Papers Presented at the

August 8, 2013 — Houston, Texas
Papers Presented at the

Twenty-Ninth Annual
Summer Intern Conference

August 8, 2013
Houston, Texas

2013 Summer Intern Program for Undergraduates
Lunar and Planetary Institute

Sponsored by
Lunar and Planetary Institute
NASA Johnson Space Center
HIGHLIGHTS

Special Activities

June 3, 2013    Lunar Curatorial and Stardust Lab Tour          NASA JSC
June 28, 2013   Ellington Field Tour                           Ellington Field
June 12, 2013   Meteorite Lab Tour                            NASA JSC
July 30, 2013   NASA Building 9N, CCT II, SM, Soyuz, FGB and JEM NASA JSC
                (Crew Compartment Trainer 2, Service Module, Functional Cargo Block, Japanese Experiment Module)
July 30, 2013   Image Science and Analysis Lab                 NASA JSC

LPI Summer Intern Program 2013 – Brown Bag Seminars

*Wednesdays, 12:00 noon – 1:00 p.m., Lunar and Planetary Institute*

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AGENDA

8:00 a.m.  BREAKFAST

8:25 a.m.  Introductory Remarks by Drs. Paul Spudis and Dave Draper

8:30 a.m.  CAITLIN ALTOMARE, Lafayette College
(Advisor:  David Kring)
Eolian Deposits of Pyroclastic Volcanic Debris in Meteor Crater [#1011]

8:50 a.m.  MICHAEL BOUCHARD, Missouri University of Science and Technology
(Advisor:  Lee Graham and David Meledrez)
Crewed Martian Traverses; Building on Lessons Learned from Apollo, Robotic Missions, and Planetary Analogs [#1007]

9:10 a.m.  ANDREA BRUCK, Illinois State University
(Advisor:  Brad Sutter)
An Investigation of Calcium Chlorine Oxoanions Role in the Evolved Oxygen Release Found by Mars Curiosity Rover Sample Analysis at Mars (SAM) Instrument at Rocknest Eolian Bedform [#1003]

9:30 a.m.  LAURA CORLEY, University of Hawaii
(Advisors:  Patrick McGovern and Georgiana Kramer)
Olivine Exposures on the Moon: Origins and Mechanisms of Transport to the Lunar Surface [#1012]

9:50 a.m.  ANGELA DAPREMONT, College of Charleston
(Advisor:  Carlton Allen)
The Gale Crater Mound in a Regional Geologic Setting: Comparison Study of Mt. Sharp and 1000 km Radius [#1002]

10:10 a.m.  BREAK

10:30 a.m.  WANDA FENG, Smith College
(Advisor:  Cindy Evans, John Gruener, and Dean Eppler)
Comparing Geologic Data Sets Collected by Planetary Analog Traverses and by Standard Geologic Field Mapping: Preliminary Results [#1013]

10:50 a.m.  ANTHONY FRUSHOUR, Appalachian State University
(Advisor:  Sarah Noble, Lindsay Keller, and Roy Christoffersen)
Alteration of Lunar Rock Surfaces Through Interaction with the Space Environment [#1008]

11:10 a.m.  CHRISTOPHER HOFF, University of Massachusetts
(Advisor:  John Jones)
HED Parent Body Differentiation [#1006]

11:30 a.m.  EVE LALOR, Temple University
(Advisor:  Virgil “Buck” Sharpton)
Analyzing Rim Crest Variations in Lunar Impact Craters [#1009]

11:50 a.m.  DAYL MARTIN, University of Manchester
(Advisor:  Paul Spudis)
Geology of the Orientale Basin and its Impact Melt Sheet [#1004]

12:10 p.m.  LUNCH
1:00 p.m.  **MOUNA PETITJEAN, Université Paris-Sud**  
(Advisor: Stephen Clifford)  
*The Correlation of Martian Crater Ejecta Morphology with Geologic Units and Radar Surface Permittivity in the Northern Plains* [#1014]

1:20 p.m.  **ALASTAIR TAIT, Monash University**  
(Advisor: Justin Simon)  
*Preliminary Strain Measurements of the Chondrules and Refractory Inclusions in the Allende CV3 Meteorite* [#1005]

1:40 p.m.  **ATSUSHI TAKENOUCHI, University of Tokyo**  
(Advisor: Michael Zolensky)  
*What are Space Exposure Histories Telling us About CM Carbonaceous Chondrites?* [#1001]

2:00 p.m.  **KELSEY WILLIAMS, Brown University**  
(Advisors: Yann Sonzogni and Allan Treiman)  
*Amphibole in Martian Meteorite Tissint: Composition and Implication for Water Content of Parental Magma* [#1010]
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EOLIAN DEPOSITS OF PYROCLASTIC VOLCANIC DEBRIS IN METEOR CRATER

C. M. Altomare¹ and D. A. Kring², ¹Lafayette College, Easton, PA 18042, USA altomarc@lafayette.edu, ²Lunar & Planetary Institute, 3600 Bay Area Blvd., Houston, TX 77058, USA, kring@lpi.usra.edu.

Introduction: Barringer Meteorite Crater, more commonly known as Meteor Crater, was excavated by an impact event approximately 50 ka [1-2]. It is Earth’s best preserved impact crater, and is therefore fundamental to our understanding of impact cratering processes. Consequently, it is a classic analogue for impact craters on other planetary surfaces, such as Mars. The impact site is 29 km southeast of the San Francisco Volcanic Field (SFVF), Arizona and 53 km east of Flagstaff, Arizona. The crater is 1.2 km in diameter, 180 m deep, and is embedded in the Colorado Plateau, a region composed of relatively planate sedimentary lithologies [3].

Although Meteor Crater is well-preserved, erosion has occurred and sediments eroded from the crater’s wall are deposited on the floor. Six years ago, an eolian deposit with a minimum thickness of 86 cm was discovered (Figures 1 & 2) along the southwest crater wall, beneath a layer of colluvium. The deposit contains sediments, derived from the local target material, and subsequently eroded from the crater walls (i.e. carbonates and quartz sandstones), as well as unexpected pyroclastic volcanic ash and cinder debris.

The purpose of this research is two-fold: (1) conduct petrologic analyses of the pyroclastic debris within the eolian deposit, and (2) identify the source of the volcanic particles by comparison with petrologic characteristics of regional volcanic vents. By doing so, it may be possible to identify the mode of transport of the particles to the crater, and thus serve as an analogue to similar processes acting on impact craters on Mars.

Sample and Analysis Methods: Four samples of the eolian deposit (MC51211-1a, 2a, 3a, and 4a; see Figure 2 for location relationships) were first examined under an optical petrographic microscope to characterize the internal textures and particle sizes. Two of these samples (1a & 4a) were selected for further compositional analyses of the mineral phases (olivine, pyroxene, and plagioclase).

Results: Petrography. Sample MC51211-1a is bimodal and captures two horizons in the sediment sequence with a porosity of 11.5%. The lower, and coarser, horizon contains sand-sized sedimentary...
particles of hematite-cemented quartz siltstone (16.1%), quartz-sand-bearing micritic carbonate (10.3%) and micritic carbonate (5.1%). The overlying, finer-grained, layer is a mixture of sedimentary (25.3%) and volcanic (28.1%) particles of ash (<2 mm), “other” particles (0.5%), and undetermined material with mineral/melt inclusions (3.0%). Undetermined material may be epoxy resin contamination during sample preparation. Sediment particles in the lower layer (1.0 to 10.0 mm long dimension) display low sphericity and are generally sub-rounded, though shape does vary. In contrast, the overlying layer is composed of fine silt-sized particles (0.1-0.5 mm long dimension) showing diversity in both sphericity (low to high) and shape. Poor sorting is indicated by the variety of particle size, sphericity and shape within both horizons.

Volcanic particles within the overlying layer (28.1%) are composed of phenocrysts of olivine, plagioclase, and occasional clinopyroxene (25.8%); some particles are vesicular. Microcrystalline volcanic particles compose the remaining 2.3%. The volcanic particles are coarse to medium silt size fractions and display a diverse range of sphericity and shape, which is typically more angular than the sedimentary particles, indicating poor sorting.

Sample MC51211-4a, which was collected ~87 cm above sample 1a (Figure 2), is unimodal and represents a single horizon in the sediment sequence with 20.2% porosity. It contains ash-sized sedimentary (39.3%) and volcanic (35.6%) particles, as well as undetermined material with mineral/melt inclusions (4.9%). Sedimentary particles are composed of silt-sized grains of quartz (35.2%), sutured quartz (0.1%), and micritic carbonate (1.4%), as well as sand-sized particles of hematite-cemented quartz siltstone (0.05%) and quartz-sand-bearing micritic carbonate (2.5%). The sedimentary particles (0.1 - 0.5 mm long dimension) have diverse sphericity (low to high) and shape.

The vesicular, silt-sized volcanic particles of sample MC51211-4a (35.6%) contain olivine and plagioclase phenocrysts with occasional clinopyroxene phenocrysts (33.3%) and microcrystalline volcanic fragments (2.3%). Particle sphericity and shape diversity indicate poor sorting. Overall shapes of the volcanic lithologies are more angular than the sedimentary lithologies.

Mineral Chemistry. Forty-two particles of ash and cinder were analyzed using the EMP from samples 1a and 4a. These samples show a range in mineral composition with microcrystalline plagioclase of An_{20-25.5}, Ab_{42.8-27.2}, Or_{2.0-1.0}. Sometimes associated with groundmass pyroxene (W_{4.4}, En_{44.7-36.8}, Fs_{9.9-11.5}) and/or olivine (Fo_{90.5-74.9}). Olivine phenocrysts (Fo_{96.3-72.4}) and/or plagioclase phenocrysts (An_{74.2-57.5}, Ab_{41.2-25.1}, Or_{1.2-1.0}) may also be present (both 100-200 µm long dimension). Olivine phenocrysts are normally zoned with rims of Fo_{84.3-73.4} (Figure 3a).

Discussion: Potential Source Constraints: Particles containing Mg-rich olivine, Ca-rich plagioclase and Ca-rich clinopyroxene can be classified either as vesicular basaltic or vesicular andesitic basalt. This suggests the potential source to be a strombolian-like eruption.

![Figure 3: Olivine Forsterite histograms for (a) MC51211 (this study), (b) Sunset Crater [5,6] and (c) Merriam Crater samples [6,7].](image)
potential volcanic sources for the eolian deposit; two erupted >12 ka (West Sunset Mountain and Merriam Crater), two have unreliable ages which may be older or younger than 12 ka (SP Crater and Saddle Mountain) and one erupted <12 ka (Sunset Crater)

**Potential Sources and Suitability:** West Sunset Mountain (4.8 Ma) is approximately 13 km south of Meteor Crater and is basaltic in composition [8]. Despite predating the Meteor Crater impact event (50 ka), its close proximity could potentially allow ash and cinder transportation through ground transport flow via current prevailing winds (SW-NE, [9-10]).

Merriam Crater cinder cone is in the southwestern portion of the SFVF, ~42 km N-NW from Meteor Crater. Vent lava flows have compositions ranging from alkali olivine and alkali-rich high-alumina basalt to basaltic andesite [6]. There is debate on the precise age of the crater with one study suggesting 150±30 ka [11], while another suggests 293±10 ka [12]. Thus, the cinder cone likely predates Meteor Crater and the subsequent eolian deposit; ground level flow would be required for particles to travel to and be deposited into Meteor Crater. For this to occur, however, prevailing winds at the time of deposition would need to transport particles N-S.

The basaltic SP Crater cinder cone is in the northern region of the SFVF. Some studies estimate the cinder cone to be older than Meteor Crater [13,14]. In particular, one study reported a SP flow age of 71 ± 4 ka [15], but another suggested an age of only 5.5-6 ka [16]. If SP Crater predates Meteor Crater, it is unlikely to be the source of the pyroclastic debris due to the required ground transport distance (~82 km), and the necessity of N-S prevailing winds. If, however, SP Crater is younger than Meteor Crater, then the explosive eruption could easily transport particles to Meteor Crater in an ash cloud.

Similarly, Saddle Mountain (17 ka) is a possible source due to its basaltic andesite composition [17], close proximity to Meteor Crater (~78 km) and an age that might post-date the eolian deposition.

Sunset Crater (0.9 ka) is a basaltic cinder cone directly west of San Francisco Mountain in the SFVF. Based on its young age and proximity (~57 km) to Meteor Crater alone, Sunset Crater is the most likely source of the MCS1211 samples. Pyroclastic material originating from the vent is known to cover at least ~315 km² and nearly reaches the edges of Flagstaff, Arizona, where layers of ash and cinder are up to ten cm thick [18]. Thus, material could have easily been deposited within Meteor Crater from the eruption itself and subsequent ash cloud. Ground level transport to the crater is unfavorable given the SW-NE prevailing winds. However, prevailing winds become more northerly in the winter months [10,11], and in such a case, ash and cinder could also be transported by ground level eolian flow to Meteor Crater. Strong wind storms in the region are able to transport sediments across distances of 60 km, which is greater than the distance between Sunset Crater and Meteor Crater. Additionally, Shoemaker attributed two thin ash layers in the floor of Meteor Crater to Sunset Crater [4], implying that deposits from the eruption could indeed travel to the crater.

**Mineral Compositions.** No material was available for direct analytical comparison from the potential sources, thus any comparisons were reliant on data from the literature. Published mineral compositions were only available for Merriam and Sunset Craters (Figure 3b-c), which have similar compositions. Analyzed Sunset Crater olivine phenocryst cores may be remnant xenocrysts from the interior of the vent, causing misinterpretation of the composition. The mineral compositions from the eolian deposit within Meteor Crater are indistinguishable from the two potential sources, thus composition cannot be used to confirm or rule out either cinder cone. The chemical similarities to the eolian deposit suggest that either cone is a likely candidate.

**Conclusion:** The source of volcanic debris in Meteor Crater is unclear, although Sunset Crater and Merriam Crater are the most likely sources. Both cinder cones have similar mineral compositions to the pyroclastic particles and are basaltic in nature. Furthermore, both are <60 km from Meteor Crater. The age difference between the two sources, however, likely requires two distinct transport mechanisms.

CREWED MARTIAN TRAVERSES; BUILDING ON LESSONS LEARNED FROM APOLLO, ROBOTIC MISSIONS, AND PLANETARY ANALOGS. M. C. Bouchard, Missouri University of Science and Technology (319 Shady Meadows Drive, Ballwin MO 63011, mcbmv4@mail.mst.edu)

Introduction: The process of planning a planetary traverse is a delicate balance between engineering, science, and operations protocol. An effective planetary traverse will maximize the science return and proof of engineering concepts while operating within the mission constraints. It requires detailed pre-work and multiple iterations to complete, and studying previous extra-terrestrial traverses can lend insight into how to best continue the process. Traverses involving humans add many layers of complexity. Human missions to Mars will include many of these complex and detailed traverses. Studying the only other extra-terrestrial traverses, those of NASA’s Apollo program (1966-1972), can lend insight into these future traverses. However many of the logistics of field studies on Mars are more complicated, most notably the time delay in communication between the field crew and NASA centers on Earth. The robotic missions of the Mars Explorer Rovers (MER) and now the Mars Science Laboratory have experience working with this delay and have lessons that can be applied. Much can also be gained from studies and demonstrations here on Earth. Many planetary traverse and field study analogs exist, one of which, the Desert Research and Technology Studies (D-RATS), seeks to test procedures, technologies, and traverse concepts in order to inform and progress the future of planetary traverses. These programs and projects all contribute to the process of planning and executing a crewed Martian traverse. This report summarizes the traverse planning method as specifically used by D-RATS, introduces some of the significant changes in culture that must occur for the execution of a crewed Martian traverse, and outlines a high level architecture for this new mission operations concept.

