

**MET 00436,35** is an unbrecciated diogenite containing large chromite grains. The Mg# for orthopyroxene grains in this sample is low at ~70-71. Minor and trace element concentrations are low with narrow ranges (Fig. 1 & 2) and the Eu/Eu* ratio is 0.5-0.8.

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**Figure 2.** A) Yb vs. CaO for the laser ablation analyses. B) Yb vs. Eu for the analyses in A. The unfilled points indicate values from [3] for MIL 07001, MET 00425, MET 00436 and GRO 95555. Values from [7] are also included for GRO 95555.
Discussion: The major and minor element compositions of the diogenites analyzed here are within the ranges of other diogenites [6]. Increases in concentrations of minor and trace elements within GRO 95555 and MET 00425 are linked to their proximity to a silica phase (Fig. 3). This suggests that the silica phase and adjacent pyroxene was formed from a more compositionally evolved magma and/or interacted with interstitial melt. This is also suggested by the association of higher minor element concentrations with more Fe-rich orthopyroxene in MET 00425. The correlation of Al₂O₃ and CaO appears to be linear in all diogenites except SAN 03473, suggesting that the geologic history of this diogenite is more complex.

There is a clear relationship between the minor and trace element concentrations in diogenites, apart from Sr and Eu (Fig. 2). While these elements can be correlated with the minor element concentration within individual diogenites, there is no cross sample trend as observed for the other elements. This suggests that the diogenites did not form from a single fractionation sequence. This is also supported by the wide range of concentrations of trace elements observed. If all diogenites formed from the same fractionation sequence, then the magma chamber would have to have undergone 70-90% of perfect fractionational crystallization to create the range in incompatible trace element concentrations. This is inconsistent with the small range of Mg# observed in orthopyroxene; if a magma chamber undergoes 70-90% fractionation then one would expect a very low Mg# in orthopyroxene with high incompatible element contents. A more plausible scenario is that the diogenites formed from multiple magma chambers with different initial composition [8]. However, it is unclear how different diogenite parent magmas acquired vastly different trace element contents, although the magma contamination model [3] is one possibility.

The REE concentrations and Eu/Eu* ratios calculated for GRO95555, MIL07001 and MET00425 agree well with those obtained from previous studies [3, 7], although the concentrations obtained in this study are a factor of 1-5 higher than the previous studies for some diogenites. This difference may reflect heterogeneity in the diogenites (e.g. [9]). We find MET00436 is more LREE depleted than determined by [3] in addition to having up to a factor of 10 higher concentrations for the mid to high REE’s, again, suggesting possible heterogeneity.

Conclusions: In GRO95555 and MET00425 there is a link between high minor element concentrations in orthopyroxene and proximity to a silica phase. Orthopyroxene in these diogenites may have formed from an evolving magma or interacted with interstitial melt.

The trace element variations within orthopyroxene are several times higher than the minor elements; ~5x for Al vs. ~10x for Yb (Fig. 1 &2). The wide range in trace element contents would require that the magma chamber underwent more fractionation than is plausible based on Mg#. This favors the idea that diogenites may have formed from chemically distinct magmas rather than a single, asteroid-wide magma ocean.

In situ measurements of trace elements in orthopyroxenes gives results comparable to those obtained using bulk rock methods. While the precision of LA ICP-MS is not as good as solution analyses the ability to analyze small areas rather than bulk samples will prove invaluable when trying to decipher the magmatic history of diogenites. The continuing evaluation of the data and method will hopefully result in a better understanding for the formation of the HED meteorites.

References:
SPINEL-RICH LITHOLOGIES ON THE MOON: AN EXPERIMENTAL STUDY OF POSSIBLE PRECURSOR MELT COMPOSITIONS

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Introduction: The lunar crust contains some of the most important and accessible clues to the history of the Moon’s chemical evolution [1-3]. The M3 near-infrared (NIR) imaging spectrometer on the Chandrayaan-1 spacecraft discerned several areas on the rim of the far-side Moscoviense basin which had near-IR reflectance spectra characteristic of abundant (Mg,Fe)Al$_2$O$_4$ spinel with less than 5% mafic silicate minerals. Pieters et. al. [4] suggested that these deposits represent a previously unknown lunar rock type, spinel anorthosite. However, the origin and mechanism(s) of formation of such spinel-rich rocks are not clear.

Figure 1: Spinel-rich clast 2 in ALHA81005

Recently, a spinel-rich clast (clast 2; Fig. 1) in meteorite Allan Hills (ALH) A81005 has been reported; it contains ~30 vol% spinel, the most reported in any lunar sample [5]. Thus, it represents a great opportunity to study the formation of spinel-rich lunar lithologies. Gross and Treiman [10] suggested that this clast could have formed as a spinel cumulate from a picritic magma body that assimilated crustal anorthosite on its margins. This explains the petrographic and chemical features of the clast [10], and is consistent with the regional setting of the Moscoviense spinel deposit.

However, it remains possible that this clast could have formed as a spinel cumulate from an impact melt of troctolitic composition. Here we present data of our liquidus/crystallization experiments at low pressure to provide constraints on the origin and formation history of Mg-Al spinel-rich and spinel-bearing lunar highland samples. Experiments on the bulk composition of ALHA81005 (Table 1) will help us determine if melting of the meteorite source region can produce such spinel-rich rocks. Experiments on the bulk clast composition (Table 2) will reveal whether spinel-rich samples can be melt compositions themselves or if other concentration processes must have taken place in order to enrich the rock in spinel.

Table 1: Representative chemical analyses of experimental run products, ALHA81005 bulk composition.

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>1500°C</th>
<th>1400°C</th>
<th>1300°C</th>
<th>1200°C</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>45.50</td>
<td>44.85</td>
<td>45.00</td>
<td>43.86</td>
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<tr>
<td>TiO$_2$</td>
<td>0.26</td>
<td>0.26</td>
<td>0.40</td>
<td>0.02</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>25.00</td>
<td>26.08</td>
<td>19.28</td>
<td>35.53</td>
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<tr>
<td>FeO</td>
<td>5.11</td>
<td>3.01</td>
<td>7.73</td>
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</tr>
<tr>
<td>MnO</td>
<td>0.08</td>
<td>0.05</td>
<td>0.11</td>
<td>0.01</td>
</tr>
<tr>
<td>MgO</td>
<td>8.08</td>
<td>8.52</td>
<td>12.18</td>
<td>0.45</td>
</tr>
<tr>
<td>CaO</td>
<td>16.11</td>
<td>17.19</td>
<td>14.30</td>
<td>19.83</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.05</td>
<td>0.01</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
<td>Total</td>
<td>99.99</td>
<td>100.00</td>
<td>100.01</td>
<td>99.99</td>
</tr>
</tbody>
</table>

Table 1: Representative chemical analyses of experimental run products, ALHA81005 bulk composition.

<table>
<thead>
<tr>
<th>Starting composition</th>
<th>ALHA81005 Bulk</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weight %</td>
<td></td>
</tr>
<tr>
<td>SiO$_2$</td>
<td>44.70</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.250</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>26.60</td>
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<tr>
<td>FeO</td>
<td>5.50</td>
</tr>
<tr>
<td>MnO</td>
<td>0.075</td>
</tr>
<tr>
<td>MgO</td>
<td>7.90</td>
</tr>
<tr>
<td>CaO</td>
<td>15.00</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.20</td>
</tr>
<tr>
<td>Total</td>
<td>99.99</td>
</tr>
</tbody>
</table>

Samples: The meteorite ALHA81005 is a polymict anorthositic regolith breccia from the lunar highlands [6,7] (possibly from the farside [8,9]). Its clasts include granulites, ferroan anorthosites, mare basalts, impact melt, and impact glasses [5].

Clast 2, the spinel-rich clast, is approximately 350x150 μm in size and is exposed in thin section ALHA81005.9. The clast is a spinel-anorthositic-troctolite and contains ~30 vol% Mg-Al spinel and ~20 vol% mafic minerals, including olivine and low Ca-pyroxene [10].

Methods:

Experiments. One-bar liquidus/crystallization experiments were conducted in a Deltech gas mixing furnace (ARES Laboratory, JSC). The oxygen fugacity was regulated by mixing CO-CO$_2$ gas to a desired Moon-like $f_{O_2}$, at ~1 log unit below the iron-wüsite buffer. A synthetic powdered starting material with the bulk compositions of ALHA81005 and the clast 2 composition were made from homogenized mixtures of element oxides. The starting mixtures were pasted onto Re-wire loops (rather than Pt) to prevent Fe loss [11]. The experiments were held above the liquidus at 1500°C for several hours up to 1 day and then quenched at 1500°C, 1400°C, 1300°C, 1200°C, and 1150°C. The synthetic powder of the ALHA81005 bulk composition was fired and completely melted at 1-bar and quenched to a homogeneous anhydrous glass before the experimental
run. The glass was grounded to powder and remixed to ensure homogeneity.

The synthetic powder of clast 2 bulk composition could not be melted to a glass before the experimental run due to technical problems with the furnaces. Thus experiments on this composition were conducted directly from the oxide mixtures.

Phase abundances: To determine mineral abundances and ensure that no phases are missed during microprobe analysis, we performed mass balance calculations using the least squares method. Equilibrium-distribution-coefficients were calculated for olivine-glass (liquid), $K_{D_{FeO-MgO}} = \frac{[X_{FeO(Ol)}][X_{MgO(L)}]}{[X_{MgO(Ol)}][X_{FeO(L)}]}$.

EMP: Cameca SX100 electron microprobe analysis was used to obtain backscattered electron images and quantitative chemical analyses. Quantitative analyses were obtained by wavelength dispersive spectrometry. Operating conditions were 15kV accelerating voltage, and 20nA beam current. Analyzing the standards as unknowns ensured data quality.

Experimental Results:

**ALHA81005:** The results of the 1-bar experiments on ALHA81005 bulk composition show that spinel does not crystallize from the starting material. The bulk composition plots in the plagioclase field (Fig. 2). Thus, plagioclase is the first phase to crystallize; olivine is the second phase, and (presumably) pyroxene would follow at temperatures lower than in our experiments. Experimental run products include glass, plagioclase and olivine in different proportions, with 100 vol% glass at 1500ºC decreasing to 33 vol% glass at 1200ºC while plagioclase and olivine increase to 57 vol% anorthite and 10 vol% olivine at 1200ºC (Fig. 3).

