Papers Presented at the

16th Annual

Summer Intern Conference

August 11, 2000
Houston, Texas

2000 Summer Intern Program for Undergraduates
Lunar and Planetary Institute
Papers Presented at the

Sixteenth Annual
SUMMER INTERN CONFERENCE

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Houston, Texas

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Sponsored by
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The Lunar and Planetary Institute is operated by the Universities Space Research Association under Contract No. NASW-4574 with the National Aeronautics and Space Administration.
8:00—8:30 a.m.  **Breakfast** (LPI Great Room)

**Intern Presentations**  
Chair: Dr. Paul Spudis

8:30 a.m.  
**LYDIA L. BOROUGH**  
Advisor: Paul D. Spudis  
*The Stratigraphy of Lava Flows in Northern Oceanus Procellarum, Moon*

8:50 a.m.  
**ALEXANDRA L. DICKIE**  
Advisor: Pat McGovern  
*Measurements of Extension and Strain at Alba Patera, Mars, from MOLA Topography*

9:10 a.m.  
**VERONICA L. LaMOTHE**  
Advisor: Julianne Moses  
*Ammonia and Phosphine Photochemistry of the Jovian Troposphere*

9:30 a.m.  
**TERESA G. LONGAZO**  
Advisors: David S. McKay and Sue Wentworth  
*Biological and Inorganic Weathering of Olivine*

9:50 a.m.  
**KOTARO OE**  
Advisor: Gordon McKay  
*REE and Sr Partition Coefficients for Nakhla Pyroxenes*

10:10 a.m.  
**Break**

**Intern Presentations**  
Chair: Dr. Mike Zolensky

10:30 a.m.  
**KELLY PETERSON**  
Advisor: Walter Kiefer  
*Gravity Studies of Coronae on Venus*

10:50 a.m.  
**JOEL SAYLOR**  
Advisor: Michael Zolensky  
*Occurrence of Fluid Inclusions in CM2 Chondrites*

11:10 a.m.  
**RACHEL T. SCHELBLE**  
Advisor: Carlton Allen and Frances Westall  
*The Relationship Between Hematite and Bacteria in the Gunflint Iron Formation from Ontario, Canada: Implications for the Hematite Deposits on Mars*

11:30 a.m.  
**RACHEL L. SHANTEAU**  
Advisor: Robert Herrick  
*The Effects of Impact Angles on Martian Craters Using MOLA Data*

11:50 a.m.  
**JENNIFER G. WARD**  
Advisor: Laurel Kirkland  
*Surface Composition of Mars: Examination of Syrtis Major Using Thermal Infrared Data*

12:10 p.m.  
**ABIGAIL A. WASSERMAN**  
Advisor: Carl Agee and Dave Draper  
*Phase Relations of a Model Martian Mantle at 5 GPa*

12:30 p.m.  
**Adjourn**
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**INTRODUCTION:**

The lunar surface is composed primarily of two distinct geologic units. The bright, low titanium, low iron highlands make up the majority of the surface area and volume of lunar crust. The darker maria are high iron, high to low titanium, regions resembling terrestrial basaltic lavas and occur primarily on the near side of the Moon. The maria occur in basins, assumed to be the result of impacts which fractured the highlands crust, allowing lava to seep through the cracks and flood the surrounding highlands regions.

The lunar maria are important because they tell us about the thermal history and interior composition of the Moon, and because of their relatively high concentration of useful materials. The titanium-oxide mineral ilmenite traps solar-wind hydrogen, which can be mined along with other gases for fuel and life support uses in space [1,2]. Information about the structure of lava flows, their complex history and composition therefore become important in the advancement of human space travel.

Northern Oceanus Procellarum, a mare region on the northeastern near side of the Moon, is the area chosen for mapping in this study. Procellarum is the largest maria on the Moon, and occupies most of the western near side, and has not been previously mapped using the methods described here.

The NASA / Department of Defense Clementine mission in 1994 provides spectral reflectance data for the entire lunar surface, which show in high resolution (200m/pixel) the iron and titanium compositional variations of the lunar surface. These variations reveal the locations of different lava flows and differences in crater ejecta, which can be used to map the three dimensional stratigraphy of flows in the region.

As a surface is exposed to space, it accumulates impact craters, and the cratering rate has been calibrated using radiometrically dated samples returned by the Apollo missions. Crater densities derived from Lunar Orbiter IV photographs reveal the ages of the features in Oceanus Procellarum, providing a map showing the sequence of mare flooding in this region.

**METHODS**

Clementine image swaths, obtained in a polar orbit, were used to piece together mosaics of the region of study using the ISIS software from the USGS [3]. A mosaic approximating the "true" color of the area was made with the 415 nm filter controlling the blue channel, 750 nm controlling the green, and 950 nm controlling the red channel. A false color mosaic was constructed with the 415/750 ratio controlling the blue, the 750/950 ratio controlling the green and the 750/415 ratio controlling the red channels. This exaggerated and extremely brightly colored image helps to eliminate noise caused by surface maturity, albedo, and shadowing. The result shows, in extreme contrast, the compositional differences of the lunar surface material, with highlands appearing bright red and mare materials spanning a wide range of colors.