The Traverse Planning Model: The primary goal of traverse planning is to maximize the science return of an Extra Vehicular Activity (EVA) within the greater mission objectives (1). The traverse planning process is a very integrated effort requiring the input of scientists, engineers, mission operations, and the crew. Each traverse targets a series of science objectives as well as engineering and flight operations objectives (2). However the traverses must remain within the given operational and technical constraints of the available hardware, environment, and crew ability (Fig A). The process begins with a detailed analysis of high resolution satellite imagery of the traverse region (3). From this imagery, detailed photogenic maps are created, and the areal extents of geologic formations are interpreted. These maps inform the creation of a series of specific science objectives and questions. The science objectives are then ranked in order of priority and sites where these questions can be addressed within the traverse region are identified. It is from these sites that the traverse is determined.

In the context of D-RATS 2010, this pre-work was done by the USGS, and informed the first traverse planning workshop. At this workshop an integrated team formalized the science objectives, and specified the first order operational constraints (1). Using communication constraints such as the deployment of equipment on topographical highs, a series of preliminary traverses were determined. During the process several operational questions arose that could not be answered with remote sensing such as the location of gates, fences, property lines, and 18-wheeler road access (1). In order to address these issues a field survey was tasked with all mission critical operations represented. Answering these questions lead to major revisions in a few of the preliminary traverse paths (1). Further iterations of traverse planning took place, each in more detail, until there were time tables specified down to the 5 minute mark (1).

Mission Control vs Mission Support:

Communication and Control: The largest difference between the current experiences and future
human missions to Mars will be the operations control structure. The Apollo mission style set the standard for how current crewed missions are run: directly connected to the Mission Control Center (MCC) in Houston, with every step broadcasted and followed by a team of support scientists and engineers. This “backroom” method has a proven value to human field science. A lesson learned from the D-RATS missions was that having well-trained geologists in the field supported by a science backroom helps to maximize this science return (4). However the time delay between Mars and the Earth can reach up to 20 minutes. The D-RATS analog missions experimented with time delay and different communication architectures and discovered that even a few second delay was nearly prohibitive to real time mission assistance from Earth. This was due to overlapping transmissions and the resulting requests for broadcast. Data management will be an important issue and effectively collecting, packaging, and dissemination of the traverse data will need to be addressed. Lessons learned from the D-RATS missions suggests that having team members with these specific skills as well as major automation of these tasks would be of measurable benefit (5). A centralized geo-referenced data portal for Martian traverse information should be developed and made accessible to the field crew, the Earth Base, and the general scientific community.

Choreography vs Flexibility: Apollo mission traverses were very well choreographed before launch but involved only a few second communication delay. This allowed for MCC to direct the traverse. Due to the complexity of the mission and limited durations there was no time planned for deviations from the EVA schedule (4). However the nature of field science is an evolving exploration science, and as insights are gained the new knowledge informs the next decision. This lead to situations where the crew took actions that were not pre-approved by MCC, such as Apollo 15 astronauts inventing a seatbelt malfunction in order to collect an interesting basalt sample that was off of their EVA path, or the Apollo 17 astronauts shifting priorities to investigate orange glass they discovered for longer than the apportioned EVA period (4). An understanding of the broad science objectives allows the field crew to respond to unexpected developments appropriately. From these experiences it was learned that building in “flex-time” into a traverse can help maximize science return by allowing sufficient time for the field crew to perform the scientific process (4).

Strategic vs Tactical: The Mars Explorer Rover model of operational concepts must account for delayed control of the mission. The MER team separates their operations into tactical and strategic levels. The tactical team had direct fine detail control of the rover, and the strategic team is responsible for the long term high-level planning (2). Robotic Mars missions present longer timelines due to communication delay and the lack of a human crew. The tactical team focused on operating the daily scientific tasks, while the strategic team completed long term planning for the totally mission duration (5). The 2010 D-RATS mission applied this division as a Tactical Science Operations Team (TSOT) and a Strategic Science Operations Team (SSOT) (2). The tactical level or TSOT for each of the two rovers was responsible for actually commanding the traverses, and synthesizing the daily data (2). Both rovers’ reports were passed off to the single SSOT that met through the night and tracked the overall science objectives (2). The SSOT informed the TSOT’s if they needed to alter the remaining traverses (2). Future crewed Mars missions will require a delicate balance of real time tactical support and long term strategic planning. The solution is placing most of the tactical duties, and emergency response, in the hands of the local crew and relying on MCC to fill the role of strategic planning and non-immediate emergency support.

The Crewed Martian Traverse Model:

Mission Hardware: Planning a crewed Martian planetary traverse will involve all areas of expertise and experience. Once a landing site has been selected the traverse region can be defined. This region is limited by operation constraints such as consumables, power, communication, and the available equipment. Having a pressurized rover increases the amount of terrain covered as well as the safety of the mission. In order to best perform the preliminary remote sensing of the region survey satellites will need to be tasked. Placing an orbiter or orbiters in a polar orbit over the traverse region will allow for a more focused regional study, as well as set up a communication array to support the mission. Similar to the 2010 D-RATS experience, the preliminary remote sensing may leave outstanding questions about the region. These can be answered by a field study conducted by robotic rovers. Sending robotic rovers ahead of a human mission will insure that the maximum amount of information is known about the region’s conditions, resources, and terrain. These rovers should survey the landing site, the base station, and the approximate traverse course. Once the crew arrives they can service and continue to use these rovers for their surface operations. Air-breather exploration may serve the same purpose. The 2010 D-RATS mission also incorporated a mobile field lab, the Habitat Demonstration Unit, to simulate returning to a Martian habitat for more detailed examination of samples. The Mars base will be important as a staging and return point for the traverses, and should include more specialized experimentation equipment as well as facilities for sample cache and curation.
**Standardization of Traverse Procedures & Training:** Whereas in Apollo missions the astronauts could rehearse every minute of the traverse before arriving, the added complexity of several multi-day traverses with pressurized rovers that may occur weeks apart makes this practice impractical. Instead astronauts must train to perform standardized traverse procedures, and then apply these skills to the variety of traverses. This coupled with training in the regional geology will increase the onsite value of the field crew. This is also why professional geologists would be a large benefit for any planetary traverse. A crewed mission on Mars could be anywhere from a few days to six months. A crew that is on the Martian surface for extended time will have the opportunity for multiple traverses, revisiting sites, and using the knowledge gained from early traverses to inform later plans. The field crew will become the subject matter experts on that region of Martian geology, and will be uniquely qualified to plan and carry out the traverses. The later traverses therefore will end up being edited, altered, and potentially even generated by the field crew.

**Mission Operations Concept:** The team or teams of the field crew will operate out of pressurized rovers that will provide the crew with a shirt sleeve lab environment while between traverse sites. The field crew will include at least one trained Geologist, and one crew member with more familiarity with EVA and/or the rover systems. The rover or rovers will stay in direct, real-time communication with the Mars Base via a communication satellite array or repeater stations. The Mars Base, which can be on the surface, a near by moon, or in Martian orbit, will act as the real-time, human-in-the-loop, tactical team for the field crew. The Mars Base crew will also provide safety monitoring and mission operations support to the field crew. The tactical team should include a trained physical scientist(s) to provide science support, as well as a crew member whose primary focus is data management. This crew member will be responsible for capturing all of the data produced by the traverse, pictures, locations, transcripts, etc, and packaging it for transmission to Earth. The more this process is automated the better the crew can utilize their time. In incremental amounts the Mars Base will transmit the data packages, comments, and results back to Earth. Once received these data packages will be reviewed, cataloged, and disseminated. The Earth Base will follow the model of Mission Support, providing extra information or expertise when requested and will participate in the strategic level of traverse operations (Fig B). After the traverse is completed the returning field crew will meet with the Mars Base crew and they will review the mission. Together they will create a formal finalized report for Mission Support, much the same way an exploration Geologist would for a resource company or research institute. Based on the discoveries and new information Mission Support may draft sequential traverse plans, but these will be sent to the crew who will edit, and adjust the traverses to their own specifications. This process will repeat for the duration of the mission.


**Additional Information:** In addition to the above references insight for this article was gleaned from interviews with the following individuals: Dean Eppler, Friedrich Hörz, Farouk El-Baz, and John Gruener, as well as guidance from Lee Graham and David Melendrez of NASA’s Johnson Space Center.
AN INVESTIGATION OF CALCIUM CHLORINE OXOANIONS ROLE IN THE EVOLVED OXYGEN RELEASE FOUND BY MARS CURIOSITY ROVER SAMPLE ANALYSIS AT MARS (SAM) INSTRUMENT AT ROCKNEST EOLIAN BEDFORM. A. Bruck* and B. Sutter**, 1Department of Chemistry, Illinois State University, Normal, IL 61761. (ambruc2@ilstu.edu), 2Jacobs Engineering Technology and Science/NASA-Johnson Space Center, Houston, TX 77058 (brad.sutter-2@nasa.gov).

Introduction: A major oxygen release between 300 and 500°C has been detected by the Mars Curiosity Rover Sample Analysis at Mars (SAM) instrument at the Rocknest eolian bedform. Decomposition of perchlorate (ClO$_4^-$) and/or chlorate (ClO$_3^-$) salts in the Rocknest samples is a possible explanation for this evolved oxygen release [1]. The existence of perchlorate salts is confirmed by Hecht et al. [2] from the results of the Mars Phoenix Lander’s Wet Chemistry Laboratory (WCL). However, the detection of the perchlorate salts may have masked chlorine anions that are known to naturally accompany perchlorate in various environments [3,4]. This implication led to a suite of experiments conducted on perchlorate/chlorate species in the JSC-SAM-testbed to evaluate decomposition temperatures. Calcium perchlorate was determined to be the most likely candidate out of the chlorine oxoanions because its decomposition and corresponding O$_2$ release temperatures (400-500°C) were closest to the O$_2$ release temperatures observed for the Rocknest material. Furthermore, calcium perchlorate released small quantities of Cl that could have been the source of Cl in the chlorinated-hydrocarbons species that were detected by the SAM quadrupole mass spectrometer (QMS) and gas chromatography/mass spectrometer (GCMS) [1].

Materials and Methods: Reagent grade ACROS ORGANICS Ca(ClO$_4$)$_2$·4H$_2$O, pure was dissolved in deionized water and flash frozen with N$_2$(l). It was then allowed to sublime under vacuum with a Labconco Freezone 4.5 Plus evaporator to dryness. When dehydrated the samples were transferred to Plas labs controlled atmosphere chamber for storage under N$_2$(g). City Chemicals Ca(ClO$_3$)$_2$, purified was purchased and stored in the Plas labs controlled atmosphere chamber. These samples were prepared in inert atmosphere conditions (< 100 ppm H$_2$O present), with ground an agar mortar and sieved to ensure that all samples were ≤ 125 microns in particle size.

Synthetic magnetite (MTS4), synthetic hematite, and fayalite with a magnetite contaminant from the Forsyte Iron Mine in Ontario, Canada (# 49 V 1555) were ground with an agar mortar and pestle and sieved to ensure particle size ≤ 125 microns. HWMK919, a Mauna Kea Volcano dust sample known to be a Martian soil analog [5] (≤ 5 microns), and College Station-Rockshop pyrite (≤ 125 microns) were also obtained and all Fe-bearing minerals were stored in standard laboratory conditions.

Each mineral used as a mixture represented a different component in the Martian soil. Synthetic hematite a fully oxidized iron, and magnetite (MTS4) a partially oxidized iron, represented the iron oxides and are known to exist in the Rocknest samples [6] and could catalyze perchlorate/chlorate thermal decomposition. Fayalite an olivine analog, pyrite is a sulfur containing iron species, and HWMK919 contained amorphous material and similar components of those found at Rocknest.

Simulated SAM analysis were conducted on a Setaram Sensys-Evo differential scanning calorimeter (DSC) coupled to a Stanford Research Systems Universal Gas Analyzer (MS) at Johnson Space Center which has been configured to operate similarly to the SAM oven/QMS system [1]. Experiments on the JSC-SAM-testbed operated at 30-75 mbar, at a heating rate of 35°C/min taken to 715°C maximum temperature, with He carrier gas and a flow rate of 3 mL/min. Samples were prepared by weighing out 1 and 5 fold molar quantity of magnetite, hematite, fayalite, or pyrite based on calcium perchlorate (3.216 μmol) or calcium chloride (3.390 μmol). The HWMK919 sample was prepared with 0.5 or 2.5 mg to mimic the relative amounts of the molar quantities used for the other Fe-bearing minerals, but its molar ratio could not be determined. Once prepared, an absolute amount of the chlorine oxoanion of interest was removed from the controlled atmosphere chamber, weighed in standard laboratory conditions then mixed thoroughly with the Fe-bearing mineral of interest and placed into the JSC-SAM-testbed furnace for analysis.

Thermal Decomposition and Catalysis: The various individual chlorine oxoanion species that were tested in the JSC-SAM-testbed, could not explain the evolved oxygen temperature found by SAM at the Rocknest eolian bedform [1]. If the evolved oxygen detected by SAM is from calcium perchlorate or a different chlorine oxoanion species, reactions with another component in the soil may be occurring that would increase the decomposition rate resulting in O$_2$ release temperatures consistent with those detected in the Rocknest materials. The objective of this work is to investigate three different interactions, 1) catalytic interactions on calcium perchlorate from various iron-bearing minerals, 2) catalytic interactions on calcium...
chlorate from various iron-bearing minerals, and 3) a calcium perchlorate/calcium chlorate mixture interaction.

**Calcium perchlorate:** Reagent grade calcium perchlorate typically is only found in its hydrated state, Ca(ClO$_4$)$_2$•4H$_2$O, calcium perchlorate tetrahydrate. It was apparent that in standard laboratory conditions further hydration occurred and therefore can be represented by Ca(ClO$_4$)$_2$•nH$_2$O with n unknown. A study of its tetrahydrate form thermal decomposition by Migdal-Mikuli et al. [7] provided the following reaction scheme based on TG, DSC, and QMS techniques.

\[
(1) \text{Ca(H}_2\text{O)}\text{2} \rightarrow \text{[Ca(H}_2\text{O)}\text{2]}(\text{ClO}_4)\text{2} + 2\text{H}_2\text{O}
\]

\[
(2) \text{[Ca(H}_2\text{O)}\text{2]}(\text{ClO}_4)\text{2} \rightarrow \text{Ca(ClO}_4)\text{2} + 2\text{H}_2\text{O}
\]

Equation (2) also corresponds to the melting of calcium perchlorate. Then equation (3) represents the final decomposition at higher temperatures [7].

\[
(3) \text{Ca(ClO}_4)\text{2} \rightarrow \text{CaCl}_2 + 4\text{O}_2
\]

This reaction mechanism correlates with the data provided in Figure 1. The first endotherm corresponds to equation (1) (≤ 150°C), the second endotherm equation (2) (≤ 400°C), and the large exotherm equation (3) (≥ 400°C), Also presented in Figure 1 is the catalytic effect of 1 fold of hematite. The decomposition rates of various alkali metal perchlorates are known to increase in the presence of a catalyst [8, 9, 10]. Hematite, a well-known iron oxide that is found in the Martian soil also provided a significant decrease on the temperature difference with the 1 to 5 fold mixture of magnetite and an absolute amount of calcium chlorate differed by 50°C.