**Clast 2:** The results of the 1-bar experiments on clast 2 show spinel crystallization at all experimental temperatures. The bulk composition plots outside the SiO₂-Olivine-Plagioclase diagram due to the enrichment in spinel. Spinel is the first phase to crystallize; its abundance reaches a maximum of 30 vol% at 1500ºC. Olivine is the second phase to crystallize followed by plagioclase at 1200ºC. However, backscattered images (BSE) of the experimental product show swirls and pockets of different melt compositions (Fig. 4). Calculated distribution-coefficients for olivine-glass show a wide range of $K_D = 0.27$ up to 0.90.

**Figure 2:** Phase diagrams for into the system olivine-anorthite-silica, projection after Longhi (1977, Fig. 2a). Liquidus boundaries shown for several values of Mg#. Data points are glass compositions for experiments on the ALHA81005 bulk composition.

**Discussion and Implications:** The observation of new lunar lithologies on the lunar surface is one of the more intriguing discoveries raising new questions about igneous processes in the lunar crust. Especially perplexing is the question of how spinel-rich samples such as the spinel-rich fragment in ALHA81005 [5, 10] and the spinel-rich deposits detected by M³ [4] fit into the petrogenetic schemes? Do they represent material that was
excavated by an impact from deep within the crust or do they represent crystallized material from an impact melt sheet?

**ALHA81005:** The calculated distribution-coefficients for olivine-glass (liquid) fall within the expected range for equilibrium (0.3-0.35), thus it is safe to assume that they were in equilibrium with the melt [12,13]. However, the experiments show that melting and crystallization of the meteorite composition, and hence its presumed source region (e.g. as in an impact melt sheet), will not produce any spinel and thus will not lead to spinel-bearing or spinel-rich lithologies.

**Clast 2:** Using the oxide mix as a starting material rather than a homogenized glass mix is not ideal, as our results show. The wide range (0.27-0.90) of the Mg/Fe distribution-coefficients for olivine-glass indicate that equilibria for these experiments were not achieved. Thus we cannot determine whether clast 2 could be a melt composition itself, or if other processes must have taken place in order to enrich this sample in spinel. The maximum spinel crystallization of 30 vol% has to be taken in with care. It could be possible that some oxide grains - e.g. Al₂O₃ - served as nucleation grains to crystallize the spinel.

However, the bulk composition plots within the spinel field (though outside this projection), and it is not surprising that spinel is the first phase to crystallize. The next phase to crystallize is olivine at 1400°C, indicating the melt composition follows the proposed line on Figure 2. Thus spinel crystallization is likely.

Further experiments using the homogenized glass powder will need to be run in order to determine how much volume % spinel can crystallize. Only then can we determine whether accumulation is a major process in the formation of spinel-rich lithologies.

**Figure 3:** Variation in phase abundances (wt%) for ALHA81005 bulk composition starting mixture at 1 bar as a function of temperature, showing the crystallization of plagioclase and olivine

**Figure 4:** BSE images of melt at 1200°C and 1150°C showing the varying glasses as well as the spinel, olivine and plagioclase phases.

**Acknowledgments:** We are especially grateful to L. Le for her patience and assistance with the 1-bar furnace experiments.

TRACKING THE PROCESS OF VOLATILE RELEASE FROM THE LUNAR HIGHLAND BRECCIA METEORITE NWA 2996 USING VESICLE SIZE DISTRIBUTIONS. S. R. Jacob$^{1,2}$ and C. N. Mercer$^1$. $^1$Lunar and Planetary, 3600 Bay Area Blvd., Houston, TX 77058; Jacob@lpi.usra.edu, $^2$University of Hawaii at Manoa, 1680 East-West Rd., Honolulu, HI 96822; srjacob@hawaii.edu.

Introduction: Vesicles are frozen records of degassing processes in melts. Textural studies of vesicles in pyroclasts and lava flows are commonly used to understand processes of gas exsolution that drive terrestrial magmatism [1]. In this study we apply the same techniques to further understand volatile escape from lunar materials.

Relative rates of bubble nucleation and growth control primary vesicle textures, which can later be modified by vesicle coalescence, deformation, or volatile escape. By measuring characteristics of vesicle populations such as size, spatial pattern, numbers density, and shape we can determine the physical processes by which volatiles escape. For example, these characteristics can tell us if the growth and nucleation of vesicles in a sample occurred as a single event or in multiple stages, whether vesiculation occurred continuously or accelerated through time, and whether the vesicles coalesced, collapsed, or underwent Ostwald ripening [2, 3].

Understanding vesicle textures in lunar materials provides a window into understanding the role of volatiles in lunar materials. This is particularly important since recent studies show that the moon is less dry than previously thought [4, 5, 6]. Volatiles exist at the surface and in the interior of the Moon. Solar-implanted volatile elements at the surface may supply volatiles (and produce vesicles) in impact-generated materials. In contrast, volatile elements from the interior are expected to supply volatiles (and produce vesicles) in volcanic-generated products like basalt flows [7]. Lunar volatiles include H, CO₂, F, S, and Cl [7].

The goal of this project is to quantify the vesicle textures and glass compositions in lunar meteorite NWA 2996 to better understand the process of volatile release from the matrix melt during meteorite formation, and to compare the vesicularity with other lunar materials. The NWA 2996 thin section, TAL-1*, is a feldspathic breccia that contains large clasts of anorthosite, non-mare mafic-rich clasts, and impact melt breccias with minor evolved lithologies basalt, glass beads [8, 9,10], and has an extensive glassy matrix that is dense in some areas but is highly vesicular in >50% of the thin section.

Methods: We used vesicle size distribution theory [11] to quantify the vesicle textures in NWA 2996. SEM backscattered mosaic images (Fig. 1a) were acquired at the Johnson Space Center using a JEOL field-emission JSM7600F SEM. All images were simplified (Fig. 1b) using Adobe Photoshop and ImageJ following the processing methods of Shea et al. [2], whereby each phase (vesicles, matrix glass, and minerals) were assigned a single grayscale value.

The simplified image was imported into the Matlab-based program Fast Object Analysis and Measurement System (FOAMS) [2]. We used FOAMS to calculate and tabulate vesicle size and shape parameters (e.g., number of vesicles, diameter, area, perimeter, aspect ratio, etc.) as well as the 2-D vesicle number density, $N_a$ (number of vesicles of a given size per unit area). FOAMS employs a stereological conversion from [12] to convert $N_a$ to a more accurate 3-D vesicle number $N_v$ (number of vesicles of a given size per unit volume).

The major and minor element compositions of matrix glass, dense glassy clasts, and glass beads were determined using a Cameca SX100 electron microprobe at the Johnson Space Center. The sample
was analyzed with a 15 keV accelerating potential, 10nA beam current, and a 10 µm spot size, using natural and synthetic standards.

**Results and Discussion:** The calculated bulk vesicularity of TAL-1* is ~9.4%. Currently there are no other quantitative reports of vesicularity in lunar feldspathic breccias with which to compare this sample. However, bulk “porosity” of many lunar materials have been reported and represent a maximum estimate for vesicularity, as these measurements include porosity due to vesicles and cracks. Feldspathic breccias range in bulk porosity from <1% to 36%, while impact melt breccias (basaltic and feldspathic) range from 2% to 28% [13, 14]. In the context of bulk vesicularity, NWA 2996 is only moderately vesicular (Fig. 2). However, the volatiles that created the vesicles exsolved from the shock melted matrix and not from entrained clasts, so considering a corrected “matrix-only” vesicularity of 20% would indicate that NWA 2996 is moderate-to-highly vesicular relative to other feldspathic and impact melt breccias (Fig. 2). Vesicles in NWA 2996 document degassing of volatiles that are likely derived from shock melted rock and lunar surface volatiles. For comparison, we report values of vesicularity for the most highly vesicular Apollo basalts (Fig. 2), which derived their volatiles from the Moon’s interior. These basalts have higher vesicularities than most impact-generated materials.

The volume fraction size distribution (Fig 4) shows that vesicles with diameters between ~0.02 to 0.08 mm account for the majority of the overall sample vesicularity.

The single-mode Gaussian shape of the distribution implies that nucleation and growth of vesicles occurred in a single pulse, as expected for a single shock metamorphic event. The slight negative skew of the distribution indicates that Ostwald ripening occurred [2, 3]. Ripening takes place as volatiles transfer from one vesicle to another through thin vesicle walls. During ripening, large vesicles continue to grow as small vesicles shrink.

The vesicle size distribution for NWA 2996 can be represented as a population density function [16, 17]. For a steady-state population of nucleating and growing bubbles:

\[ n = n^0 \exp \left(-\frac{L}{G\tau}\right) \]  

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**Figure 2:** Comparison of porosity of lunar products (small symbols) [13, 14, 15] with vesicularity measurements from this study (large symbols).

**Figure 3:** Aspect ratio histogram of NWA 2996.

**Figure 4:** Histogram of the volume fraction size distribution. Bins are geometric, the first bin represents the smallest size vesicle in mm and each successive bin is multiplied by 100.1.
where $n$ is the volumetric number density of vesicles per vesicle size class (number per volume melt per bubble diameter), $n^0$ is the volumetric number density of bubble nuclei, $G$ is the mean bubble growth rate, and $\tau$ is the time scale of nucleation and growth. A plot of $\ln(n)$ versus $L$ has a slope of $-1/G\tau$ and an intercept of $n^0$. Figure 5 shows the vesicle size distribution for NWA 2996. The distribution is concave up for smaller vesicle size classes, signifying continuous acceleration of nucleation and growth of bubbles. For comparison, a linear distribution would indicate a constant rate of nucleation and growth.

Figure 5: Vesicle size distribution of NWA 2996.

The abrupt shallowing in slope at ~0.15 mm shows that the product, $G\tau$, abruptly increases, but then abruptly decreases again at 0.2 mm. The abrupt changes in slope at these larger vesicle diameters are likely the result of large bubble coalescence.