Compositional maps of iron (FeO) and titanium (TiO₂) abundance were constructed using the methods described by Lucey et al. [4,5] and Blewett et al. [2]. The resulting images provide maps of Fe and Ti at 200 m/pixel resolution. This method has been calibrated using known iron and titanium concentrations from Apollo landing sites, and Lunar Prospector gamma-ray data [6].

Using compositional differences obtained from true- and false-color mosaics, as well as the iron and titanium maps, the generalized flow boundaries were located and mapped. The resulting 52 sub-units were studied using ISIS software. The regolith covering each flow was measured for composition, carefully avoiding crater ejecta and exposed highlands material. These data provided a preliminary grouping of these units into approximately 20 generalized units, a list that was continually modified and shortened throughout the study with the arrival of new data (Table 1).

A defined area of each of the flows was studied using Lunar Orbiter IV photographs to determine crater densities. All craters larger than 500 m in diameter were counted within those defined areas. The resulting crater densities, number (N) of craters (D > 500m) per unit, was then used to determine the relative ages of the flows. The grouping of sub-units into larger units was then refined and finalized based on relative age. Seven more units were identified. Each unit was dated by comparison of crater densities with radiometrically determined ages of Apollo (Figure 3, Table 1). A geologic map was constructed, showing the flow units and highlands, shaded according to age (Figure 1).

The composition of the lava associated with each flow unit was determined by studying the ejecta blankets of craters. Craters that punch through the regolith layer into the lava, but stop short of puncturing through the lava, throw out an ejecta blanket that is composed primarily of relatively fresh lava material and a negligible amount of highlands contamination. These ejecta blankets are measured for their iron and titanium concentrations, which correspond very closely with the compositions of the lavas underlying the regolith (Table 1).

Larger craters which punch all the way through the lava layer into the highlands material produce an ejecta blanket which contains some mixture of highlands and lava materials. Once the iron concentration in the lava and the highlands iron concentration was measured, the ratio of highlands to lava material originally excavated by the impact was calculated. The total volume of excavated material was calculated using a standard model of crater geometry. The geometric assumptions are a standard depth (d) to diameter (D) ratio of 10d = 1D, and a crater shape similar to an inverted spherical cap with dimensions $V = \frac{\pi d^2}{6} \left( 3r^2 + d^2 \right)$, where $V = \text{volume}$, $d = \text{crater depth}$ ($\approx 0.1D$), and $r = \text{radius of crater rim}$. For each prominent crater that penetrated through the lava, the crater diameter was measured using the mosaic imagery, and the volume calculated using the previous equation. Using these values and the highlands to lava ratio calculated earlier, the thickness of lava was calculated. These values were plotted.
Stratigraphy of Northern Oceanus Procellarum, L.L. Boroughs and P.D. Spudis

on a map of the region, and depth contour lines drawn to produce an isopach map of the area (Figure 2).

The geologic map and isopach map (Figures 1 and 2) were analyzed using pixel counting to determine surface area and depth of each flow unit, as well as total volume of lava in the region. Areas within each flow unit and major highlands area were measured for total pixels. The scale of the image was used to determine total surface area of each flow (Figure 1), with corrections made for small areas within the flow regions lost due to small highlands fragments, line thickness of outlines, and crater ejecta blankets. The pixels within each flow unit in the isopach map were similarly counted to determine the average thickness of each flow (Figure 2). The total surface area of the flows in the region was combined with this average thickness data to determine the total volume of lava in the region, which was then used to calculate the magma production rate as a function of time.

RESULTS AND DISCUSSION

Northern Oceanus Procellarum contains some very young flows, several of which dispute the long-held view that the Moon ceased volcanic activity around 3 Ga [7,8]. The flows in this region show a range of 2.4 billion years, from 1.2 to 3.6 billion years old (Table 1). Flow Units 1 and 3 and the Rumker area were deposited in the Late Imbrian times (3.8-3.2 Ga), and Flow Units 2, 4, 5, and 6 in the Eratosthenian era (3.2-1.1 Ga).

The lavas show little variation in iron, ranging from 19.6% to 22.1% FeO, considerably higher than the 12.9% average calculated for the highlands. A larger variation is seen in titanium, from less than 0.5% TiO2, similar to the very low titanium concentration in the highlands, to a high of 6.0% TiO2. The general trend in this region is for the higher Ti lavas to be younger, contradicting the generalization that the lower Ti basalts on the Moon tend to be younger than higher Ti basalts [1].