**Calcium chlorate:** Calcium chlorate exists in a hydrated state as proven by the results from the JSC-SAM-testbed which evolved a significant amount of water (m/z 18), and is therefore regarded as Ca(ClO$_4$)$_2$•nH$_2$O with n unknown. This species has little supplementary literature on its thermal decomposition. However, there have been numerous studies on the alkali metal chlorate species and on the catalytic effect of metal oxides [11, 12]. This reaction differs from the decomposition scheme of perchlorates by a perchlorate and chlorite intermediate, the scheme is listed below with M = alkali metal [11].

\[
(4) \text{2MClO}_3 \rightarrow \text{MClO}_4 + (\text{MClO}_2) \text{ (slow)}
\]

\[
(5) \text{(MClO}_2) \rightarrow \text{MCl + O}_2 \text{ (fast)}
\]

\[
(6) \text{2MClO}_3 \rightarrow \text{MClO}_4 + \text{MCl + O}_2 \text{ (overall)}
\]

This suggests that there is an oxygen transfer between two chlorate anions and provide a stable perchlorate intermediate and an unstable chlorite intermediate. This does not prove that this is the case for calcium chlorate, but it does provide an explanation for the significant difference in decomposition of the two species. In Figure 1, the first endotherm (≤ 200°C) is associated with dehydration because of a large water (m/z 18) release at the same temperature (data not shown). The second endotherm is assumed to be melting of calcium chlorate (≤ 300°C) and the final exotherm(s) (≥ 400°C) is accompanied by the oxygen release.

**Results and Discussion:** Varying molar quantities was essential because the Martian soil contains varying ratios from 1 to approximately 20 fold of different iron bearing minerals [13]. With the calcium perchlorate, the increase from 1 to 5 fold Fe phases had little effect on the decomposition temperature (Fig. 2). However, when increasing the amount of iron oxide in the calcium chlorate mixtures, a decrease in the temperature of the evolved oxygen release was observed (Fig. 3). The temperature difference with the 1 to 5 fold mixture of magnetite and an absolute amount of calcium chlorate differed by 50°C.

After an adequate amount of data was collected of calcium perchlorate mixtures it was concluded that calcium perchlorate interactions with the soil could not account for the oxygen release detected by SAM, as shown by Figure 2. The lowest decomposition temperature found was approximately 445°C catalyzed by magnetite. This result led into an investigation of calcium chlorate with various iron-bearing minerals.

Pure calcium chlorate had a much higher decomposition temperature at approximately 565°C and has a broader decomposition range than calcium perchlorate. When mixed with the iron oxides it caused a drastic decrease in decomposition temperature (~150°C). This decreased temperature falls in the range of peak oxygen releases detected by SAM and is therefore the most likely calcium bearing candidate investigated as
Figure 2: Evolved oxygen gas data from the SAM instrument compared to the MS m/z 32 analysis performed on the JSC-SAM-testbed. Calcium perchlorate was kept at an absolute amount, the varying numbers following the mineral is the fold amount or weight amount in mg for HWMK919.

shown by Figure 3. Figure 3 also shows magnetite, hematite, and HWMK919 – calcium chlorate mixtures all have their peak oxygen release within the range of that found by SAM at Rocknest. This result was confirmed with duplicate trials that are not shown. The pyrite-calcium chlorate mixture is within the range of the minor oxygen release of SAM and was also reproducible during trials. This sharp peak at approximately 250°C also evolved large quantities of SO₂ (m/z 64), H₂S (m/z 34) that could be an isotope of oxygen, and CO₂ (m/z 44). The DSC thermogram data provided a large exotherm (> 40 mW) that correlated to the oxygen release at approximately 250°C.

Fayalite had little effect on the perchlorate and chlorate species. However, with the calcium perchlorate/chlorate mixtures it showed evidence of HCl (m/z 36) and Cl (m/z 35) releases at higher temperatures when mixed at higher folds (data not shown). Evolved chlorine species were mainly observed with the HWMK919 mixtures and also increased with additional HWMK919 (data not shown). Although evidence of evolved chlorine was previously reported [1] this investigation found no evidence during decomposition of pure calcium perchlorate or calcium chlorate.

Varying perchlorate/chlorate mixtures were also attempted with inconclusive results. When mixing calcium perchlorate and calcium chlorate, three drastically different decomposition temperatures were recorded. This is likely explained by the hydroscopic tendencies of the two salts. When mixing they become too hydrated to mix efficiently. In future this mixture should be investigated in a controlled atmosphere that would allow easier handling of the substances.

Conclusions: Iron oxides (magnetite and hematite) have the most effect on decomposition temperature of calcium perchlorate/chlorate species. However, when investigating the catalytic tendencies of these iron bearing minerals with calcium perchlorate, none of the components from the Martian soil analyzed resulted in a similar oxygen release as the SAM evolved gas analysis data. Calcium chlorate-Fe phase mixtures were analyzed and provided results that simulated the oxygen release temperatures detected by SAM. Evolved chlorinated species were only detected during analysis of mixtures of HWMK919 and the calcium perchlorate with fayalite mixtures. Future work will evaluate magnesium and sodium chlorates as potential sources of evolved O₂ from Rocknest. Heated X-Ray Diffraction (XRD) analysis will provide data on reactions occurring during heating as well as provide a better understanding of the mechanisms occurring in future experiments.

OLIVINE EXPOSURES ON THE MOON: ORIGINS AND MECHANISMS OF TRANSPORT TO THE LUNAR SURFACE. L. M. Corley¹, P. J. McGovern², and G. Y. Kramer², ¹Hawaii Institute of Geophysics and Planetology, University of Hawaii at Manoa, Honolulu, HI 96822, USA (lmc44@hawaii.edu), ²Lunar and Planetary Institute, Houston, TX 77058, USA

Introduction: According to the lunar magma ocean theory, the Moon’s crust and mantle crystallized from a magma ocean. Olivine crystallized early and throughout much of the crystallization of the mantle. Spectral data have recently allowed for the detection of olivine located at large impact basins (Fig. 1) [1, 2]. Olivine exposures at the surface may be exposures of the mantle but could also be exposures of shallow intrusions that solidified from melt generated in the mantle. In either case, olivine exposures can provide insight to the structural and magmatic evolution of the Moon.

Yamamoto et al. [1] used data from KAGUYA’s Spectral Profiler (SP) to identify olivine located on rims and central peaks of large impact basins, including Crisium, Humorum, and Nectaris. They concluded that the olivine spectra were from dunite, and thus likely originated in the upper mantle. Powell et al. [2] examined Crisium basin using Chandrayaan-1’s Moon Mineralogy Mapper (M³) spectra. They detected olivine on mare, intrusive landforms, and on crustal maxima, all which indicate magmatic transport. This conclusion is supported by findings of McGovern and Litherland [3] showing that areas around large basins are favorable for magma ascent via dikes.

Methods: We identified olivine on the Moon at Crisium, Nectaris, and Humorum basins and near the crater Roche, using M³ data (Fig. 2). Unlike SP, M³ is an imaging spectrometer that allowed us to link potential olivine spectra to geologic features. 85 bands from 460 to 2980 nm give M³ high spectral resolution that can be used to distinguish mineral signatures. We used level 2 M³ reflectance data, which is both thermally and photometrically corrected [4]. For our olivine investigation, we used an olivine index based on the algorithm developed for the imaging spectrometer CRISM onboard the Mars Reconnaissance Orbiter [5], but optimized for M³ wavelengths.

Olivine spectra have three absorption features at 0.85, 1.05, and 1.25 µm. In M³ spectra, these three absorption features can be easily recognized as one wide composite absorption band centered near 1 µm (Fig. 1). This is the only absorption band for rocks that are composed mainly of olivine. However, a small amount of pyroxene in a rock will result in an additional absorption band centered near 2 µm. We classified our spectra with a weak 2-µm band as olivine-dominated pyroxene mixtures. Based on work by Singer [6], the spectra we classified as olivine-dominated mixtures likely had no more than 25% pyroxene.

![Figure 1: Spectra from this study of Mg-dominated and Fe-dominated olivines, and an olivine-dominated pyroxene mixture.](image)

Results:

Crisium. At Crisium basin we identified 62 spectra with signatures that represent olivine or olivine-dominated mixtures. These spectra are located on the...
rims of small craters, on massifs at the rim of Crisium, and on mare both inside and outside of Crisium basin.

We were able to confirm many olivine locations that were detected by [1, 2], including Lacus Perseverantiae and a potential dike at Eimmart A. However, there are a few areas where we were unable to confirm their detections. We report several olivine detections where olivine was previously undiscovered, including on the main basin-filling mare of Crisium. One such discovery is located on the rim and in the ejecta of Picard crater. Crustal thickness models [8] based on GRAIL measurements and LOLA topography indicate that Picard crater is large enough to have penetrated through the thin crust. In contrast, many of our olivine detections correlate with crustal thickness maxima.

Nectaris. Nearly all of the olivine detections at Nectaris are confined to the mare. Our investigation confirmed every olivine location detected by [1], including the rim and central peak of Theophilus. In addition, we were able to find many more olivine locations, for a total of 45 olivine spectra. Although crustal thickness inside Nectaris basin is as low as 5 km in some areas, none of the craters where olivine is found are large enough to have penetrated through the crust (Fig. 3).

Humorum. A total of 8 olivine spectra were detected at Humorum basin. In this investigation, only one of three olivine locations detected by [1] was confirmed. A new olivine discovery was made at Lee crater, where olivine appeared to be Fe-dominated. In addition, two Mg-dominated olivine spectra were found on a graben in northwest Humorum. All of the olivine detections at Humorum are located on its rim.

Roche. Roche crater, located on the lunar farside, had not been previously investigated for olivine using high spectral resolution data. Work by Andrews-Hanna et al. [11] identified a linear gravity anomaly at Roche using GRAIL data, which they interpreted to be an ancient dike. We identified 9 olivine spectra near Roche. Several of our detections, both Fe-dominated and Mg-dominated spectra, were in close proximity to the proposed dike.

Discussion: Our findings have important implications for the origins of the exposed olivines and the mechanisms of transport to the lunar surface. Possible transport mechanisms include mechanical transport of mantle or lower crustal material by basin-forming impact, or magmatic transport of cumulates or xenoliths.

There is a diversity of origins for olivines located on maria within and around Crisium. Our examination
of the geophysical setting at Crisium basin suggests that olivine exposed on the main mare at Picard crater may be primary olivine from the mantle. The crustal thickness inside Crisium basin is thin enough that the 23-km diameter Picard crater likely penetrated through the crust, exposing mantle material. Olivines located at Lacus Perseverantiae were likely transported magmatically because here the crust is too thick for the crater where olivine is found to have penetrated the mantle. Other olivines are located on the rim of Crisium near local crustal maxima, which may indicate that mantle material was excavated by basin impact and scattered over the surface as a veneer [12]. However, these olivines may have instead been transported by magmatic processes. Consistent with [2], we identified olivine on the rim of Eimmart A where there is morphologic evidence for a potential dike.

Olivine exposures in Nectaris basin are confined to the rims of small craters that penetrated the mare. These craters are not large enough to have penetrated the estimated thickness of the mare [8]. Instead, these small craters probably exposed an olivine-rich mare basalt or olivine cumulates of the Nectaris basin impact melt.

At Humorum basin, the grabens exhibiting olivine sites were likely created by extensional stresses caused by the loading of mare basalt [3, 13]. Extensional stresses would also have been favorable to the formation of dikes. Thus, we conclude that olivines detected at the graben in northwest Humorum were likely transported by magmatic processes.

Due to the thicker crust on the lunar farside, mantle material near Roche crater was not likely exposed by the Roche impact. However, the presence of olivines at this location suggests that magmatic intrusions reached the shallow subsurface and were exposed by small impacts. Detection of olivine is consistent with the presence of a dike near Roche [11].

THE GALE CRATER MOUND IN A REGIONAL GEOLOGIC SETTING: COMPARISON STUDY OF MT. SHARP AND 1000 KM RADIUS. A. Dapremont1, C. Allen2, and D. Oehler2, 1Department of Geology and Environmental Geosciences, College of Charleston, Charleston, SC 29424 (amdapremont@gmail.com), 2Astromaterials Research and Exploration Science, NASA-JSC.

Introduction: Gale Crater is a Late Noachian/Early Hesperian impact crater located on the dichotomy boundary separating the southern highlands and the northern lowlands of Mars [1,2]. NASA’s Curiosity Rover is currently exploring Gale crater, searching for evidence of habitability early in the planet’s history. With an approximate diameter of 155 km, and a ~ 5 km central mound informally titled Mt. Sharp, Gale represents a region of geologic interest due to the abundance of knowledge that can be derived, through its sedimentary deposits, pertaining to the environmental evolution of Mars [2][Fig. 1]. This study was undertaken to compare the mineralogy of sedimentary deposits and wind erosion features of Gale Crater, Mt. Sharp, and a 1000 km radial area.

Phyllosilicates are known to be present at Gale in the crater’s lower mound (LM), and spectral signatures are consistent with nontronite, an iron (Fe)-smectite [2]. Monohydrated (kieserite) and polyhydrated (likely magnesium-rich) sulfates have been identified in the lower mound as well [2]. Previous studies of the upper mound (UM) indicate a deficiency in hydrated minerals [1]. The sedimentary package at Gale is indicative of an overall transitional sequence from phyllosilicates to sulfates which correlates with hypotheses indicative of a shift from wet climatic conditions early in the planet’s history to a dry, oxidative state within the past several billion years [2].

Large scale aeolian erosional features called yardangs are a proxy for prevailing wind direction over an extended period of time. Yardangs on the Martian surface have Earth analogues in numerous locations such as central Asia and the Sahara Desert [3]. These wind-carved products are formed in a variety of rock types on Earth including sandstones and basalt [3]. Thomson et al. [1] mapped several yardang units in the LM and UM of Gale, and these features have also been noted in studies of the Medusae Fossae Formation (MFF) near the dichotomy boundary [4, 5]. The MFF is thought to be formed from volcanic ash [5]. Yardangs are valuable tools for correlation studies of Gale and the surrounding regional geology.

Methodology: Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) ir_phy browse products were evaluated for aluminum and iron/magnesium phyllosilicate signatures. Altitude values were obtained using ArcMap10.1 software and the full resolution Mars Orbital Laser Altimeter (MOLA) data prepared by the U.S. Geological Survey in the Mars DVD v. 1.6. High Resolution Imaging Science Experiment (HiRISE) and Mars Orbiter Camera (MOC) images of phyllosilicate locations, as well as yardang identification sites, were acquired for more detailed examination of geologic features present. These locations and altitudes were compared with areas of interest from Korn and Allen [6] outside of Gale which were defined using night-time thermal infrared brightness, age, and altitude.

Wind directions of yardangs inside and outside of Gale were mapped using the Mars Reconnaissance Orbiter (MRO) Context Camera (CTX) mosaic in GoogleMars [Fig. 2]. Geologic map units of Thomson et al. [1] were used to define mound boundaries. Image overlays were created to correlate Thomson et al.’s units with mapped yardang units inside the crater.

Results: Phyllosilicates and Aeolian Patterns Inside Gale. Phyllosilicate signatures of the upper mound (UM) were found within an altitude range of -2400 to -100 m with weaker signatures at higher altitudes [Fig. 3]. These signatures were found below elevations of ~2900 m in the lower mound (LM) [Fig. 3]. In many cases, phyllosilicate concentrations corresponded to yardangs and cliff forming units [Fig. 4]. The UM exhibited small...
scale yardangs with diverse orientations (N-S, NE-SW, and NW-SE) [Fig. 2]. In contrast, the larger yardangs of the LM, with widths of hundreds of meters and lengths of several kilometers, exhibited a more consistent nearly N-S orientation, suggesting a N-S prevailing wind during their formation. [Fig. 2].

Fig. 2. Wind directions (determined form yardang orientations) of LM (black) and UM (white) overlain on CTX mosaic.