Microprobe analyses show both the matrix glass and dense glass clasts are compositionally similar to three lithologies represented in NWA 2996 clasts: anorthosite, non-mare mafic-rich clasts, and basalts (Fig. 6). Glass bead compositions are more restricted and represent shocked anorthosite. Maskelynite swirled into the matrix glass represents shocked Ca and Na-plagioclase grains with variable meteoritic Fe. Future SIMS analyses will reveal whether the volatiles contents are similar amongst these different glassy products. The target lithologies that generated these glasses are anhydrous, however, the relatively high matrix vesicularity indicates volatiles existed. This implies that the target site for NWA 2996 may have been volatile-rich, supplying the volatiles needed to generate vesicles. However, we cannot test this hypothesis until there are more data on vesicle size distributions and volatile contents in feldspathic breccia samples from mature lunar regolith.

Figure 6: Mg # vs. An #, shaded region indicates Apollo lunar glass samples [18].

**Conclusion:** NWA 2996 has a moderate to high vesicularity compared to other lunar feldspathic breccias, but is less vesicular than the most vesicular lunar basalts. Vesicles nucleated and grew in a single, accelerating pulse, and experienced Ostwald ripening, coalescence, and post-formation deformation. Glass compositions indicate formation by shock melting of anhydrous lithologies, implying that addition of surface volatiles are likely required to explain the vesicularity.

**Acknowledgements:** The NWA 2996 thin section was graciously loaned by Dr. T Bunch of Northern Arizona University. This work was funded by the Lunar and Planetary Institute summer internship program. Thank you to Ramiro Padilla and Brian Fessler for help with computer software.

**References:**

THE RELATIONSHIP BETWEEN RADAR SCATTERING AND SURFACE ROUGHNESS AT LUNAR VOLCANIC DOMES.
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Introduction:
Understanding the morphology of volcanic terrains on the Moon is essential in discovering their emplacement mechanisms. Terrestrial lava flows have been studied using radar imaging in an attempt to distinguish differing flow morphologies. SIR-B radar data taken over the Kilauea Volcano in Hawaii have shown it is possible to distinguish between rough a'a and smoother pahoehoe flows, as well as pyroclastic ash deposits [1]. It is likewise possible to use radar to distinguish different volcanic features on the Moon. Rugged lava flows on the Moon such as that in Mare Serenitatis have been identified using Earth-based radar, and by observing areas of higher backscatter and circular polarization ratio (CPR) it is possible to infer their original flow morphology [2]. Ground-based radar has also been used to identify fine-grained pyroclastic deposits by their low backscatter and CPR values in Mare Vaporum and southern Mare Serenitatis [3]. In this work, we explore the relationship between surface roughness and the radar signatures of volcanic domes, lava flows, and pyroclastic deposits on the Moon.

The largest concentration of lunar domes is the Marius Hills Complex in Oceanus Procellarum, containing over 200 domes of varying sizes. Based on infrared spectroscopy, these domes appear to be of similar composition to the surrounding mare [4]. The steep slopes of many Marius Hills domes are thought to be caused by eruptions of cooler, more crystalline basaltic magma, as opposed to a more evolved, silicic melt [5]. Previous studies of these domes show that they have a much higher CPR value as compared to the mare plains, suggesting that the domes are rougher than the surrounding terrain [5]. In order to quantify the roughness of the Marius Hills domes, we have compared radar CPR values at S- and P-band wavelengths (12.6 and 70 cm, respectively) with RMS topographic variation obtained from LOLA data. Comparing these different parameters will tell us about the domes’ roughness on various length scales.

For comparison with the Marius Hills domes, several other regions on the Moon were analyzed using this technique: the Gruithuisen and Mairan domes [6], as well as the Rümker Hills volcanic complex [5]; mare basalt lava flows in Imbrium, Oceanus Procellarum and regions of the Marius Hills plains; regions of pyroclastic dark mantle deposits [3,7]; rough lava flow features from Mare Serenitatis [2]; and crater features such as ejecta blankets and crater floor impact melts.

Methods:
LOLA: RMS Topographic Roughness and Pulse Width.
The Lunar Orbiter Laser Altimeter measures the Moon’s topography. Each shot of the surface consists of five spots arranged in a cross pattern [8]. Each sequential shot containing the five spots is separated by ~57 m, while the smallest point-to-point baseline is ~25 m [8]. We used the LOLA data to calculate RMS topographic roughness to compare with CPR. In order to do this, we wrote a script in MATLAB which grouped three LOLA shots together and used the spot data to create a least-squares plane. In this script, all invalid data points were removed from the data set [9]. To ensure robust least-squares planes, we only generated RMS misfits if there were at least nine functioning spots; if all spots were receiving data, there were a total of fifteen points. Our results typically included 14-15 points on each calculated plane. We used this plane to calculate RMS misfit of each data point from the plane, which represents the roughness of the lunar surface on a 25-50 m scale.

The LOLA altimeter also measures the time interval of each returned laser pulse, referred to as the pulse width [9]. The MATLAB script also calculated average pulse width for each group of three shots. Pulse width is controlled by the small scale topographic roughness within each 5 m diameter spot, with rougher topography causing the returned pulse to be broadened in time. The pulse widths are the averages for each study area and are reported in terms of the measured time interval in nanoseconds. An approximate scaling analysis shows that a pulse width of 15 nanosec corresponds to a characteristic roughness of 20 cm inside the laser spot, and that a 25 nanosec pulse width corresponds to a characteristic roughness of 70 cm. A more complete analysis, paralleling a study of martian altimetry [10], may modify these values somewhat, but the basic relationship between increasing pulse width and increasing surface roughness is robust.

Mini-RF: Circular Polarization Ratio.
The radar systems used in this study transmit in circular polarization. When an incident beam reflects off a surface, it changes polarity to the opposite sense of
that which was transmitted (OC); likewise, if the beam reflects off two objects, it reverses polarity twice, resulting in the same sense in which it was transmitted (SC). The Circular Polarization Ratio (CPR) is the ratio of the same sense to opposite sense of received backscatter, relative to the transmitted sense (SC/OC) [5]. Rough, blocky surfaces have the potential to cause incident radar beams to bounce more than once, leading to the conclusion that high SC signals, therefore high CPR, are caused by rough surfaces. Depending on the wavelength of the radar and the composition of the regolith, it is possible to observe features below the surface [5]. The probing depth of the radar is a function of wavelength – S-band wavelength (12.6 cm) can penetrate up to a meter or two in typical mare regolith, while P-band wavelength (70 cm) is sensitive to diameters ~2 cm and larger in diameter, while P-band signals are sensitive to diameters ~10 cm and larger [5]. By comparing the roughness on small scales from CPR to larger-scale roughness from RMS misfit and pulse width, we hope to gain an understanding of the large- and small-scale topography of the Marius Hills.

S-Band CPR values were collected for the Marius Hills domes as well as our areas of comparison. Using the image processing programs ISIS and ImageJ, we extracted S-Band CPR from observations made by the Mini-RF radar on Lunar Reconnaissance Orbiter [11]. The radar results were measured for the same study regions as the altimetry. Our data collection for P-Band CPR [2,5] is ongoing, and will provide a measure of roughness at larger scales.

Results:

Our results are summarized in Figures 1 and 2. Each symbol is the average of all results calculated in a 6x6 km study area. We typically included at least three study averages for each geologic region.

Figure 1 shows the relationship between the RMS topographic roughness, which measures roughness at 25-50 m scales, and the pulse width, which measures topographic roughness at scales less than 5 meters. Not surprisingly, the crater units (floor and ejecta blanket of Tycho and Copernicus) are roughest at both scales. The mare lava flows and the dark mantle pyroclastic deposits are smoothest on both scales. Interestingly, the Gruithuisen and Mairan domes are distinctly rougher than the other volcanic domes in the pulse width data, possibly reflecting a different emplacement mechanism and different magma composition.

Typically, large values of CPR correspond to relatively rough topography (e.g., [5]). The relationship between CPR and topographic roughness holds at a length scale of a few times the radar wavelength (~1 meter for the S-band CPR) and thus should be most testable in our pulse width data. A comparison between S-band CPR and pulse width shows a weak positive correlation, but the comparison is complicated by the currently incomplete quantification between pulse width and short length scale roughness. Our longer length scale topographic roughness is more quantitatively reliable, and Figure 1 indicates that there is a correlation between topographic roughness as defined by pulse width and by RMS roughness. Therefore, in Figure 2 we show the relationship between the S-band CPR and the RMS topographic roughness. Most of the patterns seen in Figure 1 also are evident in Figure 2. In particular, the mare lava flows and the dark mantle pyroclastic deposits are the topographically smoothest units and also have the lowest CPR values. The roughest unit is the crater materials, which also has the largest CPR values. These patterns are consistent with the expected relationship between topographic roughness and CPR. On the other hand, the Marius Hills and Gruithuisen domes cannot be distinguished in the S-band radar, despite their topographic roughness differences.

Future Work:

This project is to be continued as a Senior Thesis at Mount Holyoke College. P-Band CPR values are currently being collected, and will be integral in comparing roughness features at larger scales. Likewise, X-Band CPR values may be collected to provide small-scale roughness measurements. Additionally, this method of analysis will be applied to lava flows of different ages in order to observe the evolution of regolith development on lava flow surfaces with time.

References:

Figure 1. Mean Pulse Width plotted against RMS Topographic Roughness

Figure 2. S-Band (12.6 cm wavelength) Circular Polarization Ratio plotted against RMS Topographic Roughness
CORE-MANTLE PARTITIONING OF VOLATILE ELEMENTS AND THE ORIGIN OF VOLATILE ELEMENTS IN THE EARTH. K. Nickodem¹, K. Righter², L. Danielson², K. Pando², C. Lee³ ¹ Department of Civil Engineering and Geological Sciences, University of Notre Dame, Notre Dame, IN 46556 (nickod1@nd.edu) ²Johnson Space Center, Houston, TX 77058 ³Department of Earth Science, Rice University, Houston, TX 77005

**Introduction:** There are currently several hypotheses on the origin of volatile elements in the Earth. One hypothesis is that they were added during Earth’s accretion and core formation and mobilized into the metallic core [1], others claim multiple stage origin [2], while some hypothesize that volatiles were added after the core already formed [3].