Phyllosilicates and Aeolian Patterns Outside Gale. Within the 1000 km study area outside of Gale crater, aluminum-rich phyllosilicate signatures were found over an altitude range of +1726 to -2734 m. Within this region ridges, crater rims, valley floors, and uplifted areas displayed phyllosilicate signatures. The latter was determined through detailed analysis of CRISM imagery. Several locations where phyllosilicate signatures were abundant were identified within the areas of interest of Korn and Allen [6].

Yardangs outside of Gale were predominantly located to the northeast, in the northern plains units, and along the dichotomy boundary. Yardang orientations included NW-SE, NE-SW, N-S, and E-W. Many yardangs exhibited aluminum and iron/magnesium-rich phyllosilicate signatures [Figs. 5, 6]. The altitude range over which these signatures and yardang locations correlated was -979 to -2846 m.

Fig. 3. MOLA elevation of Gale. Black circles indicate aluminum phyllosilicate locations and white squares indicate aluminum and iron/magnesium phyllosilicate locations.

Fig. 4. Infrared surface brightness image from CRISM browse products showing large, nearly N-S wind erosion features (yardangs) of LM of Gale (CRISM M: FRT00018285). Strong spectral signatures indicated aluminum and iron/magnesium phyllosilicate presence associated with erosion features concentrated in top and middle of image.
**Discussion/Conclusion:** *Phyllosilicates.* CRISM data illustrate phyllosilicate presence inside and outside of Gale crater across various altitudes. Due to the strength of phyllosilicate signatures in the lower mound, it is probable that these units were altered in the presence of water. Due to the weakness of phyllosilicate signatures at high altitudes of the upper mound, these data are consistent with deposition of these sediments in a dryer period.

* Aeolian erosion: Differences in wind directions deduced from yardang orientations in the lower and upper mounds indicate changes in prevailing wind direction over the course of Gale’s history, which suggests a significant gap in time between these two parts of the mound. It is possible this shift is related to an overall climatic change.

The correlation established between phyllosilicate presence and yardangs of Gale crater and the surrounding area may imply that these erosional features are formed from easily-altered material such as volcanic ash. Differences in prevailing wind direction and alteration mineralogy of the LM and UM of Gale imply that the former was emplaced and eroded well before the latter. Future investigations should pursue the aforementioned relationships to further understand the crater’s history. Discoveries of the Curiosity Rover, currently en route to Mt. Sharp, will aid in a more comprehensive understanding of Gale’s past.

COMPARING GEOLOGIC DATA SETS COLLECTED BY PLANETARY ANALOG TRAVERSES AND BY STANDARD GEOLOGIC FIELD MAPPING: PRELIMINARY RESULTS. Wanda Feng¹, Cynthia Evans², John Gruener², and Dean Eppler², ¹Departments of Geology and Astronomy, Smith College, Northampton, MA (wfeng@smith.edu), ²NASA, Johnson Space Center, Houston, TX 77058 (cindy.evans-1@nasa.gov, john.e.gruener@nasa.gov, dean.b.eppler@nasa.gov).

Introduction: Geologic mapping involves visualizing an area in 3D and using the present landforms and compositions to convey formative history. Traditional field techniques are used to accomplish this on earth. Mapping proves more challenging for other planets, which are studied primarily by orbital remote sensing and, less frequently, by robotic and, human surface exploration. Use of these differing techniques induces inherent discrepancies, which have not been previously evaluated. The objective of this project is to produce a geologic map from data collected on the Desert Research and Technology Studies (RATS) 2010 analog mission using Apollo-style traverses, and to compare this with a geologic map of the same area produced using standard field geologic techniques.

Background: The Apollo Missions (1961-1972) yielded data that revolutionized our understanding of lunar geology. No subsequent planetary human exploration has been conducted since the end of the Apollo Program; consequently, there has been no opportunity to field check Apollo results with more detailed field investigations of the study sites. The Desert RATS missions (1997-present) have been conducted in northern Arizona to exercise science operations, test multi-mission space exploration vehicles (MMSEVs) and EVA protocol to prepare for future human exploration [1]. Since 2009, these analog tests have used “Apollo-style” traverse planning and extravehicular activities (EVA) to understand regional geology and sample conjectured geologic units [2-3], but the strengths and weaknesses of this style of planetary exploration have yet to be examined.

The most extensive RATS mission was completed in 2010 in the San Francisco Volcanic Field north of Flagstaff, AZ. Over the course of 14 days, a 580 km² area was explored by 2 prototype pressurized rovers with crews of astronauts and geologists [3]. Two communication modes between the rovers and the support scientists (“backroom”) were utilized: continuous communication (CC) with the backroom and twice a day (2/Day) communications, leaving crew members to operate autonomously, and downloading their observations and data at the end of the day for backroom review [3-4]. Overall, 448 samples were collected from 69 EVA stations [5]. This study took a 15 km² field area adjacent to SP mountain that was visited by RATS 2010 crews to evaluate in further detail using RATS data and data derived from standard geologic mapping techniques. Although this area has been studied at a reconnaissance scale [6-7], detailed geologic mapping has not been carried out on this volcanic center.

Methods: Much like the Apollo Missions, RATS yielded samples, photographs, and crew videos for further study with remote sensing data. In the study area, 19 EVA stations from 5 days contributed to 122 samples. Of these, 9 stations with 58 samples were taken on CC days. Data and samples were collected with guidance from the backroom scientists. The remaining 10 stations with 64 samples were from 2/D days [5]. The data for these samples was acquired from the downloaded RATS data (videos, GPS tracks, and photos).

The sample locations were georeferenced using crew videos, field photos, and GPS data from the EVA backpacks using Google Earth and Picassa. These locations, as well as crew and rover traverses, were mapped for each station in ArcMap (Fig. 1). The sample contexts were inferred from the ESRI world imagery basemap, which has a 1-m resolution. A GeoEye-derived digital elevation model (DEM) of the area that was provided by the USGS in Flagstaff was used to create contours and extract elevation data.

Figure 1: 1:1500 image of site 25B with sample numbers, EVA track (green), rover track (red). The basemap is ESRI world imagery, including an aerial view of the MMSEV.

Sample contexts were important for characterizing composition and mineralogy, ultimately allowing for differentiating between different sedimentary and vol-
canic units. The age relationships of the units were finally determined by geomorphologic relationships and relative weathering of the samples.

**Results:** The 1:24,000 scale product map (Fig. 2) was created in ArcMap (ArcGIS 10.1) using the World Geodetic System (WGS) 1984 WebMercator projection. The photogeologic pre-mission units [2] were re-evaluated using RATS 2010 data.

The field area included a basement sedimentary unit, several different volcanic flows, and cinder cones. The sedimentary basement units, originally mapped as two distinct units, were determined to be one unit of limestone (ls) interbedded with limey siltstone and sandstone. Distinguishing between the different volcanic flows and cones proved difficult. The samples indicated that all of the flows were basalts. The oldest identifiable basalt flow (b1) was interpreted by photo-interpretation; however, due to the lack of samples and distinguishing topography, the number of flows is indeterminate. The b1 flow was inferred to be older than the other basalt flows by superposition.

The central volcano complex was mapped as two cones. On the basis of RATS 2010 samples, the northern cone (p1) is composed of massive lava flows with olivine and pyroxene phenocrysts, while the southern cone (p2) is composed of pyroxene- and plagioclase-phenocryst dominated agglutinates. The topography and sample compositions suggest that p1 is related to and constructed on top of the adjacent flow (b2).

The southeastern corner of the study area was interpreted as an older, weathered flow (b4) of a plagioclase-rich agglutinate basalt. The westside of the study area comprises another basalt flow (b3) composed of massive, vesicular basalt rich in olivine and pyroxene phenocrysts. The morphology of the b3 flow suggests that it is younger than b4. The clear lobate features of both the b3 and b4 flows in proximity to the central volcano complex supports that the flows are relatively younger in age. An area with elevated horizons along the eastern edge of b4 was sampled and is composed of basalt with plagioclase phenocrysts and pyroxene xenocrysts, similar to the p2 cone. The RATS data indicate these features were subsequently surrounded by the b4 flow. SP flow (b5) and SP mountain (p3) are the most recent volcanic features in the study area. Both units are composed of massive and vesicular basalt rich in olivine and plagioclase. The map boundaries for these units remain unaltered from the pre-mission map.

The interpreted boundaries and types of the surficial units (originally referred to as surficial plains) were difficult to validate based on the RATS data. The limited data and photo-interpretation results suggest that the units include alluvium (al) and colluvium (cl) with possibilities for eolian deposits and ashfall.

**Discussion and Conclusions:** The RATS 2010 data is spatially limited, as manifested on the map (Fig. 2). Mapping the study area has therefore involved interpolation of rock mineralogies and unit boundaries through photo interpretation. However, our detailed analysis of the RATS data indicate the high quality and efficacy of the pre-mission mapping and traverse planning [2-3]. The compositional study of the limestone basement and differentiation between lava flows and cinder cones reflects that the pre-mission process of identifying traverse and sampling locations was successful. At the same time, there are regions that were not sampled or visited by the Desert RATS 2010 traverses that might have improved our maps and understanding of the geology of the area. For example, we lack data east of the limestone unit, in a deep N-S channel within the basalt. Since no basalt or surficial samples were collected, the contacts between these units have been inferred from imagery entirely. Samples of the older, underlying basalt would have also been helpful in determining the number of basalt flows as well as the contacts between them.

Processing the RATS 2010 data has also contributed to technical lessons learned that may be useful for future studies. The organization of the RATS database and recovering pertinent information after 3 years was difficult. Since none of the photos were geotagged, and some samples remain unlabeled, their context remains unclear. Similarly, some of the crew videos exclude information regarding whether certain samples were collected in situ or as float. These limitations to the Desert RATS data set create discrepancies between our map and the map constructed by the field team. In future work we will present full details of map uncertainties and inferences as well as the differences in our RATS and field map comparison.

**Acknowledgements:** We would like to thank Jim Skinner, Jake Bleacher, Debra Hurwitz, Barbara Janikko, and Peggy Whitson for collaborating in field mapping. Aid by ESRI specialist Kelly Boyd is also greatly appreciated. This project was supported by a Moon and Mars Analog Mission Activities Program grant. I would like to thank the Lunar and Planetary Institute for providing me with this opportunity.

Figure 2. A 1:24,000 map of the study area with limestone (ls), basalt flows (b1-h5), cinder cones (p1-p3), alluvium (al), and colluvium (cl) units labeled. The sample numbers, EVA and rover tracks have been superimposed.
ALTERATION OF LUNAR ROCK SURFACES THROUGH INTERACTION WITH THE SPACE ENVIRONMENT.  A. M. Frushour, S. K. Noble, R. Christoffersen, and L. P. Keller. 1Department of Geology, Appalachian State University, ASU Box 32067, Boone, NC 28608 (frushouram@appstate.edu) 2NASA Goddard Space Flight Center, Mail Code 691, Greenbelt, MD 20771 (sarah.k.noble@nasa.gov) 3Jacobs Technology, Mail Code JE23, P.O. Box 58447, Houston, TX 77058 (roy.christoffersen-1@nasa.gov) 4Astromaterials Research and Exploration Science, NASA Johnson Space Center, Mail Code KR, Houston, TX 77058 (lindsay.p.keller@nasa.gov)

Introduction: Space weathering occurs on all exposed surfaces of lunar rocks, as well as on the surfaces of smaller grains in the lunar regolith. Space weathering alters these exposed surfaces primarily through the action of solar wind ions and micrometeorite impact processes. On lunar rocks specifically, the alteration products produced by space weathering form surface coatings known as patina [1]. Patinas can have spectral reflectance properties different than the underlying rock. An understanding of patina composition and thickness is therefore important for interpreting remotely sensed data on celestial bodies without atmospheres. The purpose of this study is to try to understand the physical and chemical properties of patina by expanding the number of patinas known and characterized in the lunar rock sample collection.

Methods: We searched the Lunar Sample Compendium [2] to find rock samples which have been reported to have patina coatings. The search criteria included any mention of patina, micrometeorite craters, or “zap pits” in the Compendium sample descriptions. Available thin sections of these rock samples in the JSC Sample Curatorial Center were then studied with a petrographic microscope to survey samples for patina. Once good patina candidates were found, permission was obtained from the Lunar Sample Curator to carbon coat the thin sections. We then imaged the thin sections using a JEOL 7600F field-emission analytical scanning electron microscope (FE-STEM). The thickness, lateral extent, microstructure and chemical composition of the patinas were measured using the FE-STEM capabilities.

Results: Relative to previous studies [1,3,4,5] we increased by six (6) the number of lunar rock samples with characterized patinas. The studied patinas were in thin sections 15485,6; 12017,23; 14301,85; 10045,39 and 60025,163.

Thin section 15485,6 is a vitrophyric pigeonite basalt that has a discontinuous patina that is about 7 mm long and varies in thickness from about 10 µm to 350 µm. This patina is composed of multiple layers, each on the order of 1 to 10 µm thick (Fig. 1). Each layer is generally defined by rounded grains of glass <1 to 10 µm in diameter with a sub-layer of glass <1 to 5 µm thick separating each layer. In some areas the glass layers are more fragmental and in others more massive; layer boundaries become indistinguishable where this occurs. The individual layers are not flat, but rather undulatory. There are areas where the patina thickness is composed of just one layer and other areas that have 10s of layers. The glass grains and mineral fragments commonly fill in depressions in the outer surface of the rock. One localized region containing spherules of nanophase metallic Fe was identified. Based on energy-dispersive (EDS) compositional spectrum imaging, and spot analyses, the 15485,6 patina is fairly uniform in composition with subtle variations in Al and Mg (Fig. 2). 12017,23 is a pigeonite basalt that has a continuous layer of glassy patina that extends about 4.5 mm and varies from ~35 µm to ~600 µm thick. The patina is made mostly of silicate glass with entrained vesicles, mineral fragments, and spherules of nanophase Fe (Fig. 3). The patina contains vesicles on the scale of about 10 to 100 µm. There are two areas of nanophase Fe. One area consists of chain-like linear arrays of spherules <1 to 3 µm in size. The other area is in the form of multiple thin layers 1 to 5 µm thick. There are also partially melted rock fragments near the surface of the rock. The composition of the 12017,23 patina glass is overall fairly homogeneous, with the exception of local regions of partially or wholly melted minerals, such as ilmenite (Fig. 4). 12017,23 has an accompanying thin section 12017,22 on which patina also occurs but was not investigated in detail.

14301,85 is a regolith breccia that has a layer of patina approximately 2 mm long which ranges from 5 to 50 µm in thickness. The patina has distinct sub layers, one composed predominantly of mineral grain and lithic fragments, the other containing partially melted mineral grains and nanophase Fe (Fig. 5). The inner layer may or may not be patina, because it has a microstructural resemblance to types of clasts in the rock interior.

10045,39 is a low K ilmenite basalt which has a continuous patina that is about 400 µm long and varies from 1 µm to 130 µm thick. The patina is made mostly of a mix of highly vesicular glass and lithic fragments (Fig. 6). The glass is spread out along about half of the patina and contains schlieren-like layers with a high
concentration of nanophase Fe. The thickest part of the patina contains lithic fragments that are angular to sub-rounded with various degrees of melting. The patina has an area ~100 µm long composed predominantly of unconsolidated mineral fragments.

60025,163 is a ferroan anorthosite which has a continuous patina about 1 mm long that varies from 5 µm to 50 µm thick. The patina is composed of a single layer of silicate glass that is essentially identical in composition to the anorthositic host rock such that the boundary between the patina and rock surface is compositionally indistinguishable (Fig. 7). The glass does, however, contain some large nanophase FeO blebs in localized regions.

Discussion: Wentworth et al. [1] originally proposed a classification scheme for lunar rock patinas with three main categories. Fragmental patinas [1] are composed predominantly of aggregated rounded to sub-rounded particles, well sorted in the 1-5 µm size range and held together by various degrees of melt welding. Glazed patinas [1] are composed of continuous glass layers in which it is generally difficult to distinguish different small glassy constituents, i.e., the glass is microstructurally homogeneous. Classical patinas [1] are also dominantly glassy, but have complex microstructures containing multiple identifiable glass units, typically referred to as “splits” or “pancakes”.

We find that some, but not all, the patinas characterized in the current study fit well with the Wentworth classification [1]. TS 15485,6 has attributes fairly close to the fragmental type, particularly because the glassy spherules defining the layers are well sorted with respect to size. TS 60025,163 falls well into the “glazed” category given that it dominantly consists of continuous, homogeneous glass. TS 12017,23 fits somewhat into the glazed patina type, but it also contains regions with mineral fragments, making it a composite type not well covered by the Wentworth [1] scheme. Although TS 14301,85 clearly consists of fragmental material, the fragments are angular and poorly sorted, not consistent with Wentworth’s fragmental designation. TS 14301,85 is an example of a patina type we would call heterogeneous fragmental, as an addition to the Wentworth scheme. 10045,39 is another example of a very heterogeneous patina, with complex glassy regions, and mineral fragments, that does not fit existing types and should define a new “heterogeneous” type.

Conclusions: We were successful in finding more patinas, the yield of thin sections with patina being quite good considering the time invested. Our results considerably expand the overall microstructural and morphological diversity of lunar rock patinas. The variety of patinas found illustrate that they do not easily fall into the previous categories defined by Wentworth et al. [1].


Figure 1. Back-scattered electron (BSE) image showing fine-scale layering of patina in 15485, 6.

Figure 2. X-ray energy-dispersive (EDS) red-green-blue composite elemental map of 15485,6 patina showing the subtle variations in Mg (red), Ca (green), and Al (blue). Upper region of image is host rock.
Figure 3: BSE image of nanophase Fe\textsuperscript{0} (arrows) and a partially melted mineral grain in 12017,23.

Figure 4: EDS elemental map of Mg (red), Ti (green), and Fe (blue) showing a melted grain with ilmenite composition.

Figure 5: BSE image showing the fragmental outer layer, and what is possibly a glassy inner layer of patina on 14301,85.

Figure 6: BSE imaging showing example of glass, mineral and lithic fragments making up the patina on 10045,39.

Figure 7: BSE image of patina (arrow at right) on 60025,163 that has similar composition to the host rock (arrow at left).
HED PARENT BODY DIFFERENTIATION: C. Hoff\textsuperscript{1}, J. Jones\textsuperscript{2}, and L. Le\textsuperscript{3} \textsuperscript{1}University of Massachusetts, Amherst Department of Geology, \textsuperscript{2}NASA Johnson Space Center, Houston, TX, \textsuperscript{3}JETS, NASA Johnson Space Center, Mail Code JE-23, Building 31, Houston TX 77058

**Introduction:** The most well-known achondrites form what is known as the HED meteorite group which stands for howardites, eucrites, and diogenites. These meteorites include intrusive or extrusive igneous basaltic eucrites, intrusive orthopyroxenite diogenites, eucritic breccias, diogenitic breccias, and howardites, which are polymict breccias containing both eucritic and diogenitic clasts.

These meteorites are assumed to be related. The Frankfort and Yamato howardites display pyroxenes that resemble both eucrites and diogenites [1]. A physical mechanism that mixed both eucrites and diogenites is the likely explanation, implying that howardites, eucrites, and diogenites at some point came from the same parent body. The recent Dawn mission to Vesta has provided spectra of the lithology of the asteroid with different areas matching the spectra of howardites, eucrites, and diogenites [2]. The connection of HEDs to Vesta has allowed researchers to use the HED meteorites to postulate about the formation history of Vesta. Vesta is the largest asteroid to have a basaltic surface, making it the largest differentiated dwarf planet [3]. Vesta is also the only known asteroid inferred to have a metallic core, ultramafic mantle, and a basaltic crust [4].

Several different models exist that describe the formation of Vesta, but this abstract is going to deal with the last stages of the Righter-Drake magma ocean model [5]. The idea that a magma ocean could have been present on an asteroid is supported by magmatic iron meteorites that have very high liquidus temperatures and were once completely molten [6]. Temperatures high enough to completely melt such meteorites could have been high enough to form a magma ocean, or at least a substantial amount of melting on asteroids such as Vesta [6]. Vesta would have started off as a magma ocean that had a small metallic core that would have been surrounded by a mostly molten mantle [5]. Until a critical crystal mass fraction of ~0.80 could be attained, the mantle would have undergone convection and equilibrium crystallization following core formation. At some point, when the crystal-liquid mixture reaches $10^4$ Pa s, the system ceases the high activity of convection [6], which for the Righter-Drake model is ~80%. With this much crystallization, there would not be enough convection to prevent crystal settling. The slowing of convection would mark the formation of a dunite inner mantle and a diogenitic outer mantle.

The eucrites are generally divided into four categories: Main Group, Nuevo Laredo Group, Stannern Group, and Cumulate Eucrites. Main Group eucrites would have formed from residual liquids [from olivine and orthopyroxene crystallization] that were brought to the crust. The Stannern trend eucrites were either formed by low degrees of partial melting of a chondritic source or from contamination of Main Group eucritic magma by liquid formed from crustal partial melting [4]. The Nuevo Laredo and cumulate trend liquids however, would have been formed from crystal fractionation of a Main Group eucritic parent liquid [5].

This abstract describes experimental constraints on Main Group eucrites, which are intended to be compared to the crystal fractionation model of Righter and Drake [5]. They used MELTS to calculate a low-pressure fractional crystallization trend for a residual, 1250°C melt. At around 1250°C, ~0.80 crystallization would have occurred, mobilizing the residual liquid, and subsequent fractionation of that liquid would have produced a suite of diogenitic and eucritic materials [5]. This abstract will experimentally describe how this 1250°C residual liquid fractionated and will compare the results to the MELTS calculation. The purpose of the project was to evaluate the effectiveness of MELTS to model extraterrestrial magma systems by choosing a realistic situation in which MELTS was employed, as well as to further understand the origin of eucrites and diogenites.

**Experimental:** A starting composition was produced by quenching a synthetic liquid whose composition closely approximates that of the Righter-Drake 1250°C melt. Experiments were run in 1-bar, gas-mixing Deltech furnaces, initially for short duration (SD) runs (5-7 hrs), and then long duration (LD) runs (24-180 hrs). These experiments were
performed at 1250-1180°C, at oxygen fugacities near the iron-wüstite (IW) oxygen buffer, and a pressure of one bar. The lowest temperature, 1180°C, is Stolper’s original eucrite liquidus [8]. The temperature is controlled via a platinum-rhodium (Type B) thermocouple system. The oxygen fugacity is controlled through adjusted, flowing CO₂/CO gas and measured through a zirconia oxygen-fugacity electrochemical cell.

The charges were suspended on a rhenium loop to prevent iron loss, and the rhenium loop was in turn suspended from drop quench wires by thin platinum wire. The charges were then sealed in the furnace and set to equilibrate for the time allotted and then drop quenched into water. A typical experimental charge is shown in Figure 1.

The experiments were analyzed using a Cameca SX100 electron microprobe for oxide wt. % for the 9 elements in the starting composition (Si, Ti, Al, Cr, Fe, Mn, Mg, Ca, Na). For phase identification and visual orientation, a scanning electron microscope (JEOL SX100 electron microprobe for oxide wt. % for the 9 elements in the starting composition (Si, Ti, Al, Cr, Fe, Mn, Mg, Ca, Na). For phase identification and visual orientation, a scanning electron microscope (JEOL 5910LV) was used.

The initial experiments were three SD 1250°C, 1225°C, and 1200°C experiments. However, while the SD experiments gave a general sense of phase assemblages and K₀ values [K₀ = D(Fe)/D(Mg)], the large (>5%) zoning in the olivines and the large standard deviations for K₀’s indicated that longer experiments were required in order to reach equilibrium.

Results: The experimental set-up initially proved to be unsuccessful in providing adequately stable fO₂ (oxygen fugacity) and the redox state of the experimental charges was subject to random, rapid changes. Therefore, for the first three experiments (1250, 1225, 1200°C), the unstable fO₂ forced constant adjustments to the CO₂/CO gas mixture, necessitating SD experiments.

Table 1. Summary of Experimental Results

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>MgO %</th>
<th>CaO %</th>
<th>% Melt</th>
<th>Duration (hrs)</th>
<th>Phases Present</th>
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Table 2. Representative results comparing the 2013 and 1997 instances of MELTS results and experimental data collected

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>2013 Exp</th>
<th>1997 Exp</th>
<th>MgO %</th>
<th>CaO %</th>
<th>% Melt</th>
<th>Mg# Olivine</th>
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<tr>
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</table>

Near-constant fO₂ for long periods of time was obtained via hardware exchange and thus long-duration experiments were then enabled. The short-duration experiments proved to have not yet reached equilibrium. Although all experiments will be discussed in this abstract, because the LD experiments experienced stable fO₂, displayed little-to-no crystal zoning, and showed consistently different Mg#’s [molar Mg/(Mg+Fe) x 100] than the short duration experiments (Fig. 2), they are therefore given much more weight. Figure 2 shows that olivines in the SD experiments have nearly constant Mg#’s (~70), whereas the LD-experiment olivines’ Mg#’s decrease with decreasing temperature, as they should.

For our starting composition, we have three different predictions/results for how liquid composition and phase assemblage change with temperature. One of these predictions is from the current version of MELTS, another prediction is from a 1997 version of
Discussion: MELTS has long been used as thermodynamic modeling software for both terrestrial and extraterrestrial geologic environments. It seems however, that MELTS does not agree with our experimental measurements or with different versions of itself (Table 2).

Key differences between the three different results include the presence of phases as a function of temperature and the resulting and/or independent oxide weight percentage differences (Table 2). The 2013 version of MELTS predicts the first appearance of pyroxene at 1190°C and the 1997 MELTS version predicts it to appear at 1210°C: ~20°C higher. None of our experimental charges indicated any pyroxene except for one 1225°C-SD experiment (not reported in Table 2) where there may have been some issues with fO$_2$ and the drop-quench process. This experiment was redone as a 1225-LD experiment and only olivine was present. This experiment will be repeated again to see if pyroxene forms.

The MgO oxide weight percents and mafic phase Mg#'s as functions of temperature are reported side by side with the other two instances of MELTS, and none of them completely agree (Table 2). MELTS consistently picks higher Mg #'s than the experimental results showed.

Our K$_D$ partitioning results are broadly consistent with Jones [9] and Filiberto et al. [10] who demonstrate a linear relationship between D(Fe) and D(Mg) in an olivine-melt equilibrium system.

The lack of pyroxene in our experiments has implications for the formation of eucrites and diogenites. Although we were able to produce a composition similar to that of a typical eucrite (Fig. 3) [11], its liquidus temperature is higher than that found by Stolper [8]. The fact that we did not crystallize pyroxene implies that the 1250°C composition of Righter and Drake is not capable of producing both eucrites and diogenites. Therefore, the link between actual eucrites and actual diogenites remains to be established.

Typical eucrite was calculated from Sioux County and Juvinas eucrites [10]


ANALYZING RIM CREST VARIATIONS IN LUNAR IMPACT CRATERS. E. F. Lalor¹ and V. L. Sharpton². ¹Dept. of Earth and Environmental Sciences, Temple University, Philadelphia, PA 19122 (eve.fattah@gmail.com), ²Lunar and Planetary Institute, Houston, TX 77058 (sharpton@lpi.usra.edu).

Introduction: Cratering is the most common geologic process in the solar system. Understanding what controls crater shape [1-4] is therefore essential to using craters as exploration tools [5]. Though the impact process itself is generic, the variations in a crater are controlled by physical factors unique to the planetary body. On an airless body such as the moon, crater shape is controlled primarily by characteristics of the projectile (velocity, angle, mass) and the nature of the target (strength and gravity). Consequently, morphologic variations in lunar craters give insight into lunar geology and the impact process.

Understanding these variations is critical for characterizing crater formation. Pike [2] compiled morphometric measurements including depth, diameter, and rim height of 484 lunar craters based on shadow measurements and topographic data from Apollo-era Lunar Topographic Orthophoto (LTO) maps. This dataset has provided the basis for widely used scaling relationships linking crater depth and crest height to diameter. These scaling relationships are crucial to our current understanding of crater formation and for reliably using crater shape to constrain planetary stratigraphy (e.g., [5]). Here, we rely on high-resolution imagery and topography from the Lunar Reconnaissance Orbiter (LRO) to:

1. Document the full extent of rim height variations observed in morphologically fresh lunar craters;
2. Investigate the geologic controls over the observed rim crest variations; and

Technical Approach: The topographic dataset we used is the GLD100, a stereo-derived digital elevation model (DEM) from the LRO Wide Angle Camera (WAC) imagery with a horizontal resolution of 100 m and vertical resolution of ±20 m [6]. Comparatively, the LTO maps used by Pike have a resolution of ±50 m vertical, ±200 m horizontal, and a contour interval of 100 m [4]. The topographic data and WAC imagery were loaded into ESRI’s ArcMap for image processing and crater analysis.

Thirty morphologically fresh craters on mare plains were selected, ranging 9-42 km in diameter. Superposition on mare plains ensures ease of calculating the pre-impact surface, as well as a selection of relatively fresh craters that are not degraded by other impacts or flooded with lava such that pristine topography is lost. The size range includes simple and complex (i.e., central peak) craters, as well as craters within the simple-to-complex transitional zone. Craters smaller than ~5 km strain the resolution limits of GLD100 data, and larger craters are less common on the lunar mare and tend to be more heavily degraded.

Constraining the pre-impact surface. To extract the relative topography of the crater facies, the first step is to determine the pre-impact surface. For each crater, we extracted a subset of the GLD100 extending approximately five crater diameters from the crater center in order to show elevations referenced locally rather than to the entire moon. The elevation points of the

Figure 1: Comparison of our measurements with Pike’s values. Data points in black are those from shadow data rather than LTO images. Crater names are displayed with average diameter in kilometers.
crater and its ejecta were then removed from the local DEM. A third order polynomial trend surface was calculated from the remaining data points, and then subtracted from the local DEM to produce a final DEM referenced to the pre-impact surface of each crater.

**Defining Rim Crest Topography.** The rim crest of each crater was traced and resampled to have vertices every 100 meters. The evenly-spaced vertices contain X, Y, and Z coordinates of a polyline around the circumference of each crater’s rim crest. This polyline’s mean center of figure defines the crater center, irrespective of the crater’s degree of circularity. The coordinates of the center and those of each rim vertex were used to calculate the radius at each vertex and its azimuth from the crater center. This azimuthal data is shown as an elevation profile extending clockwise around the rim crest of the crater.

**Discussion:** For most craters, the average rim crest elevations from our study and the single values reported by Pike [2] agree within 200 meters or less. We infer that resolution limits rather than methodical errors are the source of discrepancies in Pike’s data. Overall, the measurements from LTO are more conformable than those from shadow measurements. The shadow measurements are not systematically inaccurate, but the points with the most disagreement from our dataset tend to be from shadow data. Pike measured the pre-impact surface using the mean elevation of the terrain beyond the rim flank. Some discrepancies between our dimensions may be due to the use of different reference elevations, especially for craters superimposed on irregular topography. Measurements of radius disagree by more than 500-1000 km in many cases. Most craters with this radial discrepancy have low degrees of circularity.