Several volatile elements are depleted in Earth’s mantle relative to the chondrites, something which continues to puzzle many scientists. This depletion is likely due to a combination of volatility and core formation. The Earth’s core is composed of Fe and some lighter constituents, although the abundances of these lighter elements are unknown [4]. Si is one of these potential light elements [5] although few studies have analyzed the effect of Si on metal-silicate partitioning, in particular the volatile elements.

As, In, Ge, and Sb are trace volatile elements which are depleted in the mantle but have yet to be extensively studied. The metal-silicate partition coefficients of these elements can be measured to determine the effect of Si. Partition coefficients depend on temperature, pressure, oxygen fugacity, and metal and silicate composition and can constrain the concentrations of volatile, siderophile elements found in the mantle.

Reported here are the results from 13 experiments examining the partitioning of As, In, Ge, and Sb between silicate and metallic liquid. These experiments will examine the effect of temperature, and metal composition (i.e., Si content) on these elements in order to gain a greater understanding of the core-mantle separation which occurred during the Earth’s early stages. The data can then be applied to the origin of volatile elements in the Earth.

**Procedures:** The samples used for the series of experiments were powders composed of 70 wt.% Knippa Basalt, composition described in Lewis et al. [6], 30 wt.% metal mixture, and varying amounts of Si ranging from 0 wt.% Si to 10 wt.% Si. The metal mixture contained 80.9 wt.% Fe, 8.2 wt.% FeS, 2.4 wt.% Ge, 3.2 wt.% As₂O₃, 2.9 wt.% Sb₂O₅, and 2.4 wt.% In. These were ground into a powder and mechanically mixed. All experiments were run using graphite capsules. The runs were conducted using a piston cylinder apparatus at constant pressure of 1.0 GPa with various times and temperatures. Once at constant pressure, samples were heated to high enough temperatures to melt and attain equilibrium for durations based on diffusion time across the capsule [7]. The temperature was measured using a type C thermocouple (W-Re) wires with an accuracy of ±2°C. Samples were then quenched to a silicate glass with large metallic spheres (Fig. 1) by turning off the power and keeping constant pressure until the temperatures reached 100°C.

Four series were performed (Table 1): two metal-silicate composition series with varying amounts of silicon performed at 1600°C and 1800°C, an Si free time series at 1600°C, and an Si free temperature series ranging between 1500°C and 1800°C.

**Analysis:** Samples were analyzed for major element composition using a Cameca SX100 for electron microprobe analysis (EMPA) at NASA-JSC. A 1 µm beam was used at 20kV and 10nA. A variety of natural and synthetic standards were used. All samples used graphite capsules and were carbon saturated, but carbon was not analyzed yet. Therefore the metal analysis have low totals. Based on previous studies, the carbon solubility within the metal can be up to 5.5 wt.% C between 1300°C and 1800°C [8].

For all samples the In, As, Sb, and Ge content of the glass was lower than the detection limit of the EMPA; therefore, the samples were analyzed for trace element composition using Laser Ablation Inductively Coupled Mass Spectrometer (LA-ICP-MS) at Rice University. Analysis was performed at Low Resolution (LR) and normalized to ⁴⁷Ca isotope. Isotopes ⁷⁵As, ¹¹⁵In, ⁷³Ge, ⁷⁷Ge, and ¹²¹Sb were the only trace elements specifically studied for this research.
Table 1. Summary of all experiments. Performed at 1.0 GPa. Time series omitted from table a. Duplicate experiment performed but unable to be analyzed in time for this report

<table>
<thead>
<tr>
<th>Sample</th>
<th>KAN150</th>
<th>KAN160</th>
<th>KAN170</th>
<th>KAN180</th>
<th>KAN160a</th>
<th>KAN170a</th>
<th>KAN180a</th>
<th>KAN160b</th>
<th>KAN180b</th>
<th>KAN1610B</th>
<th>KAN1810B</th>
<th>KAN184</th>
<th>KAN184B</th>
<th>KAN1810a</th>
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<tr>
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<td>1600</td>
<td>1700</td>
<td>1800</td>
<td>1600</td>
<td>1600</td>
<td>1600</td>
<td>1800</td>
<td>1800</td>
<td>1800</td>
<td>1800</td>
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<td>1800</td>
</tr>
<tr>
<td>Duration</td>
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<td>45</td>
<td>15</td>
<td>90</td>
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<td>90</td>
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<td>90</td>
<td>90</td>
<td>90</td>
<td>90</td>
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<td>0</td>
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<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>wt. Si</td>
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<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>ΔIW</td>
<td>-1.32</td>
<td>-1.25</td>
<td>-1.37</td>
<td>-1.24</td>
<td>-1.25</td>
<td>-5.26</td>
<td>-7.47</td>
<td>-1.24</td>
<td>-3.21</td>
<td>-4.89</td>
<td>-4.44</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

**Results:** An Si free time series at 1600°C was completed but not reported in the table. Samples were run for 30, 60, 90, and 120min. Within error measurements, results indicate that equilibrium is attained within 30min after which metal and silicate compositions do not change. Once chemical equilibrium is reached longer runs help to reach textural equilibrium allowing for larger metallic spheres to form.

Oxygen fugacity was calculated relative to the Iron-Wüstite (IW) buffer using ΔIW = -2log[XFeO/XFe]. The ΔIW values ranged from ~-1.3 to -1.37 for Si free runs compared to Si bearing runs which produced ΔIW values from -4.9 to -7.5 (Table 1). The partition coefficient can change along with oxygen fugacity (fO2). An increase in Si content will cause a decrease in fO2. The range of ΔIW values for these experiments falls in the range considered during Earth’s core formation (~1 to -5).

Partitioning behavior of As, Sb, Ge, and In can be calculated in two ways. One way is to calculate the partition coefficient for each element using the equation:

\[ D_{\text{metal/sil}} = \frac{c_{\text{metal}}}{c_{\text{silicate}}} \]

where \( D \) = partition coefficient of each element and \( c \) = wt. % concentration of the element in the metal and silicate respectively.

The second way is to calculate the KD using an exchange reaction of element \( i \) with valence \( n \):

\[ iO_{2/2}\text{silicate} + \frac{(n/2)}{2}\text{FeO}^{2-}_{\text{metal}} = \frac{(n)}{2}\text{Fe}^{2+}_{\text{metal}} + \frac{(n/2)}{2}\text{FeO}^{2+}_{\text{silicate}} \]

The KD was calculated for each sample using the equation:

\[ K_{D} = \frac{[XFeO]^{2-}_{i} [XFe]}{[XFeO]^{2+}_{i} [XFe]} \]

Where \( K_{D} \) = exchange coefficient, \( X_{i} \) and \( X_{FeO} \) = molar concentration of element \( i \) in the metal and metal oxide of \( i \) in the silicate respectively. The advantage of \( K_{D} \) is that it is independent of \( O_{2} \).

The results show a general decrease in the KD with increasing temperature and Si content for all elements (Fig. 2, 3). For Samples KAN160, KAN180B, KAN184, and KAN1810 there was significant BaO contamination in the silicate melt which could affect some of the concentrations. Results reported here have removed the BaO and renormalized. Duplicate runs have been completed but not analyzed in time for this report.

Data on As and Ge from Siebert et al [9] is included on the temperature graph (Fig. 2). The experiments performed by Siebert et al. were completed at 3.0 GPa. Comparison of the results to Siebert et al. indicate similar log KD values for As and Ge [9]. The data from Siebert et al. is consistent with the trend that log KD values decrease with increasing temperature. The 1600°C Si-bearing series shows decreasing log KD values for each element. The 1800°C series, also show a lowering of KD, but there is more scatter.
The new experimental data are compared to the equilibrium log $K_D$ values in Figure 3. The necessary log $K_D$ value for the elements based on chondritic ratios can be attained when additional Si is added to the composition. This indicates that Si has a large effect. The combined effect of temperature and metallic Si content shows that even at 1600°C the necessary concentration of these metals found in the primitive upper mantle can be met.

The large effect of silicon on the magnitude of the $K_D$ values is expected based on the work of Chabot et al. [12] who demonstrated that As, Sb, and Ge all prefer Si-free metal to Si-bearing metal. However, another factor contributing to the low $K_{Ds}$ is the sulphur present in the silicate melt. With increasing Si in the metal (and lower $fO_2$), more S dissolves into the silicate melt [13]. Some of the trace elements may then have a stronger affinity for S (As and Sb), increasing the concentrations in the silicate melt. The large effect on the partition coefficient seen in these results may be a combination between increased Si content of the metal as well as S content in silicate melt. Further calculations and data will be needed to understand this interaction between these elements and to know the individual effect of Si.

Evidence here suggests that Si has a large effect on the partitioning of As, Sb, Ge, and In. Based on the data, if Si is present in the core then the concentrations of these elements found in the mantle can be explained by early stage equilibrium between the core and the silicate melt, supporting the hypothesis that volatiles were added during Earth’s accretion and core formation.

**Discussion:** To test whether these four elements can be explained by an equilibrium core-mantle partitioning scenario, one must know the values of log $K_D$ that would be required for such a scenario. To calculate the $K_D$ or $D$ required for equilibrium in the early Earth, it is necessary to determine the approximate abundance of the metals in the core. The bulk composition of early Earth will be assumed to be the same as CI chondrites. The target log $K_D$ was calculated (Table 2) using the abundances in the upper mantle [10] and CI chondrites adjusted for volatility factors [11]. This assumes a 32 wt.% core and 68 wt.% mantle.

<table>
<thead>
<tr>
<th>Element</th>
<th>Volatility</th>
<th>Core</th>
<th>Mantle</th>
<th>Target D</th>
<th>Target log $K_D$</th>
</tr>
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<tr>
<td>As</td>
<td>3</td>
<td>5</td>
<td>1.1</td>
<td>100</td>
<td>-1.34</td>
</tr>
<tr>
<td>Sb</td>
<td>3</td>
<td>0.13</td>
<td>0.05</td>
<td>21.7</td>
<td>-0.75</td>
</tr>
<tr>
<td>Ge</td>
<td>4</td>
<td>20</td>
<td>0.011</td>
<td>18.2</td>
<td>0.038</td>
</tr>
<tr>
<td>In</td>
<td>10</td>
<td>0.0016</td>
<td>0.006</td>
<td>0.15</td>
<td>-2.85</td>
</tr>
</tbody>
</table>

**References:**
Understanding The Scattering Properties of Lunar Geological Units Using Mini-RF and LROC Data  S.M. Pantone¹, P.D. Spudis² ¹.Dept. of Geology, Weber State University, Ogden, Utah 84408 (SpenserPantone@weber.edu), ². Lunar and Planetary Institute, Houston, Texas 77058

Introduction

The lunar surface is constantly bombarded by meteorites. The bombardment process excavates bedrock from below, erodes and blankets existing blocks on the surface. Eventually, the surface becomes smoother with time, so that older surfaces display fewer blocks than younger ones. A better understanding of the backscattering properties of craters with varying age and size is crucial for interpreting data received from the Mini-RF instrument onboard the LRO (Lunar Reconnaissance Orbiter) spacecraft [1,2].