Pike’s [2] data set was the basis for his depth-diameter (d/D) analysis in [3]. The d/D ratio is a function of increasing crater complexity. Simple craters are small and bowl-shaped with a d/D of roughly 1/5 [3], and complex craters are wider with flat, shallow floors and increasingly smaller d/D fractions. Pike [3] states that the simple to complex transition occurs at 16 km for lunar mare craters. In our analysis, no craters between 15 and 20 km could be confidently determined as either simple or complex, indicating that the transition is a spectrum rather than a binary.

Crest height and radius variations constrain the transition through the degree of collapse that has occurred on the crater wall. Simple craters are small enough to be stable in their round and regular form. The first transitional features observed as diameter increase are singular slump blocks that indicate local structural weaknesses. The degree of slumping escalates across the transitional zone. Complex craters are characterized by concentric terraces indicating that the entire wall has collapsed. Consequently, complex craters show the highest degree of rim crest height variability. Transitional craters show the greatest radial width variation due to their asymmetric collapse.

![Figure 2: Simple Craters](image) Rosse crater (size) has a uniform rim crest with less than 200 m variance. Mosting A (size) has a variable rim crest reflecting the pre-impact topography. Messier (size-size) shows a predictable pattern of rim crest variation caused by low-angle impact.
Simple craters that deviate from the pattern are most strongly influenced by target heterogeneities or impact obliquity. The successive basalt flows of the mare are denser and more resistant than fragmented highland material. The crater Mosting A (Fig. 2) is superposed on an interfingering of mare and highland lithologies. The less dense highland rock was preferentially excavated during impact, building an irregular rim crest. This effect is also seen on the larger Melenlaus crater, which lies directly on the mare-highlands boundary in southern Mare Serenitatis.

Messier crater (Fig. 2) is an exceptional example of an extremely low-angle impact trajectory. The uplift and ejecta are concentrated to the elongated crossrange walls, and are negligible in the uprange and downrange directions. An oblique impact crater will show a strong negative relationship between radius and rim height. Greaves crater is another more subtle expression of this phenomenon.

Figure 3: Transitional and complex craters. A. Carrel crater exhibits a singular slump block. B. The walls of Flamsteed have collapsed to a greater extent. C. Burg crater exhibits the characteristic scalloped rim of complex craters. D. Picard has maintained its roundness despite full rim collapse.

The first group of transitional craters has low height variability but high radial width variation. These craters have low degrees of circularity. In Carrel crater (Fig. 3A), the radius varies almost 30% from the average due a massive slump in the eastern quadrant of the crater. The localized slumping in these craters appears to indicate otherwise undetected local structural weaknesses in the bedrock. The other set of transitional craters has high crest height variation and low radial variability. Flamsteed (Fig. 3B) exhibits broader sections of enhanced wall collapse than Carrel; consequently, this slumping contributes more to lowering the rim crest than to radial extension.

Formation of terraces within complex craters often coincides with a crenulated rim from multiple slumping events, as in Burg crater (Fig. 3C). These scalloped walls generate a wide range of crest height variations. However, radial variation in complex craters levels off around 20% of average radius. Picard (Fig. 3D) is anomalous as it is much smaller than other complex craters, and has concentric terraces yet maintains circularity and crest height regularity.

Conclusions:

1. Modern remote sensing data and ArcGIS analysis provide more precise measurements of the range of rim crest heights within lunar craters.
2. Pike’s [2] measurements reasonably agree with our average height values, though his shadow measurements and crater radii show some significant deviations.
3. The transition from simple to complex crater morphology occurs over the 15-20 km range on the lunar mare surface in a stepwise addition of complex characteristics. As crater size and complexity increase, so do degrees of rim height and radius variation. Transitional craters show significant variability in either crest height or radius, both within individual craters and between craters of similar sizes. Complex craters have great rim crest disparities among themselves, but similar degrees of radial variation.

GEOLOGY OF THE ORIENTALE BASIN AND ITS IMPACT MELT SHEET. D. J. P. Martin, School of Earth, Atmospheric and Environmental Sciences, University of Manchester, Oxford Road, Manchester, United Kingdom, M13 9PL; dayl.martin@student.manchester.ac.uk (Advisor: P.D. Spudis).

Introduction: Orientale is the youngest and best preserved multi-ring impact basin on the Moon [1]. Situated on the extreme western edge of the lunar near side, observation of the basin from Earth is limited and only possible at a highly oblique angle. Therefore studying the basin effectively has been restricted to the analysis of space data, such as images and compositional maps obtained from spacecraft. Recent missions such as the Lunar Reconnaissance Orbiter have provided much higher resolution images of the surface than previous missions, allowing detailed study of the surface of the moon and its features.

The most recent geological map of the Orientale Basin was made in 1978 using data from the Lunar Orbiter and Zond 8 missions [2]. The aim of this project was to create a new geological map of the Orientale Impact Basin using images and data from the Lunar Reconnaissance Orbiter Camera (LROC), the Moon Mineralogy Mapper (M3), Clementine FeO and TiO2 maps and LOLA topographic maps and to determine the extent of differentiation of the impact melt sheet (if any). The updated map shows a more accurate representation of the distribution of units both inside and outside of the basin and provides a clearer insight into the distribution of the melt sheet, the geological context of impact melt inside the basin and nature of the basin-forming impact. Determining if the Orientale basin melt sheet has differentiated can help determine the behavior of large igneous bodies on the Moon and how this behavior may differ from impact melt bodies on Earth.

Method: A WAC image mosaic of the western hemisphere of the Moon (formed from images taken by the LROC) was centered on Orientale and used as a base-map in ArcGIS 10.0. Clementine FeO and TiO2 maps [3] and the GLD100 global topographic map derived from LOLA and WAC stereo images [4] were also used to aid in the identification and mapping of various units (for example finding areas of mare material with high FeO content in the texturally similar Maunder Plains). Separate units were defined using a number of characteristics such as position within the basin, surface texture, composition, structure and stratigraphic position. The unit names of the Orientale Group [5] and similar colors of the map of Scott et al. [2] were used in the updated map for consistency with existing lunar maps but some of the formations were subdivided into a number of constituent members based on one or more of their characteristics varying throughout the formation.

Following the creation of the geological map, an analysis of the melt sheet (the Maunder Fm.) could begin. Using ArcGIS, the FeO and TiO2 contents of the different formations could be analyzed by overlaying the different formations onto the Clementine maps and performing a statistical analysis of the element concentrations and distributions within the areas of the separate formations. This could be used to give an overall average composition of the formations, detailed ‘spot’ compositions and therefore a range of compositions over a given area. Using this technique, compositions of the ejecta blankets of over 300 craters within the Maunder Formation were analyzed. To test for compositional variation with depth, I also analyzed crater size as a function of FeO content (based on the idea that larger craters excavate deeper material) [1,6].

<table>
<thead>
<tr>
<th>Formation</th>
<th>Wt% FeO</th>
<th>Wt% TiO2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mare Orientale</td>
<td>11.04</td>
<td>2.313</td>
</tr>
<tr>
<td>Maunder Plains</td>
<td>4.55</td>
<td>0.589</td>
</tr>
<tr>
<td>Maunder Rough</td>
<td>4.39</td>
<td>0.575</td>
</tr>
<tr>
<td>Montes Rook Knobby</td>
<td>4.57</td>
<td>0.496</td>
</tr>
<tr>
<td>Montes Rook Flat</td>
<td>5.05</td>
<td>0.510</td>
</tr>
<tr>
<td>Massif</td>
<td>4.10</td>
<td>0.458</td>
</tr>
</tbody>
</table>

Fig. 1 – A table of average iron and titanium compositions of the different formations. This was created using the Clementine FeO and TiO2 maps.
Results: The formations described by McCauley [5] have been largely unchanged except that some have been split into a number of constituent members. The Mare Orientale displays a number of small melt ponds previously unmapped (partly due to comparison with the Clementine FeO map [3]). The Maunder Formation has been split into 2 units – a smooth (plains) member and a rough member. Maunder rough areas have a relatively large range of surface topography. Maunder smooth member is flat and, in some areas, extensively fractured.

The Montes Rook Formation has also been split into two members – knobby and plains-like. The knobby member appears to contain large hummocks of material giving these areas a blocky and uneven appearance. However, flow lobes have been observed against some of the massifs or against the Cordillera Ring in some areas. The plains are mainly concentrated in the southwestern quadrant of the basin interior.

The Hevelius Formation (basin exterior ejecta) has been split into three members: smooth, highland plains similar to the Imbrium basin Cayley Fm., radially textured ejecta (deposits linedated radial to the basin center) and transverse ejecta (textured material oriented parallel or concentric with the basin rim). The radial ejecta is mostly situated to the north and south of the basin rim with transverse ejecta being found mostly to the east and west. Plains are present in large areas outside the ejecta blanket and in small, localized areas within the ejecta blanket. Secondary craters are abundant to the northwest, southeast and southwest; relatively few are found to the east and northeast.

The Orientale basin displays at least four distinct concentric rings [1,6]. The innermost ring (320 km dia.) is expressed as a simple scarp within the Maunder Fm., with up to 5 km of relief between the ring and the flat, mare floor of the basin. The next larger ring (480 km dia.) is expressed by the Inner Rook Mts., an irregular but circular arrangement of equant massifs. Some of these massifs are composed of nearly pure anorthosite [7]. The next larger ring is the Outer Rook Mts. (620 km dia.), again made up of massive, blocky mountains roughly arranged in a circular pattern. The outer ring (and main topographic rim of the basin) is the Cordillera Mts. (930 km dia.), which in appearance seems to be a simple scarp. In most cases, the rings mark the limit of exposure of units of the inner Orientale Group (e.g. the Montes Rook Fm. is largely confined between the Cordillera and Outer Rook rings) but exceptions occur locally.

Analysis: I interpret the Maunder rough member as thinner areas of the melt sheet that have been draped over lower-lying areas of massif material. Maunder plains member is thought to be flat due to the melt sheet ponding and being locally thicker. Topographic analysis of the Inner Rook ring can be used to estimate a thickness of the melt sheet (as there are areas where the ring is completely covered by melt). The largest range is 6.2 km suggesting the melt sheet is ~6 km or less thick.

The Montes Rook Formation may contain some portion of the impact melt due to the presence of flow lobes near to Cordillera Ring and extensive areas of the plains in certain locations.

The “bilateral symmetry” of ejecta distribution of the Hevelius Formation appears similar in some ways to the “butterfly pattern” of ejecta formed from low-angle impacts [e.g., 1]. The concentration of plains in certain areas, the distribution of secondary craters, and the distribution of ejecta facies all support the idea that the Orientale basin formed by an oblique impact from the east-northeast [1].

More than 300 craters within the Maunder Fm. were analyzed to determine the variation in composition with depth of the melt sheet (Fig. 2).

Fig. 2 – A chart showing FeO composition with increasing crater diameter. Dark line represents mean surface FeO content; shaded area shows data within 1 standard deviation of mean.
There is no correlation between crater size (and hence, depth) so the composition of the melt sheet appears to be uniform. The largest craters sampled were approximately 15 km across and excavated material from ~1.5 km depth within the melt sheet [8]. Modeling of a basin-sized melt sheet suggests a stratigraphy of noritic or anorthositic-noritic layer overlying lower troctolitic or pyroxenitic rocks [9]. Although the upper surface is anorthositic norite, there is no evidence for the presence of more mafic compositions at depth (Fig. 1). Given the estimated initial thickness of the melt sheet (as much as several km; [9]), at least some level of differentiation would have occurred by analogy with the largest impact melt sheets on Earth (e.g., Sudbury igneous complex; [10]). Some process on the Moon restricts or eliminates fractional crystallization of the melt sheet.

Conclusions: An updated map of the Orientale basin and its surrounding ejecta shows the relations of basin units. The distribution of ejecta suggest that the basin formed by an oblique, low angle impact coming from east to west. The impact melt sheet (Maunder Fm.) has a maximum thickness of a few km at most. Using this map along with compositional melt maps, I determined that the basin impact melt sheet is largely undifferentiated.


Fig. 3 – An updated geological map of the interior units of the Orientale impact basin. 
*Width of image is ~1100 km*
The correlation of Martian crater ejecta morphology with geologic units and radar surface permittivity in the northern plains. M. Petitjean; S.M. Clifford; 

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The existence of a global ice-rich permafrost on Mars is supported by the identification of a variety of cold climatic landforms, the interpretation of various remote sensing data sets (especially those of the MARSIS and SHARAD orbital radar sounders and the Gamma Ray Neutron Spectrometer), and direct investigations of the Martian surface by landed and roving spacecraft. But the origin, quantity and distribution of subsurface ice are still unclear. There are in situ drilling, geophysical, and compositional investigations which are capable of resolving these questions, but they are both expensive and technologically challenging; so geomorphic observations remain the most expedient and cost effective ways of investigating the occurrence of ice in the Martian subsurface.

The objective of our summer project was to investigate the possible correlation of MARSIS-derived radar surface permittivities with Martian crater ejecta morphology – especially those craters with fluidized ejecta - which are thought to result from an impact into ice. We are particularly interested in the potential correlations which may exist in the northern plains, where such evidence may be indicative of the presence of a frozen relic of a northern ocean or outflow channel discharge.

Based, in part, on the tentative identification of potential nested paleoshorelines along the dichotomy boundary[1], [2], have proposed that Mars may have once hosted a large ocean or ice sheet in the northern plains, covering as much as a third of the planet. Hydrologic arguments suggest that such an ocean or ice sheet was almost certainly an initial condition – condensing shortly after the planet formed and persisting until the end of the Noachian (~3.7 Ga). With the loss of the planet’s early atmosphere by atmospheric erosion, exospheric escape, and the formation of hydrated minerals and the long-term decline in the planet’s geothermal heat flow, any early ocean is expected to have rapidly frozen. With the transition to global climatic conditions similar to those of Mars today, early ocean ice would have been thermodynamically unstable – leading to its sublimation and cold-trapping at higher latitudes. However, some of it may also have been buried by volcanics and eolian and fluvial sediments, which could have potentially preserved the remaining ice over Martian geologic time.

Major episodes of outflow channel activity during the middle of Martian geologic history may have resulted in the occurrence of transient lakes and seas which, like an early ocean, would have rapidly frozen and been subject to both sublimative loss and potential burial by sediments and volcanics.

Fig.1: This figure represents the distribution of the rampart craters, according to Robbins database, compared to the permittivity values. The Polar regions, ice-rich are much more covered, and present higher mobility than the equatorial zone, in spite of the low-permittivity values.

A preliminary study (Clifford et al., EGU 2013) was done based on the global crater database of Costard [3] which was compiled from Viking Orbiter imagery, with a typical spatial resolution of the ~50 m/pixel range. It allowed making the difference between low permittivity values in the equatorial area, and those in the polar region. This can be explained by the different possible origins in these low permittivity values: a high porous ground or the important presence of ice in the sediment. It demonstrates a good correlation between high ejecta mobility and low permittivity data.

Here, we also try to resolve this ambiguity by observation of the geomorphology (fig.1).
One of the main requirements for our study was to find the most complete database of Martian crater characteristics, including: location, morphometry, and ejecta morphology. Barlow et al. [4] present the consensus definitions and procedures adopted by the Consortium. The U. S. Geological Survey in 1999, create to facilitate the standardization of terminology and methodology in defining crater size, morphology, and ejecta mobility. This consensus was particularly welcome in standardizing the categorization of fluidized ejecta craters, which have previously been referred to as rampart, ‘splosh’ or ‘flower’. For our study, we chose the Robbins database, because it is the most recent, precise and complete. Also, the database used informed us about the mobility of these craters; a high mobility is typically the evidence of the presence of volatiles in the ground.

However this information was recorded for only a small fraction of the craters in the Robbins database. For this reason, we decided to make use of a relation by Melosh [5] for the radius of an ejecta sheet as a function of the ejecta by the use of two methods. When the perimeter of ejecta is indicated, we obtain the mean radius by referring of the relation in between. When this perimeter is not indicated, we used the size of the crater to approximate the size of the ejecta:

\[ R_{\text{ejecta}} = (2.3^{1/0.5}) \times R_{\text{crater}}^{1.006} \]

Where \( R_{\text{crater}} \) is the radius of the crater, and \( R_{\text{ejecta}} \) the mean radius of the ejecta.

With such a method we obtain an estimated diameter of ejecta, according only to the crater size, and not the ice content of the ground. We obtain here a final database, mostly based on the Robbins’s one, with 384343 craters referred, a mathematic approximation of their mobility, and divided in five types: RD (no ejecta visible), Slers, Dlers, and Mlers. (the rampart craters), and the Pancake craters (ejecta present but no lobated).

Another question was the definition of the northern plains. A first attempt has been made with the topographic dichotomy border, but the different geological nature so included near the equatorial zone was not the best choice for the significance of the final data. So we decided to refer to the age of the geological unit: all geological units of the north hemisphere post Noachian have been selected for our studies.

The distribution of the craters has been plotted with their mobility characteristics in order to obtain a map showing evidence of overlapping crater (fig 3). Our idea was to use this map as a tool of repair to find areas of interest presenting overlapping craters of different types. The study of the distribution of these different relationships of overlap in the geological units was described in order the find a logical distribution of these overlaps, showing an evidence of the history of volatiles in the ground throw time. We can precise that because part of the mobility information were estimated by a calculation, we don’t have a reflection of the quantitative ice-content of the ground, but the map shows strongly enough the area of overlapping. With this tool, we were able to make a schematic draw of the different relationship of overlapped ejecta craters. Because of the complex interface provided for the HRSC dataset, we decided to use the Google Mars software, with the CTX imagery. Our method consisted in a systematic scanning of the lowland field and notes the most interesting overlap relationship. We were particularly interested in the small rampart craters overlapping, because of their evidence of near surface volatiles. A two kilometer diameter rampart crater is evidence of few hundred of meters deep volatiles. Also, we agreed on criteria about the degradation state of the...
geological figures found. Figure 2 presents a better idea of these criteria.

Fig.3: Zoom to a typical region of the map generated.

Comparing the different observations made as described and the age of the unit where they are located, we decide to accentuate the comparison in between the large area of the Vasistas Borealis Formation, as the older unit of the Northern plain, and the Amazonian formation as the most recent one.

It appears that in any case, the evidence of the depletion of volatiles were found in all geological units; a rampart crater overlapped by RD type with a large enough diameter is evident, as soon as the onset diameter of the RD morphology was large enough to be compared to the main element. But this observation has been made in very different ages of unit (Hesperian and Amazoniain); which doesn’t allow making any hypotheses about the timing of this evolution.

References:

PRELIMINARY STRAIN MEASUREMENTS OF THE CHONDRULES AND REFRACTORY INCLUSIONS IN THE ALLENDE CV3 METEORITE.  A. W. Tait1, K. R. Fisher2. 1Monash University, Clayton, 3168, Vic., Australia, (alastair.tait@monash.edu). 2University of Cincinnati, Cincinnati, OH, 45219 (kenton.r.fisher@NASA.gov).

Introduction: One of the problems in meteoritics today lies in understanding chondrule (and Calcium Aluminum Inclusion (CAI)) formation, and how that relates to planetesimal accretion. Specifically, why do individual chondrites (i.e., parent bodies) appear to exhibit extensive sorting of chondrules, but chondrule size populations between parent bodies remains heterogeneous [1]? Furthermore, is this sorting mechanism also responsible for counteracting the gas drag on CAIs allowing them to mix with the later formed chondrules [2]? Understanding such mechanisms would help clarify the timing of events in the early solar system, and by extension planetary formation.

Many hypotheses have been proposed to explain apparent sorting in chondrites, such as: X-winds [3], photophoresis [4], turbulent concentrations [5], and mass sorting [6]. Each of these mechanisms make predictions as to the shape and size of the chondrule distribution [2]. Therefore an accurate data set is needed to distinguish such models. However, the task of analyzing particle populations in chondrites (e.g., chondrules and CAIs) has its biases. Firstly, previous counting of particles has been done on thin sections which, due to their relatively small size, do not represent the size population seen in hand samples. Secondly, by cutting samples into slabs, the number of smaller chondrules is artificially inflated due to the particles being cut non-diametrically. Fortunately, statistical corrections can be employed to correct for this [6]. Currently these corrections only work if the following two assumptions are accurate: a) that the chondrules or CAIs are spherical to begin with, and b) that the sample has not been exposed to deformation. It is generally accepted that chondrules are initially spherical [7,8]. The history and extent of deformation among many meteorites are still largely unknown, but thought to be driven by impacts [9]. Less common features attributed to impacts are the presence of fabrics and lineations in meteorites that were first observed in 1960’s [10]. Yet, fifty years later only a few studies have reported that meteorites have recorded such features [7]. Although impacts are often cited as a mechanism for this deformation, plastic deformation from overburden and nebular imbrication have also been reported [10-12]. Previous work conducted on the Leoville CV3, and the Parnallee LL3, exhibited a minimum uniaxial shortening of 33% and 21% respectively [11,12]. The Allende has been looked at before; previous workers using Electron Back Scatter Diffraction (EBSD) found a minor-axis alignment of olivine grains inside dark inclusions and an “augen” like preferred orientation of olivine grains around more competent chondrules [13,14]. This study will expand on this work by using traditional strain measurements techniques, combined with X-ray computerized tomography (CT) imagery, to independently evaluate evidence for preferred orientation and to address the strain in three dimensions.

By better understanding the deformational history of chondrules, future modifications can be made to the existing chondrule size distribution correction algorithms [15] to get a more accurate picture of the true the distribution of particles in meteorites and, by extension, what mechanisms are driving particle sorting in the early solar system.

Methodology and Results: For this study a ~25 cm slab of the Allende CV3 meteorite was chosen due to its unique size. Such a large sample size allows for observations on the near “outcrop” scale; this allows for more accurate size distributions (in particular for the CAIs) than previous “thin-section” studies that may introduce biases due to the small sample size.

Data Preparation. The slab was previously photographed and stitched together at 13.88 µm/px resolution across the whole slab on both sides, with the CAIs digitized in Adobe Illustrator [16]. The slab was also subjected to X-Ray tomography with a resolution of 173.91 µm/voxel. All this data was then aligned, stacked, flipped (where necessary) and filtered at a confidence level of > 80,000 µm² for CAIs. Due to the size of the sample, comparative statistics were run on the CAIs in nine equal area selections in order to choose a smaller, representative region to further digitize for chondrules; this became our medium selection region. To minimize ambiguity, two steps were taken. First was to lump the specific phase components into more general phases. The first phase was all chondrule types (POP, PP, BO etc.). The second phase was all CAI types and Amoebic Olivine Aggregates (AOA). The last phase was chondrule rims; these were quite subjective due to their variety, alteration and similarity to granular chondrules. The second control was to have multiple workers digitize
their own data sets from the original images. Once the data was digitized, the images were processed in ImageJ [17] to record their physical characteristics: coordinates, area, major axis, minor axis, angle, circularity and solidity.

**Finite Strain Analysis.** Much of the strain analysis in this study borrows heavily from the systematic methods laid out by Cain et al. (1986) that have been used in subsequent studies of meteorite fabric analysis [7,12]. Using the medium selection, digitized chondrules, CAIs and chondrule rims were processed in ImageJ with their angle and Major/Minor axis recorded. Due to the risk of data duplication, only the topside of the slab was reported. Harmonic means were used to reduce the effect of outliers in the chondrules and CAIs. Chondrule rims were measured manually by approximately measuring the long axis of the chondrule and rim in line with the fabric (see Fry’s method). The rim axial ratio was normalized to that of the parent chondrule. See Table 1 for results.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Top (Tait) (Axial Ratio)</th>
<th>Top (Fisher) (Axial Ratio)</th>
<th>±σ</th>
<th>Minimum Mean (Axial Ratio)</th>
<th>Uniaxial Shortening (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chondrule</td>
<td>1.32</td>
<td>1.30</td>
<td>0.32</td>
<td>1.31</td>
<td>16</td>
</tr>
<tr>
<td>CAI</td>
<td>1.70</td>
<td>1.66</td>
<td>0.64</td>
<td>1.68</td>
<td>29</td>
</tr>
<tr>
<td>Rim</td>
<td>1.50</td>
<td>1.50</td>
<td>0.75</td>
<td>1.50</td>
<td>24</td>
</tr>
<tr>
<td>Matrix</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>1.40</td>
<td>20</td>
</tr>
</tbody>
</table>

**Fry’s Method.** This is a point-to-point technique for determining whole rock deformation in anti-clustered rocks such as sandstones and conglomerates [18]. Undifferentiated and initially undeformed chondrites are perfect examples for this method due to their random distribution of particles. This method was used to determine the whole rock strain and the relative strike of the fabric. Chondrules were selected over CAIs for this method due to their known original aspect ratio. Chondrules from the medium region were run through ImageJ and an objective ellipse fitting plugin [19]. The results showed a long axis angle of 038° and an ellipse ratio of 1.4 (Fig. 1).

**Flinn’s Method.** This is a technique that measures the strain of a sample in 3D by graphing the strain axis of an ellipsoid X/Y and Y/Z [20]. The descriptor term ‘k’ describes the type of deformation undergone; prolate for k > 1, and oblate for k < 1. If a population of samples is plotted in both the oblate and prolate field, the rock is likely to have been subjected to simple shear. All other strain analysis gives minimum values for strain; the inclusion of CT data allows for the ability to measure fully encapsulated grains which is closer to the absolute value for the deformation history. This was done by digitizing CAIs (n=19) from a stack of CT imagery generated from the main slab. An ellipsoid was fit to the particles using BoneJ [21]. The results showed axial ratios of X/Y = 1.42, Y/Z = 3.69, X/Z = 5.26. This translates to a uniaxial shortening of 49%. The average was k = 0.27 which puts it in the oblate field (Fig. 2).

**Rose Diagrams.** These plot the angle of the Major axis (top of the image = 0°) on a half circle of 180° in buckets of 10°. All phases record maximum values between 31-70° but due to the larger data set and the higher aspect ratio, CAIs recorded a maximum value between 31-40° which is within the range of the Fry Plot results.

**Dip Calculation.** CT imagery was used to determine CAIs (n=6) that were visible on both sides of the slab. The outline of each CAI defined in each slice was then digitized and its coordinates recorded. These offsets were recorded in excel along with the known z offset of 173.91 µm/image to determine the
dip of each CAI by trigonometry. The average dip was: 17.56° ± 4.81°.

**SEM Element Map.** Element maps were previously generated on a JEOL 7600F Scanning Electron Microscope (SEM) with a pixel resolution of 3.23 µm/pixel [15]. The elements mapped were Ca (Green), Al (Blue) and Mg (Red). Images show an apparent long axis alignment of Ca-rich pyroxenes (~50 µm) surrounding particles. These pyroxenes appear to be from the matrix and not part of the chondrule rims (Fig. 3).

![Fig. 3](Image)

**Discussion:** **Strain Description.** Both the Rose diagram and the Fry diagrams support a fabric with a strike of ~038° (relative to the north of the image) and a dip to the “west” of ~17°, supported by trigonometry of the CAIs. It would appear not all phases have partitioned the strain evenly; CAIs record the most strain with a minimum uniaxial shortening of 29% and a maximum shortening of 49%. The maximum value is based on the assumption that CAIs were initially perfect spheres like the chondrules. This is unlikely given the different CAI types, fragments and rheology. Nevertheless, the relative order of apparent strain partitioning is: CAIs > Rims > Matrix > Chondrules. Chondrules are the least strained (16%) and CAIs with chondrules recording the least strain (16% uniaxial shortening) and CAIs the most (29-49%). Deformation in rims places their formation times before deformation. Matrix deformation confirms previous work [14] and casts doubt on the imbrication hypothesis [10].

What are space exposure histories telling us about CM carbonaceous chondrites? A. Takenouchi and Michael. E. Zolensky, 1 The University of Tokyo (7-3-1 Hongo Bunkyo-ku Tokyo Japan, a.takenouchi@eps.s.u-tokyo.ac.jp), 2 NASA Johnson Space Center (2101 NASA Pkwy, Houston, TX 77058, michael.e.zolensky@nasa.gov).

Introduction: Chondrites are chemically primitive and carbonaceous chondrites are potentially the most primitive among them because they mostly escaped thermal metamorphism that affected the other chondrite groups and ratios of their major, non-volatile and most of the volatile elements are similar to those of the Sun. Therefore carbonaceous chondrites are expected to retain a good record of the origin and early history of the solar system.

Carbonaceous (C) chondrites are chemically differentiated from other chondrites by their high Mg/Fe ratios and refractory elements and have experienced various degree of aqueous alteration. They are subdivided into eight subgroups (CI, CM, CO, CV, CK, CR, CB and CH) based on major element ratios and oxygen isotopic ratios. Their elemental ratios spread over a wide range though those of ordinary and enstatite chondrites are relatively uniform. It would be very useful to know how many separate bodies are represented by the C chondrites.

In this study, CM chondrites, the most abundant carbonaceous chondrites, are examined. They are water-rich, chondrule- and CAI-bearing meteorites and most of them are breccias. High-temperature components such as chondrules, isolated olivine and CAIs in CMs are frequently altered and some of them are replaced by clay minerals and surrounded by sulfides whose Fe was derived from mafic silicates.[…]. On the basis of degrees of aqueous alteration, CMs have been classified into subtypes from 1 to 2, although Rubin et al. (2007) assigned subtype 1 to subtype 2 and subtype 2 to subtype 2.6 using various petrologic properties [1]. Their classification is useful to distinguish CMs based on petrographic and mineralogic properties. For example, though tochilinite (2[(Fe, Mg, Cu, Ni)[S] 1.57-1.85 [(Mg, Fe, Ni, Al, Ca)(HH)])) clumps are produced during aqueous alteration, they disappear and sulfide appears with increasing degrees of alteration.

Cosmic ray exposure (CRE) age measurements of CM chondrites reveal an unusual feature. Though other chondrite CRE ages range from several Myr to tens of Myr, generally with one peak age, CMs exposure ages are not longer than 7 Myr with 5 distinct age peaks. Because a CRE age reflects how long a meteorite is a separate body in space, the peaks presumably represent collisional events on the parent body (ies) [2].

In this study we systematically characterized the petrography of the CMs in each of the 5 exposure age groups to determine whether the groups have significant petrographic differences, with such differences probably reflecting different parent body (asteroid) geological processing, or multiple original bodies.

Samples and Method: We observed thin sections of 125 CM and CM-related chondrite by optical microscopy and scanning electron microscopy (SEM). Moreover, we made whole mosaics of each thin section by reflected light and backscattered electron imaging (Figure.1). We then grouped the meteorites into several groups based on the following nine petrographic criteria:

1. Abundance of chondrules
2. Abundance and thickness of chondrule rims
3. Chondrule sizes
4. Degrees of chondrule fracturing
5. Degrees of mafic silicate alteration in chondrules
6. Amount of metal iron
7. Amount of CAIs
8. Amount of tochilinite clamps
9. Brecciation

These criteria follow characterizations mentioned in Rubin et al. [1]. These grouping are performed qualitatively because this is the first attempt to grope them with their CRE ages and textures.

Some element maps are also made by SEM and some quantitative analyses were made of matrix by electron microprobe to compare compositions between each CRE age groups.