A major objective of the Mini-RF experiment is to distinguish lunar surfaces that may contain water/ice deposits from lunar surfaces that do not contain water/ice deposits [2, 3]. The Mini-RF transmits a circularly polarized RF electromagnetic energy and coherently receives orthogonal linear polarization echoes [1]. The Mini-RF maps in two separate bands (λ=12.6 and 4.5 cm) at a high resolution mode of 30 m/pixel [1]. Given the variables mentioned, the four stokes parameters are reconstructed. The Circular Polarized Ratio (CPR) is calculated for the purposes of understanding subsurface and surface roughness. The CPR is determined from reflections acquired from the ratio of power of the transmitted radio wave in same sense to the reflected radio wave in the opposite sense [1].

Radar has the ability to penetrate further into the ground were subsurface water/ice is believed to occur in various permanent shadowed regions (PSRs) near the poles of the Moon [6]. Ice in the PSRs would be transparent to radar, but the inclusions of materials and imperfections would cause the radio wave to reflect multiple times [3], enhancing the number of same sense reflections and increasing the CPR. In addition, ice also displays the coherent backscatter opposition effect (CBOE), an interferrometric addition of same sense backscatter that further increases CPR in ice targets [7]. High CPR values correlate to multiple reflections and are typically associated with very rough surfaces (Fig.1). The average dry lunar surface has a CPR in the range of 0.2-0.4 at 48° incidence [3].

(Fig.1) Linne crater in Mare Serenitatis from Mini-RF. CPR Regions are color-coded in image on right.

Methods

In this study, we have taken very high resolution images of craters for a variety of ages and degradational states and counted the number of blocks on the surface. These counts are expressed as an average areal density of rocks of a given size. Because the resolution of the Lunar Reconnaissance Orbiter Camera Narrow Angle Camera images (0.5 m/pixel) is coarser than the blocks to which Mini-RF are sensitive (~10 cm), We took the cumulative frequency of blocks, made a best-fit estimate of that function, and computed the theoretical block density at 10 cm. These values were then compared directly to CPR data from the Mini-RF for the same area to derive any dependence of CPR on wavelength-scale surface roughness.

High resolution images of 2-20 km diameter craters taken by the LROC NAC in non-polar regions with a variety of ages were downloaded from the Planetary Database System (PDS). The images were made into map projections using ISIS (Integrated Software for Imagers and Spectrometers) from the USGS [8]. In
these images, subareas were selected for analysis. The subareas coincide with the crater floor, wall, rim, and beyond (Fig. 2a). The subareas were selected to give the best representation of a given region (Fig. 2a). The subareas were cropped from the main images and converted into PNG files. After cropping and converting the subareas for each crater, the PNG files were opened with the Feature program (IDL; [9]), which allows for block counting and measuring. Blocks that could be resolved were counted and measured along the longest axis. The feature program records every measurement and block counted. The data was then imported into Microsoft Excel program, where block size was converted from pixels to dimension in meters. These data were plotted per unit area and least-squares fits were calculated. The trend of the data agrees with the general consensus that the block distribution follows a power law distribution [4, 5] (Fig. 2b).

(Fig.2a) The lunar crater Linne showing subareas used for analysis [11].

(Fig.2b) Block size frequency of subarea 1 from crater Linne.

(Fig.3) Histogram of CPR data from crater east of Lassell C.

Table 1 Boulder abundances and CPR values for selected lunar craters.

<table>
<thead>
<tr>
<th>Name</th>
<th># of boulders above 10 cm</th>
<th>CPR</th>
<th>Sigma</th>
<th>lat/long</th>
<th>D(km)</th>
<th>Age</th>
</tr>
</thead>
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<td>Byrgius A</td>
<td>83.12</td>
<td>1.0</td>
<td>0.5</td>
<td>(25.4, 43)</td>
<td>19 C</td>
<td></td>
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<td>Byrgius A</td>
<td>0.08</td>
<td>1.0</td>
<td>0.6</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Byrgius A</td>
<td>7.28</td>
<td>0.9</td>
<td>0.5</td>
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<tr>
<td>Byrgius A</td>
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<td>0.4</td>
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<td>Euclidides</td>
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<td>0.5</td>
<td>(7.3, 29)</td>
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<tr>
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<tr>
<td>Dolland</td>
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<td>0.5</td>
<td>0.3</td>
<td>[14, 44, 10, 49]</td>
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<td>Bode</td>
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<td>0.4</td>
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<td>[-9, 24, 13]</td>
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<td>0.4</td>
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<tr>
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<td>0.2</td>
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</table>

The Mini-RF CPR data for these same subareas were analyzed using Image J software [10]. Histograms were calculated from the subareas that were cropped for block counting (Fig. 3). The histograms display mean CPR values and standard deviations from the mean values. These values are recorded in Excel and plotted with the combined data from block counting and histograms in the form of, CPR vs. log no. of boulders above 10 cm/100 m² (Fig. 4, Table 1).
Combined data from all craters analyzed by this study and [11].

**Results**

When the data was plotted, the expected trend was that older craters would plot in low CPR value ranges due to low block abundance. Young fresh craters would plot in high CPR value ranges due to high block abundance. This would inadvertently yield a trend line that increased with a moderate slope. Initial results show a very weak correlation between CPR values and Block abundance. The trend line in figure 4 has an increase in its slope, but is very shallow. This implies that CPR values are not increasing concurrently with block abundance and may be controlled by other, more complex factors. Previous work show similar trends, but only when sub area 2 is excluded from the data sets plotted (Fig.4b).

![Graph](image1.png)

(Fig. 4a) Combined data of this study and [11] with sub area 2 excluded.

![Graph](image2.png)

(Fig. 4b) Combined data of this study and [11] with sub area 2 excluded.

**Conclusions**

Seven craters with varying age, diameter, and non-polar locations were processed and analyzed. The CPR values collected varied and were not completely consistent as expected. Though, there does appear to be a weak trend indicating that high CPR values correlate with craters that have high block densities. Sub area size could be the cause of the weakly constrained data. Sub area 2 represents the crater wall and it is believed that the slope of the crater wall influences the CPR values, by changing the incidence angle of the radar illumination. The differences in data sets and trends from previous works most likely stems from sub area size. If the sub areas are cropped too small then the sub area will not represent the respective parts of the crater accurately. More data is required in order to resolve the statistical noise and so for the future one may want to increase the sub area size in the processing phase.

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OLIVINE DETECTIONS AT THE RIM OF CRISIUM BASIN WITH MOON MINERALOGY MAPPER.

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Introduction: The lunar magma ocean theory proposes that a large outer layer of the Moon was once molten. The lunar crust and mantle would have crystallized from this early magma ocean. The mineral olivine was one of the earliest products to crystallize, and a major component of the mantle. Finding olivine on the surface of the Moon could indicate exposure of such mantle material, which would be of interest in the study of the Moon’s interior. Such materials may have been exposed by large (i.e. basin-forming) impacts that tapped the mantle, or conveyed to the surface by other means. Determining where olivine-rich rocks are present on the surface of the Moon may yield information about the various geologic processes that have been active throughout its history.

Using Kaguya’s Spectral Profiler, Yamamoto et al. [1] reported detections of olivine in regions near large impact basins. Such detections were almost exclusively limited to basin rims and central peaks. They proposed that this reflects exposed mantle excavated by the impacts that formed these basins. However, analysis of stress states from lithospheric loading by mare flows [2] suggests that the areas around large impact basins are particularly favorable for magma ascent through vertical conduits. We independently assess the presence of olivine at the lunar basin Crisium, using spectra from the Moon Mineralogy Mapper (M$^3$), on the Chandrayaan-1 mission. Using associated data products from images and spectra, as well as topographic and crustal thickness maps from LOLA and Kaguya, we address the origin of olivine at Crisium.

Methods: M$^3$ is an imaging spectrometer, and therefore provides visual context to directly link spectra to features. With the more complete coverage provided by M$^3$ (an imaging spectrometer unlike the Spectral Profiler) we were able to conduct a more comprehensive exploration of the region. M$^3$ has 85 bands across a range of 460 to 2980 nm, giving it high enough spectra resolution to distinguish mineral signatures. The M$^3$ data used here are “R”-calibrated and converted to apparent reflectance. These spectra are not thermally corrected, so wavelengths longer than ~2300 have a significant emission component. We therefore truncate the spectrum at 2300 nm for our analysis. This does not diminish the efficacy of our analysis because the relevant features are located below 2300 nm.

We take a conservative approach in declaring spectra to be indicative of olivine. The spectral signature of olivine has three absorptions features: at 0.85, 1.05 and 1.25 μm (Fig. 2). These features are expected to appear in the M$^3$ data as a wide band centered near 1μm. This signature must be distinguished from the pyroxenes; which display a 1 μm band and also a wide band near 2 μm. Rocks that are mainly olivine should lack this 2 μm band. Often our spectra show a weak 2 μm band, indicating that the rocks may be a mixture of olivine and pyroxene instead of a pure olivine.

The search for olivine with M$^3$ spectra was assisted by using an olivine index based on the algorithm developed for CRISM [3], an imaging spectrometer on the Mars Reconnaissance Orbiter. Although Clementine spectra lack the spectral resolution for an unambiguous olivine identification, from this data set we were able to create an optical maturity index [9] and TiO$_2$ [10] and FeO [11] abundance maps, providing us with additional information about the different compositional units in the region. We mapped the major geologic units in and around Mare Crisium (Fig. 3). We also benefited from Lunar Reconnaissance Orbiter (LRO) Narrow Angle Camera (NAC) imagery of particular areas of interest.
We assessed the geophysical setting of the Crisium basin with products derived from altimetry and gravity observations. We referenced LOLA topography [6] to the lunar geoid [7] and used a model of crustal thickness derived from Kaguya gravity and topography data [8] (Fig. 1).