Fig. 1 BSE mosaic of Murchison
**Result and Discussion:** Figure 2 is the CRE age distribution map of CMs observed in this study. In this plot, the 73 best determined CMs are shown. According to this plot, there are several distinct peaks [2] and the large number of samples permits investigation of each CRE age group. We label the perceptible 5 peaks approximately around 0.2 Myr, 0.6 Myr, 1.5 Myr, 2.5 Myr and 4.3 Myr as group 1, 2, 3, 4 and 5 respectively and briefly summarize their characteristics below:

**Group 1:** the CRE ages in this group extend from 0.04 to 0.31 Myr (21 meteorites). It seems that a lot of meteorites in this group have considerable tochilinite and medium-sized chondrule rims and half of them are brecciated. However, the amount of chondrules and the degree of alteration are quite different between each meteorite in the group.

**Group 2:** the CRE ages extend from 0.49 to 0.65 Myr (8 meteorites). Most of the meteorites in this group have good chondrule rims. Although 6 meteorites have considerable tochilinite, the other 2 have no tochilinite but much sulfide. However, like group 1, the amount of chondrules and the degree of alteration considerably varies between these meteorites.

**Group 3:** This group is arbitrarily defined by 20 meteorites whose CRE ages from 0.8 to 1.9 Myr. In this group, more than half of meteorites have relatively many chondrules and good chondrule rims and are uniformly weakly altered. However, the amount of tochilinite and metallic iron varies significantly between each meteorite.

**Group 4:** This group has 20 meteorites and their CRE ages spread from 2.0 to 3.2 Myr. This group seems to have no common feature.

**Group 5:** This is composed of 4 meteorites whose CRE ages are from 4.0 to 4.6 and exceptional 7.1 Myr. These meteorites look similar though Santa Cruz, whose CRE age is 7.1 Myr, is significantly more altered than the others.

Figure 3 shows elemental maps of 4 meteorites selected from 3 different CRE age groups and one belonging to no CRE age group. From these images we can find that Mg is rich in ALH88045 and Cold Bokkeveld, however it is low in Murray and LEW90500, and Al and Ca are more abundant than Mg in LEW90500. As above, we can find the compositional differences between each meteorite. However we have only a small amount of compositional data, and it is therefore unclear whether the compositions of meteorites with similar textures are alike or not.

CAIs are very small and it is hard to determine their abundance, therefore they are not considered at this time.

According to above results, it seems that there is no definite common feature between meteorites in each group (except group 5). However there are some meteorites similar to each between the groups. For example, Cochabamba (CRE age: 2.1 Myr), Mighei (2.3 Myr), Y-791198 (3.0 Myr), Y-74662 (3.2 Myr), Murray (4.3 Myr) and ALH84033 (4.6 Myr) all have many chondrules, good rims, considerable tochilinite...
and only a small amount of metallic iron in mafic silicates, although there are minor differences. (Figure 4).

The CMs parent body(ies) is (are) composed of many kinds of textures because these meteorites are gathered around the same peaks in CRE age distribution map and the peaks therefore probably represent a successive collisional events on the same parent body or that all the sampled parent bodies have seen essentially identical processing [2].

**Conclusions:** In this study, although we sought correlations between the CRE age and textures of CMs, we found none. However we found CMs in each CRE age group have many kinds of textures and this means that the parent body(ies) of CMs is (are) complex and uniform. We also find some CMs have similar textures regardless of their CRE ages. As they have distinct CRE ages, there is a possibility that these meteorites came from the same parent body which experienced sequential collisional events or several nearly identical parent bodies. Further investigation is required to permit selection of the correct situation.

AMPHIBOLE IN MARTIAN METEORITE TISSINT: COMPOSITION AND IMPLICATION FOR WATER CONTENT OF PARENTAL MAGMA. K. B. Williams¹, Y. Sonzogni², and A. H. Treiman³,
¹Department of Geological Sciences, Brown University, Providence, RI 02912 (kelsey_williams@brown.edu),
²Lunar and Planetary Institute, Houston, TX 77058.

Introduction: Titanium-rich amphibole is present in melt inclusions in many martian (SNC) meteorites, suggesting a hydrous martian source melt. Amphibole has been reported in melt inclusions within olivine grains of Chassigny [1, 2], but only occurs in melt inclusions within pigeonite grains in olivine-phyric shergottites [3-9]. This study focuses on comparing amphibole compositions within the same SNC class, shergottites. Two olivine-phyric shergottites were analyzed: Tissint and Elephant Moraine (EETA) 79001, Lithology A.

While amphibole (commonly of kaersutitic composition) occurs as a minor phase in martian meteorites, the mineral is widespread and may be fundamental in understanding abundances in the martian mantle. Amphibole incorporates hydroxyl into its mineral structure on a site known as the O(3) site. The O(3) site can contain OH, F, Cl, and O²⁻. Previous compositional analyses of amphiboles in martian meteorites show that the amphiboles have low halogen abundance, implying high proportions of OH⁻ and/or O²⁻ on the O(3) site [5, 10]. This study seeks to expand the compositional database of amphibole occurring in martian meteorites and to provide amphibole water content estimates that can then be used to constrain water content of the parental magma.

Sample Petrography: Amphiboles were identified in polished thin sections of Tissint and EETA79001A. The amphiboles occur as subhedral grains that are pleochroic yellow-orange to light brown and range up to 15 µm in diameter (although exposed surface rarely exceeds 5 µm). Consistent with previous observations of amphibole in shergottite [3-9], amphibole-bearing melt inclusions were only observed in the core of pigeonite grains, never in augite or olivine. Inclusions are < 25 µm in diameter and irregular in shape. Aside from silicic melt, amphibole was the only phase observed in the melt inclusions.

Methods: Chemical analyses were obtained with the Cameca SX-100 Electron Microprobe at Johnson Space Center, using a 15 keV accelerating potential and a 10 nA, 1-micron-diameter beam. Point analyses of the melt in the inclusion and line analyses of the pyroxene grain surrounding the melt inclusion were taken for each of the analyzed amphiboles.

Stoichiometric calculations of amphibole formulae were performed according to a 23 O normalization and assuming halogens and hydroxyl fully occupy the O(3) site (OH⁺ + F⁺ + Cl⁻ sum to 2.0 atoms per formula unit (apfu)). Cation normalization schemes could potentially yield more precise structural formulae, however the 23 O normalization has been proven to give reasonable results consistent with those of other methods [11]. If ferric iron is present in the sample, an all-ferrous-iron derived formula may have cation sums that are too high [12]. A method that combines all-ferrous and all-ferric calculations has been suggested to obtain a more representative cation total [11-13]. This method was considered for calculations on amphibole data from Tissint and EETA 79001A. The all-ferrous and all-ferric calculations resulted in chemical limits that varied amphibole water content by less than 0.1 weight percent (wt.%) H₂O. Given the narrow range of variability, it is safe to treat all iron as one oxidation state. All-ferrous conditions were assumed as this method provides the more conservative water content estimate.

Essential to water content estimation is the assumption that halogens and hydroxyl fill the O(3) site. This assumption may not be accurate because O²⁻ is potentially present on the site. Variability in iron oxidation and the possibility of internal amphibole dehydrogenation provide uncertainty to OH estimations based on microprobe analyses [5, 10, 14]. Shock effects on the reactions that control OH⁻ and O²⁻ content contribute to the complications in ascertaining water content from measured amphibole compositions, particularly in meteorites [15]. In the absence of direct analyses for H and/or ferric iron, we assumed OH⁻ + F⁺ + Cl⁻ = 2 apfu.

We used the computer code Petrolog3 [16] to model fractional crystallization of a bulk Tissint composition to that of the amphibole-bearing melt inclusions. The composition of the melt in each inclusion prior to amphibole crystallization was estimated by...
mass balance calculations based on the proportion of amphibole in the inclusion. Petrolog3 models of crystallization were calculated for low pressure (1 bar) and an oxygen fugacity of iron-wüstite [17, 18]. Using the results from Petrolog3 crystallization models and the compositional data obtained from the Tissint sample, we determined the proportions of olivine and pyroxene crystallization prior to melt inclusion entrapment. We applied the obtained olivine and pyroxene crystallization percentage to the water content of the melt inclusions in order to estimate water content of the parental martian magma.

Table 1: Representative electron microprobe analyses of amphibole and melt pairs in Tissint and EETA79001A. Amphibole formulae were normalized to 23 O and calculated so that OH⁺F⁺Cl⁻=2.

<table>
<thead>
<tr>
<th>Element</th>
<th>Tissint Amph</th>
<th>Tissint Melt</th>
<th>EETA Amph</th>
<th>EETA Melt</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>37.53</td>
<td>62.13</td>
<td>36.33</td>
<td>73.91</td>
</tr>
<tr>
<td>TiO₂</td>
<td>7.19</td>
<td>0.52</td>
<td>10.52</td>
<td>0.29</td>
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<tr>
<td>Al₂O₃</td>
<td>14.38</td>
<td>17.70</td>
<td>13.56</td>
<td>18.40</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.10</td>
<td>0.00</td>
<td>0.41</td>
<td>0.01</td>
</tr>
<tr>
<td>FeO</td>
<td>18.08</td>
<td>4.05</td>
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</tr>
<tr>
<td>MnO</td>
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<td>0.00</td>
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<tr>
<td>NiO</td>
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<tr>
<td>MgO</td>
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<td>CaO</td>
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<tr>
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<tr>
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</tr>
<tr>
<td>F</td>
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</tr>
<tr>
<td>Cl</td>
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<tr>
<td>H₂O</td>
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</tr>
<tr>
<td>F+Cl+O</td>
<td>-0.16</td>
<td>-</td>
<td>-0.27</td>
<td>-</td>
</tr>
<tr>
<td>Total</td>
<td>98.94</td>
<td>98.21</td>
<td>99.09</td>
<td>100.95</td>
</tr>
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</table>

Results: Phase Compositions. We obtained compositions for 7 amphibole-melt pairs in pyroxenes from Tissint, and 6 from EETA79001A. Amphibole compositions are ferro-kaersutitic [12] and similar to those reported in other olivine-phryic shergottites [3-9]. Table 1 presents representative microprobe analyses of amphibole, melt, and host-pyroxene for melt inclusions in both Tissint and EETA79001A. Tissint amphiboles (Mg′ = .38) are less magnesian than those in EETA79001A (Mg′ = .48). Both samples have low fluorine and chlorine abundances (F < 0.4 wt.%, Cl < 0.03 wt.%) suggesting a high OH⁻ (or O²⁻) occupation of the O(3) site in the amphibole structure. Stoichiometric calculations suggest 1.6-1.8 wt.% H₂O in the amphiboles. Melt inclusion glasses are silicic, with 55-65 wt.% SiO₂ in Tissint and 70-75 wt.% SiO₂ in EETA79001A. Host pyroxenes are zoned from picrolite cores (En₄₅₋₇₀Fs₂₄₋₄₃Wo₁₋₄₋₁₉ in Tissint; En₅₁₋₄₁Fs₂₅₋₃₆Wo₁₋₁₁ in EETA) to augite rims (En₄₂₋₅₅Fs₂₁₋₃₇Wo₂₁₋₂₅ in Tissint; En₅₈₋₆₂Fs₂₄₋₃₉Wo₂₁₋₂₃ in EETA). These pyroxene compositions are consistent with previous analyses of Tissint samples [19-21].

Water Content of Tissint Parental Magma. Corresponding compositional analyses of the inclusion melt and host pyroxene grain were obtained for each amphibole studied. We made rough estimates of the proportion of amphibole in each inclusion (ranging 30-50% amphibole). Using the estimated proportion of amphibole in the inclusion, stoichiometrically calculated amphibole composition, and corresponding melt composition, we determined the composition of the melt in the inclusion prior to amphibole crystallization (here referred to as “trapped melt prior to amphibole crystallization”, or TPMA).

Elemental comparison of TPMA compositions and bulk Tissint composition (deduced from fusion crust analyses of our sample and consistent with published bulk Tissint compositions [19]) gives insight into the crystallization history of the sample. The enrichment of elements incompatible with olivine and pyroxene (Ti, Al, Na, K, P) will demonstrate the degree of fractionation between the melt inclusion and the parent magma. Based on this relationship, we calculated a proportion of olivine and pyroxene crystallization prior to melt entrapment (~80%). We used this crystallization proportion as an observation-derived comparison for values generated by the Petrolog3 modeling.

Petrolog3 models of Tissint crystallization were run in attempt to determine the proportions and compositions of olivine and pyroxene needed to produce a residual melt with the same composition of the TPMA. The accuracy of our results were assessed by ensuring that the compositions of olivine, pyroxene, and residual melt produced by the models matched those measured in the meteorite. Petrolog3 is successful in reproducing olivine proportions and compositions observed in the Tissint meteorite. The program predicts 25 volume percent (vol.%) Fo₆₆₋₈₁ olivine crystallization from the parental magma, which is consistent with the olivine proportions and compositions reported in previous studies (24-29 vol.% Fo₆₆₋₈₁) [21, 22]. Modeled pyroxene crystallization is 45 vol.% and agrees with petrographic estimates placing pyroxene proportion at 46 vol.% [22]. Petrolog3 crystallization models are less effective in reproducing the composition of pyroxenes in Tissint. Figure 2 compares a representative rim-to-core analysis of a Tissint host pyroxene to the pyroxene compositions predicted by two Petrolog3 crystallization models. Inconsistency between predicted and observed pyroxene compositions may reflect the unreliability in modeling the crystallization of unusual
(martian) melt compositions with the crystallization models used by Petrolog3. Despite this discrepancy, we used the pyroxene proportions suggested by Petrolog3 as conservative estimates. The total proportion of crystallization generated by Petrolog3 (70%) is less than the proportion we calculated using the measured abundances of elements incompatible with olivine and pyroxene (80%).

Figure 2: Al2O3 and Mg' (Mg/(Mg+Fe)) of pyroxenes from two Petrolog3 model crystallizations and one representative Tissint pyroxene line analysis. The blue arrow depicts direction of crystallization.

Once olivine and pyroxene crystallization proportions were established, the total proportion was used to trace the crystallization history of Tissint backwards from TPMA composition to parental magma composition. The amphibole-bearing inclusions contain 0.49-0.87 wt.% H2O, 0.20-0.34 wt.% F, and 0.01-0.44 wt.% Cl. These inclusions are calculated to have formed after 70-80% crystallization of Tissint’s parent magma. Thus, if no water or chlorine was lost during the crystallization, Tissint’s parent magma must have contained 0.13-0.26 wt.% H2O, 0.04-0.10 wt.% F, and < 0.13 wt.% Cl. This water content approximation agrees with studies ofapatite compositions that suggest shergottite parent magmas (depleted and enriched) contained 0.07-0.29 wt.% water prior to degassing [23]. The referenced study used the stoichiometry and halogen content ofapatites to calculate a percent crystallization of the parental magma and infer an initial water content. Consistency between our estimated water content and that of [23] has further implication towards all shergottite parent magmas, as the Tissint parental melt is suggested to have originated from a mantle source similar to that of other olivine-phryic shergottites [24].

Presence of O2- on the O(3) amphibole site has been dismissed for the purpose of this study, however Ti- oxy kaersutite has been shown to be stable at 1 bar pressure [10, 14]. Oxy-amphibole component would reduce estimates for water content in the amphibole, and consequently the parental magma. Internal dehydrogenation, iron oxidation, and shock-induced metamorphism all exhibit effects on possible O(3) occupation by O2 [5, 10, 14, 15]. Direct analyses for H in the amphiboles (and the adjacent glass) would remove this uncertainty, and greatly improve our estimate of the water contents of the amphibole and parental magma.