Results: We report detections of olivine in 30 locations around the rim of Crisium (Fig. 1). They exist in several geophysical contexts including the rims of small impact craters, massifs at the rim of Crisium, and in mare or impact melt beyond the main mare-filled basin. Only one of these detections is located within the basin-filling mare: on the rim of a ~4km crater near the southeast rim of the basin.

In many instances we were able to confirm the findings of [1]. However, in several areas where they reported olivine detections, our findings were inconclusive. We report several new areas of discovery, most notably in the 7-km crater Eimmart A and within Lacus Perseverantiae. Many detections occur in an area south of Crisium, on the southeast rim of Crisium near Promontorium Agarum, and on the rim of Glaisher crater, all areas where multiple detections were reported [1].

While the crustal thickness in the Crisium region is low when considered on a global scale [e.g., 1], olivine detections at Crisium often correlate with local crustal thickness maxima. (Fig. 1) If olivine here were indicative of a substantial amount of exposed mantle material, we would not expect it to be observed where the crust is thickest.

Eimmart A: One notable detection of olivine not reported by [1] occurs in the 7-km crater Eimmart A, (24.08°N, 65.64°E) located on the rim of the 45-km crater Eimmart (23.93°N, 64.77°E). Eimmart A is a very fresh crater; it appears bright in visual bands and young in an optical maturity index. Due to Eimmart A’s position on the rim of Eimmart, it likely samples both the wall of that crater and the surrounding terrain. Eimmart in turn is located outside the mare on the northeast rim of Crisium, where it would have sampled overturned material from the impact.

Eimmart A’s walls have strongly contrasting bright and dark streaks in reflected light images. The CRISM olivine index identifies three locations for candidate olivine exposures on Eimmart A; each near the rim and at a dark streak. The spectrum from the northern rim of the crater shows a strong olivine signature (Fig. 2), and the other spectra show olivine involvement. Inspection of LRO NAC imagery of Eimmart A reveals an outcrop on the southern rim of the crater. The outcrop extends northward from the edge of the rim, not radial to the center of the crater. Nearby terrain consists of larger blocks than the northern rim. Blewett et al. [4] reported a mixture of olivine and pyroxene in Eimmart A as part of a regional survey of Crisium using Clementine spectra.

Lacus Perseverantiae: This irregularly-shaped volcanic flow south of the rim of Crisium (center: 7.78°N, 61.91°E) is penetrated by several small (<5 km) impacts, one of which reveals material with a characteristic olivine spectrum (Fig. 2).

Discussion: We argue that while some of our olivine detections may be reflect deep material uplifted by the basin-forming impact, other exposures appear to be magmatic in origin. There are two broad categories for possible transport of olivine to the near-surface: mechanical transport by basin-forming impact of deep material from either the lower crust or mantle, or magmatic transport of melts, cumulates, or xenoliths through intrusive networks. The transport mechanism is distinct from the ultimate origin of the materials themselves. Lower crustal material transported by the basin-forming process might have originally been volcanic (intrusive), whereas rapidly ascending magma might entrain peridotitic mantle xenoliths. The explanation for the placement of olivine favored by [1] was that it was of mantle origin and exposed by the impact that formed Crisium. They also acknowledge a scenario where the impact exposed only the lower crust, but where a magnesium-rich pluton had intruded.

Impact basins are particularly well-suited for stresses leading to mantle exposures or volcanic processes. [2] We have detected olivine in a variety of geologic settings at the rim of Crisium and will discuss possible origins in several different cases. We propose that while some of these materials may have been transported to the near-surface by impacts at Crisium, others were transported by volcanic processes, either as cumulates from magma storage areas or as entrained mantle xenoliths.

One promising olivine site is exposed from the near surface by an impact within Lacus Perseverantiae.

![Graph](image-url)
The lacus is clearly a mare flow based on its irregular shape, dark visual appearance, and FeO and TiO₂ abundances. The source of the flow is uncertain. The olivine spectrum appears in only one of the craters on this mare, but notably in the crater that has the largest diameter and thus should be exposing material from the greatest depth.

Eimmart A is situated on the margin of Mare Anguis. We find isolated exposures of olivine only within Eimmart A, not in the area around it. The diagnostic olivine spectral features are identifiable only in the upper crater wall and can be briefly traced in the streaks down the inside of the crater until the it is swamped by the spectrally dominating pyroxene. In several cases, landslides could have exposed olivine-rich materials beneath. The promontory that is potentially a dike may have been made of mechanically stronger material than its surroundings and thus a piece of it survived the impact that formed Eimmart A. It may even be the source for the volcanic flows of Mare Anguis.

Glaisher crater (13.14°N, 49.3°E) is located near the rim of Crisium, in rough terrain that is not clearly either mare or impact melt and far from any apparent volcanic activity. This exposure is a candidate, as [1] suggests, for either a mafic intrusion or an exposure of mantle.

Many olivine detections were found on the southeast rim of Crisium on and around Promontorium Agarum, some of which were previously reported by [1]. This area is similar to the region between 7.5-9.5°N and 58-60°E, which also has many olivine detections. Both areas have rough rim terrain and a strong gradation in elevation over a relatively small region. In contrast to several other areas of olivine detection, this southeast rim is not one of the areas where we find a crustal thickness maximum. This is a case where impact transport is plausible, though we cannot rule out volcanic transport.

EXPLORATION OF ACIDALIA MENSA AND ACIDALIA PLANITIA: PROBING MUD VOLCANOES TO SAMPLE BURIED SEDIMENTS AND SEARCH FOR ANCIENT AND EXTANT LIFE. L. M. Saper1, C. C. Allen2, D. Z. Oehler2. 1Brown University Dept. of Geological Sciences (lee_saper@brown.edu), 2Astromaterials Research and Exploration Science, NASA-JSC (carlton.c.allen@nasa.gov, dorothy.z.oehler@nasa.gov).

Here we develop a plan to explore mud volcanoes near Acidalia Mensa with an MSL-class rover and propose a traverse based on geologic observations.

Introduction:

Bright pitted cones are common in the northern plains of Mars and have been documented to occur in numerous locations including Isidis, Chryse, Utopia, and Acidalia Planitia with some of the highest concentrations occurring in Acidalia and Utopia Planitiae [1,2]. Various interpretations of these features have been proposed including cinder cones, rootless cones, phreatomagmatic tuff cones, pingos, hydrothermal springs, and geyser deposits but growing consensus in recent literature has favored mud volcanism as the most likely formation mechanism [3,4,5,6,7]. Conditions that typically lead to formation of mud volcanoes include rapid deposition of volatile-rich fine-grained sediments, a buildup of pressure at depth, and vertical migration either passively due to buoyancy or due to a triggering event such as clathrate disassociation, seismic disturbances (i.e. impact), tectonic shortening, geothermal destabilization, illitization of clays, amongst others [6,8]. Mud volcanoes are provocative targets for exploration because they bring to the surface deep sedimentary materials otherwise inaccessible at typical surface exposures.

Acidalia Planitia was likely the primary depositional center for the distal facies of sediment debouched during outflow channel formation and the diverse materials deposited there are representative of an expansive catchment area in the northern highlands [7,9,10]. The lack of exposure of buried materials in the northern plains makes exploration of mud volcanoes invaluable in reconstructing the sedimentary history. Mud volcanoes are equally as important for astrobiology because of the unique ability to point sample fine-grained sediments derived from a large area and integrated over a vertical column of strata. Also, by sampling fluids from deep in the Martian crust, mud volcanoes may be among the best places to search for recent or extant life.

Geologic Setting

The geologic history of Acidalia Planitia can be summarized in five stages (i) basin formed by a mega-impact into early Noachian crust, (ii) volcanic resurfacing by Hesperian ridged plains, (iii) deposition of sediments likely derived from outflow channels, (iv) formation of Vastitas Borealis (ABv), possibly by ocean sedimentation or as a sublimation lag from a frozen water body, (v) Amazonian resurfacing by obliquity-related periglacial processes [9,11]. The current surface of the planitia contains bright pitted mounds and cones, high concentrations of pedestal and rampart craters, and is dissected by kilometer scale polygon-forming troughs, all evidence of a volatile-rich subsurface history [12]. Current gully formation is pervasive throughout the region, indicative of active ice-related processes in the subsurface.

Acidalia Mensa is an isolated massif that is one of the highest topographic landforms in the northern plains and exposes the oldest bedrock north of the dichotomy boundary [9]. The mensa is modified by troughs and fractures and is adjacent to a knobby colles indicating a volatile-rich history.

Geologic Observations in Exploration Area

At least five distinct units have been distinguished using the suite of orbital datasets (CTX, HiRISE, THEMIS, MOC, CRISM) based on morphology, thermophysical characteristics, spectral properties, decameter to meter scale texture, and the extent of tectonic modification.

The oldest unit in the exploration area is interpreted to be Nepenthes Mensa (Nn), as mapped by [9]. It is Noachian in age and is comprised of knobs and mesas that erode to aprons of high-albedo material. The unit is dissected by linear troughs that extend tens of kilometers and are up to several kilometers in width. Nepenthes Mensa forms the southern scarp of Acidalia Mensa, with up to 200 meters of relief in places. A few isolated knobs and mesas south of the scarp are interpreted to be part of the same unit, perhaps indicating that the scarp has receded. Alternatively these isolated mesas, possibly layered, may represent tuya or subglacial volcanic structures [13].

The next oldest unit is interpreted to be volcanics that appear to have flowed over Nn, flowing into and filling troughs and craters and will not be accessed by the rover.

Dome and cone shaped high-albedo mounds occur south of the mesa often expressing one to several muted central depressions with aprons of material at the base. Interpreted to be mud volcanoes, these features match morphologies reported elsewhere in the northern plains and have a visible spectral response of ferric oxide coatings consistent with previous observations of similar mounds [7]. Fully formed mounds with
central edifices are typically ~1 kilometer in diameter, though many smaller mounds exist. A few examples have well-developed flow features, extending up to 2-km from the mound and are primary targets for exploration.

The mud volcanoes are located within a low-albedo unit having concentric flow features with lobate margins and is interpreted to be younger than the mounds although evidence for the stratigraphic relationship is unclear. It flows downslope from the northwest, thinning to the southeast, and eventually grades into typical plains materials (i.e. ABv). This unit also contains circular to quasi-circular features with steep, inward dipping raised rims of unknown origin. The annular features appear to be pre-existing craters that were filled but often have unusual morphologies, tend to be associated with topographic hills, and in a few places appear to have mud-like flows protruding from their rims. The interpretation of these features is unclear and may not be confidently determined until landed investigation.

The youngest feature is a ray of ejecta and secondary craters emanating from the Bonestell Crater impact ~160 kilometers to the southeast forming high concentrations of irregularly shaped craters, streaks striking SW-NE, and superposes all other units. Outstanding Questions to be Addressed in Exploration

1. What is the nature of the supposed mud volcanoes? If sedimentary diapirs, what materials do they source? Do they contain organics? What can they tell us about the sedimentary history of the northern plains? 2. How did Acidalia Mensa form? What was the role of volatiles? Is there any evidence of a large standing body of water? Did it leave any traces and how did it interact with the mensa? 3. What is the composition of the younger flow unit (mud or volcanic)? How does it relate to the mounds? What is the composition and nature of the enigmatic circular features? How does their interpretation influence crater-retention ages in the region? 4. How have tectonics shaped the Nepenthes Mensa unit and the formation of troughs on the mensa? 5. What is the nature of the interaction of ejecta rays and secondary craters with preexisting terrains?

Exploration Plan:

A traverse has been developed that would enable assessment of the main science questions within two Martian years allowing time for winter ‘hibernation’ (Fig. 1). The total traverse is ~29 kilometers, assuming landing in the center of the proposed ellipse, which given MSL average speed projections and the terrain is a plausible distance to be covered in the given timeframe, allocating ample time for scientific observations along the way. Upon arrival to each scientific target detailed measurements of bulk chemistry, mineralogy, grain size, and sedimentary structures if applicable will provide context and allow for correlation of materials between different locations along the traverse. A possible payload of instruments would include a similar suite to MSL including color and hyperspectral imaging cameras, a microscopic imaging camera (preferably with the ability to detect microfossils), mass and laser spectrometers, gas chromatograph, XRD/XRF, rock abrasion tools, and sample caching capabilities depending on sample return architecture.

Upon landing on the unit with flow-like properties on Acidalia Mensa the rover will assess its surroundings and perform chemical analysis to determine the composition and origin of the landing materials. After these initial measurements the rover will begin several days of traverse towards the southern scarp of the mensa, utilizing imaging cameras for context and hazard avoidance. After ~75 sols of traverse at 6 kilometers the rover will encounter the first science target – a string of secondaries from Bonestell superimposed on the landing materials. After a brief investigation of impact processes and secondary crating the rover will begin to descend off the mensa following a channel of flow material that provides a passageway down into the plains. During the journey off the mensa the rover can sample the scarp of the Nepenthes Mensa unit, looking for any sedimentary structures and evidence of water and providing a detailed analysis of the composition of the ancient unit.

At the base of the mensa the rover will sidestep more Bonestell secondaries and arrive at its first mud volcano after a traverse of around 15 kilometers and 230 sols of surface operations. The first mound has been affected by Bonestell (Fig. 1A), potentially providing a window for the rover to explore the interior of the mound. Here the rover will perform detailed analyses of mound materials and make the first observations on the nature of its formation. If it is a mud volcano, the rover should administer the full-suite of ana-
yses on the fine-grained matrix as well as a collection of 5-10 different clasts. Ideally the rover would sample a diverse range of materials and begin to understand what types of rocks are present in the source region. These analyses would include a search for organics in samples with the best preservation potential and analysis of aqueous indicators (hydrated minerals, structures). Next the rover would travel towards the largest flow feature at 22-km (Fig. 1B), taking time along the way to investigate the enigmatic circular rimmed features. If these contain evidence of volatiles, they may require a longer investigation. Similar analyses will be made at subsequent mud volcanoes, with attention to sedimentary structures and apparent flow events. The rover will continue to traverse over flow materials towards the mound that best represents similar features elsewhere on Mars (Fig. 1C) at 26-km. Here the rover will attempt to drive to the central edifice of the mound to explore a potential vent. The final exploration target is a lobate structure emanating from a raised-rim circular feature (Fig. 1D). The rover will explore the interaction between the two landforms, and potentially make a final determination of what both are.

Possible long-term targets include more mud volcanoes, isolated mesas and knobs in the plains (potentially a ‘layered’ mesa to the west), more of the scarp of Acidalia Mensa, and a nearby gullied mesa.


Figure 1. Background – MOLA DEM showing topography and geographic context of study area. Stretch is -5700m to -4000m. Black rectangle shows location of lower-right inset. Inset – CTX mosaic (P18_007889_2288_XI_48N023W, P17_007520_2267_XN_46N028W) showing proposed landing ellipse (red), nominal traverse (black), and context of exploration area. White squares indicate locations of magnified insets: A – Bonestell secondaries, mensa scarp (white arrows), and first mud volcano (black arrow), B – mound with extensive flow feature, C – well-formed mound with central edifice and vents, D – flow-like feature protruding from enigmatic annular raised-rim feature (B, C, D – HiRISE image ID ESP_018649_2250).
PETROLOGY OF AN IMPACT MELT CLAST FROM LUNAR REGOLITH BRECCIA 60016.
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Introduction: As the most common lunar surface process, it is important for us to understand as much as possible about impact-related geology on the Moon. By studying lunar meteorites and Apollo samples, we can begin to categorize the available impact melts found in these rocks and learn more about the geologic history of the early solar system. One of the interesting results of studies so far is that many workers have noticed the ages of lunar impact events appearing to form a cluster around 3.9 Ga. It has been suggested that during this time there was a period of heavy bombardment by asteroids that resurfaced much of the Moon and created many of the impact basins that currently dominate the lunar surface [e.g. 1,2].

Although it is difficult to connect individual impact melts with specific impact events, grouping these samples by examining textures, chemical compositions, and radiometric ages can delineate suites that belong to distinct impacts, which could eventually be connected to specific basins and craters as more data is collected. Many large volume impact melts have been studied and placed into compositional and age-based groups, but smaller volume samples have remained largely unstudied. By studying small volume melt clast 60016, 321, we can begin to determine if these small clasts represent the same suites as the larger melts, or if different impacts are being represented.

Sample and Analysis Methods: Sample 60016, as shown in Figure 1, is a light grey regolith breccia that contains a large variety of both light and dark colored clasts, as well as some brown staining associated with metal [3]. The clast that produced split 321 is a dark, small volume impact melt approximately one centimeter in size. For this study all analyses were performed with a thin section (.321) of the clast. This thin section was analyzed with a petrographic microscope as well as backscatter imagery created by a JEOL-7600 field emission scanning electron microprobe. Chemical compositions were determined with a Cameca SX100 electron microprobe using a 15 keV, 20 nA beam with a 1µm beam diameter for minerals and a 20µm series of 453 dispersed beam analyses for bulk chemical composition.

Results:

Petrography. Sample 60016,321 is a crystallized impact melt, and has a very fine grained, porphyritic texture. There are several groups of sub-poikilitic grains, but this is not representative of the overall clast texture. The phenocrysts in the rock are all olivine and plagioclase grains. Birefringence is of little help in distinguishing between the two phases; olivine birefringence ranges from 0.007-0.040 which overlaps the 0.007 birefringence of plagioclase. Smaller phenocrysts (less than 60 or 70 µm) are euhedral or subhedral and appear to have recrystallized from the melt, but larger grains are fractured, anhedral, contain groundmass inclusions, and are most likely relict grains. Figure 2 shows a backscatter-electron image of a relict plagioclase grain surrounded by the porphyritic texture generally exhibited in split .321. The ground...
mass is composed mainly of poorly formed microcrystalline pyroxene crystals, but some elongate plagioclase laths are present as well. These laths create a seriate texture that ranges from microcrystalline to ~8µm, at which point they become less elongate and more similar in shape to the blockier plagioclase phenocrysts. There is one small relict lithic clast embedded in this sample. It is composed of anhedral anorthite grains with a small amount of pyroxene groundmass. The composition of the plagioclase in the clast is similar to that in the rest of the rock. The rock contains about 2.2% metal; half of the grains are smaller than 4µm, and a quarter of the grains are smaller than 1µm. Sample 60016, 321 also contains 0.47% sulfides, which are commonly between 4-16µm. The larger metal grains are generally circular in cross-section and range between 4-160µm in diameter.

Mineral Chemistry. Out of the 32 olivine grains measured only 5 are zoned. Three are recrystallized grains between 20 and 30 µm, and the other two larger grains are most likely relict grains. As shown by the histogram in Figure 3, the average olivine core composition for all grain sizes is Fo86, and the rims of zoned grains range between Fo82-Fo84. Plagioclase grains are all unzoned and have a composition of An90-An97, shown in Figure 4. The individual phases in the groundmass of sample 60016, 321 are comparable in size to the electron beam, so contamination of adjacent phases is possible. None of the analyzed areas in the groundmass are stoichiometrically balanced, but are compositionally similar to clinopyroxene that is slightly contaminated by surrounding plagioclase. Alternatively, the non-stoichiometric analyses could indicate the groundmass contains a glassy component [e.g. 12]. No orthopyroxene was found.

**Figure 3.** Histogram of the Mg# (atomic percent of Mg/Mg+Fe) of the core of each olivine crystal measured in sample 60016, 321, including possible relict crystals.

**Figure 4.** Composition of plagioclase grains measured in sample 60016, 321 on a ternary diagram which plots the atomic percent of K(not shown), Na, and Ca.

**Figure 5.** Distribution diagrams using the bulk split analysis of 321. Figure 5a shows K2O weight percent plotted against Al2O3 weight percent. Figure 5b shows Mg# (atomic proportion of Mg/Mg + Fe) plotted against Al2O3 weight percent. Data for Apollo 16 bulk rock analysis have been added to the plot for reference [5,6,7,8,9,10,11,12]. Compositional fields are those described by Cohen et al., 2005[1].
Discussion: The grouping system most commonly used to distinguish between Apollo 16 impact melts was proposed by Korotev [12]. Both sub-groups in group 1 rocks from Korotev’s study and split .321 have high metal contents, and are mafic. However, one of the other distinctive characteristics of group 1 is the poikilitic texture exhibited by all of the previously studied samples[12], which this impact melt clearly lacks. The Mg# of split .321 is too high to be placed in group 1F, but not as high as the samples classified as group 1M. The K$_2$O concentrations in this split are higher than other Apollo 16 impact melts. A separate study by Norman et al. [2] organizes Apollo 16 mafic melts by age as well as petrographic and chemical characteristics. Because these suites are proposed largely based on age it is difficult to place split .321 without radiometric dating results. However, since all of the group 1 samples are considered to be part of the same suite, it is possible that this impact melt could be included in this suite as well. The KREEP signatures found in this sample indicate that the melt is from an impact event which occurred on the Procellarum Formation, such as Imbrium[13].

Conclusion: The impact melt sample that was extracted from 60016 seems to have a major element composition similar to that in the samples of Korotev’s group 1. Although the K$_2$O concentrations are higher, the intermediate Mg# of split .321 could indicate that it is a mixture of the two group 1 categories. It is interesting that the poikilitic texture associated with these sub-groups is absent, but it is possible that the cm-size clast in our study is not representative of the parent rock texture [2, 12]. Thus, the melt clast is tentatively classified as a group 1 melt, and is likely from the Imbrium impact event or another event that impacted the Procellarum Formation. Further work on this rock should include radiometric dating to see if it can be more easily placed into one of the impact groups described by Norman et al. in 2006. Trace element data will also be useful, as it will help to compare this melt with those used in Korotev’s work, which primarily involves samarium, scandium, and other trace elements.

References:
Simulation of MHD Turbulence in a Spherical Domain. Yifan Wang\textsuperscript{1} and John V. Shebalin\textsuperscript{2}
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Introduction: Our knowledge of Earth’s magnetic field has advanced greatly since 1900, but finding its root cause remains a challenge. Larmor (1919) first suggested the possibility of dynamo action by magnetic field generation due to the movement of electrically conducting fluid. Dynamo theory has been developed by Elsässer and Bullard (1939), as well as others, while modern computers have permitted realistic simulations of the geodynamo through numerical solution of the magnetohydrodynamic (MHD) equations [1].

MHD turbulence is expected to be an important part of the geodynamo. Basic studies of homogeneous MHD turbulence use Fourier methods to model flows in periodic boxes. Numerical simulations of ideal homogeneous MHD turbulence can be compared to the statistical prediction that spectral coefficients are zero-mean random variables. These absolute equilibrium ensemble theories have probability densities that are based on ideal invariants such as energy and cross-helicity, along with either mean squared vector potential in 2D, or magnetic helicity in 3D MHD turbulence.

However, broken ergodicity is readily observed in computer simulations at low wavenumbers [2,3], i.e., Fourier coefficients do not behave like zero-mean random variables over time (Fig. 1). Broken ergodicity has recently been explained theoretically in 3D and 2D homogeneous MHD turbulence [4,5]. These results suggest that broken ergodicity may play a similar role in the geodynamo and other planetary dynamos.

To relate these advances in homogeneous MHD turbulence research to the geodynamo, our present work is focused on MHD turbulence in a spherical domain. As an initial effort, we consider the simpler 2D case, which greatly reduces the complexity of the MHD equations. However, in contrast to the homogeneous case, a 2D spherical formulation includes Corioli forces, which are very important in geodynamo modelling, and is thus a useful intermediate step.

Approach: For a sphere with a radius of unity, the 2D spherical imcompressible MHD equations can take a form similar to the 2D planar case [5]:

\[
\frac{\partial \omega}{\partial t} = J(\psi, \omega) + J(j, a) + 2\Omega_o \frac{\partial \psi}{\partial \phi} + \nu \nabla^2 \omega \\
\frac{\partial a}{\partial t} = J(\psi, a) + \eta \nabla^2 a
\]

(1)

(2)

Here, \( \omega \) is the fluid vorticity, \( a \) is the magnetic potential, \( \psi \) is the fluid stream function, \( j \) is the current density, \( \Omega_o \) is the angular frequency of rotation, and the Jacobian \( J \) contains the nonlinear interactions:

\[
J(f,g) = \frac{1}{\sin \theta} \left( \frac{\partial f}{\partial \phi} \frac{\partial g}{\partial \theta} - \frac{\partial f}{\partial \theta} \frac{\partial g}{\partial \phi} \right)
\]

(3)

The variables \( \omega \) and \( a \) on the surface of a sphere are represented using spherical harmonics \( Y^m_n \) [6]:

\[
\omega(\theta, \phi) = \sum_{n=1}^{N} \sum_{m=-n}^{n} \omega_{nm} Y^m_n(\theta, \phi) \]

(4)

\[
a(\theta, \phi) = \sum_{n=1}^{N} \sum_{m=-n}^{n} a_{nm} Y^m_n(\theta, \phi) \]

(5)

For ideal 2D MHD turbulence (\( \nu = \eta = 0 \)), the model dynamical system conserves total energy \( E \) and mean squared magnetic potential \( A \); the cross-helicity \( H_c \) is conserved only when there is no rotation:

\[
E = \frac{1}{2} \sum_{n,m} \left| \omega_{nm} \right|^2 + n(n+1) \left| a_{nm} \right|^2
\]

\[
A = \frac{1}{2} \sum_{n,m} \left| a_{nm} \right|^2
\]

\[
H_c = \frac{1}{2} \sum_{n,m} (\omega_{nm} a_{nm}^*)
\]

(6)
**Method:** The coefficients \( \omega_{nm} \) and \( a_{nm} \) were initialized with their magnitude defined by a Gaussian distribution peaking at a given \( n \), and their phase defined by a random number generator. Numerical solutions of equation (1) and (2) were obtained by integrating forward in time using a third-order Adams-Bashforth predictor and an Adams-Moulton corrector.

The nonlinear Jacobian terms were evaluated by a spectral transform method. The spatial derivatives \( \partial / \partial \theta \) and \( \partial / \partial \phi \) were transformed back into physical \((\theta, \phi)\) space, thus their arithmetic calculations could be evaluated at each discrete grid points to obtain the Jacobian. Subsequently, the spherical coefficients of the Jacobians were obtained by transform back to \( n,m \) space. Not having to perform a convolution sum allows significant reduction in required computer time. **Aliasing** during the evaluation of these non-linear term were eliminated using the ‘2/3’s rule’ where the coefficients corresponding to \( n>2/3*N \) were set to zero at every iteration [7].

SPHEREPACK package developed by Adams and Swartztrauber contains FORTRAN subroutines that handles conversion between spherical coefficients \((a_{nm}, \omega_{nm})\) and physical space \((\theta, \phi)\) [8]. It was utilized to perform the spherical harmonic transforms.

**Results Discussion:** The simulation was performed with \( N = 32 \) (32 latitudinal, 64 longitudinal grid points) and a time step of \( 10^{-5} \) over \( 3 \times 10^6 \) iterations. The simulation took 50 CPU hours to run.

**Invariants:** From the plot of energy, cross-helicity and mean potential (Fig. 2), we observed that the total energy and vector potential remain constant, but the cross helicity only remained constant at \( \Omega_o = 0 \) and Therefore the conditions of conservation described in (6) were satisfied, and the distribution of the coefficients in phase space can be described with a Gaussian canonical probability distribution and the coefficients are expected to be zero-mean random variables [5].

\[
D = \frac{1}{Z} \exp(-\alpha E - \beta H_c - \gamma A) 
\]

Near the end of the simulation, numerical error induced by transforming between coefficients and real space starts to affects the results, as the total energy starts to slowly deviate from the constant value of \( 4\pi \). The magnitude of this error is proportional to the grid size and the number of iterations.

**Behaviour of coefficients:** Development of broken ergodicity depends strongly on the grid size used. In the 2D homogeneous case [5], there is no evident broken ergodicity at small grid sizes \( (N = 8) \) and a clear coherent structure only develops during long runs of large grid sizes \( N \geq 32 \).

---

Figure 2. The change of kinetic energy, magnetic energy, total energy, cross helicity and mean vector potential with time. Note that the total energy is largely conserved but starts to deviate from the constant value at \( t > 20 \).
Due to the time constraint of this program, multiple long runs over large grids were not feasible. However, broken ergodicity was still observed. The contour plots (Fig 3) shows that the time average of the coefficients for large $n$ remains very close to zero, but peaks were clearly distinguishable at the lowest values of $n$.

These observations are similar to those seen in homogeneous MHD turbulence simulations [2-5], suggesting that significant large-scale coherent structures should appear at low wavenumbers when long-time simulations are performed on larger and larger grids. In addition, larger grids will allow a better evaluation of dissipative effects.

**Conclusion:** We developed a 2D spherical code and observed broken ergodicity (large-scale coherent structure) through numerical simulation of ideal 2D MHD turbulence. Analogous behavior has been observed in 3D MHD homogeneous turbulence simulations [4] and we may expect a similar outcome from 3D spherical numerical simulations, though this case is decidedly more difficult to study theoretically and numerically. In summary, ‘broken ergodicity’ may help explain the appearance of large-scale, quasi-stationary magnetic fields in planetary cores containing turbulent magnetofluids.

**Further Work:** The speed and accuracy of the code need to be improved for running at a higher grid resolution and iteration number. Currently, the associated Legendre function transform part of the spherical harmonic transform is evaluated using numerical quadrature, which is significantly slower than a Fast Fourier Transform. Thus, it is challenging to perform a spherical domain simulation on a grid size equivalent to one used for a planar simulation. In the 2D case, a detailed, systematic study using different rotation rates $\Omega_o$ and initial conditions is planned. In addition, an extension to 3D will allow a treatment more relevant to the geodynamo problem.

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Fig. 3: The time average and variance of the (a) real and (b) imaginary components of $a_{mn}$, Time averages largely obey ensemble predictions but spike at low $n$.

**References:**