Geologic mapping in this study shows refinement over previous results. Whitford-Stark and Head [10] show some slight differences in flow boundaries and groupings compared to the results published here, and even larger differences were mapped by Wilhelms [11]. Discrepancies in both cases can be explained by the arrival of new, higher resolution Clementine compositional data in 1994 (after the previous studies [10,11] were published), which is better and more reliable for this purpose.

Relative ages of some flows in this study also show refinement over previous studies. Both Whitford-Stark and Head [10] and Wilhelms [11] show Flow 2 as older than Flow 3, which is contradicted by this study. There are a considerable number of secondary features in Flow 2 which result from impacts in the highlands region to the north of the study area. These secondary features were carefully avoided when measuring crater densities in this study. If included, the secondary features would incorrectly show a higher crater density in Flow 2 than Flow 3. I estimate Flow 2 to be approximately 1.3 billion years younger than Flow 3.

The accumulation of lavas flows covers 462,000 km² of the study area, a significant 71% of the 653,000 km² total study area. Mare basalts cover an average of 17% of the total lunar surface [12]; however, this study region did not include much of the surrounding highlands material. The lavas in this region make up a significant 10% of the total 6.3 x 10⁶ km³ of lava [12] on the Moon’s surface.

The average depths of lava sequences in the region, from 232 to 1100 meters, are larger than the thickness assumed to be the lunar averages, less than 400 meters, especially in Sinus Roris in the northeastern corner of the study area. However, the average depths are consistent with the estimate that most maria are less than 2 km thick [12].

The total volume of lava produced, 266,000 km³, represents approximately 2.7% of the total 1 x 10⁷ km³ volume of mare basalts on the Moon [12]. Based on an average crustal thickness of 53 km in the area [11], and therefore an estimate of crustal volume of 5.46 x 10⁷ km³ for the study area, the lavas make up only 1% of the total crustal volume of the study area.

The total lava volume of 266,000 km³ produced over a period of 2.4 billion years, yields an average magma production rate of 1.1 x 10⁴ km³ per year in this region of the Moon for the total time span of active production. This figure represents a rate that may have fluctuated during different time periods. If constant throughout the time of production, the above rate is considerably lower than the 150 x 10⁴ km³ per year estimates of lunar magma production for the Late Imbrian [12], but is similar to the 1.3 x 10⁵ km³ per year lunar average estimates for the Eratosthenian [12].

CONCLUSIONS

Oceanus Procellarum exhibits a complex geologic history, with volcanism continuing until around one billion years ago. The relatively new Clementine color images allow a more precise study of the area than was previously possible. Contacts were previously mapped based on color, but the Clementine data provide the first opportunity to study high-resolution composition data and provide the first estimates of lava thickness. The methods described here apply to all areas of the lunar surface, and can be useful in unraveling the history, structure, and composition of other areas of the Moon.

REFERENCES

Stratigraphy of Northern Oceanus Procellarum, L.L. Boroughs and P.D. Spudis

Figure 1: Geologic map of Northern Oceanus Procellarum. Gray areas are lava flows. Black areas are highlands materials and gaps in coverage. White areas are crater ejecta blankets that obscure flow boundaries. Darkest flows (i.e. #1) are oldest, and grays lighten to the youngest flows (i.e. #4). See Table 1 for unit information.

Figure 2: Isopach map of Northern Oceanus Procellarum showing regional variations in flow depth.

TABLE 1: Oceanus Procellarum Stratigraphic Unit Data

<table>
<thead>
<tr>
<th>UNIT</th>
<th>CRATER DENSITY N/km²</th>
<th>2EST. AGE Ga</th>
<th>Flow FeO</th>
<th>TiO₂</th>
<th>Regolith FeO</th>
<th>TiO₂</th>
<th>4THICKNESS (m) Min.</th>
<th>Max.</th>
<th>Avg. (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flow 1</td>
<td>7.2 ± 0.30</td>
<td>3.6</td>
<td>19.6</td>
<td>3.5</td>
<td>16.3</td>
<td>1.8</td>
<td>0</td>
<td>1200</td>
<td>376</td>
</tr>
<tr>
<td>Flow 2</td>
<td>3.1 ± 0.20</td>
<td>2.1</td>
<td>20.5</td>
<td>&gt;0.5</td>
<td>19.0</td>
<td>0.6</td>
<td>0</td>
<td>1400</td>
<td>703</td>
</tr>
<tr>
<td>Flow 3</td>
<td>5.4 ± 0.22</td>
<td>3.4</td>
<td>19.9</td>
<td>1.9</td>
<td>20.1</td>
<td>2.2</td>
<td>0</td>
<td>1200</td>
<td>1100</td>
</tr>
<tr>
<td>Flow 4</td>
<td>1.9 ± 0.21</td>
<td>1.2</td>
<td>22.1</td>
<td>6.0</td>
<td>21.3</td>
<td>4.8</td>
<td>200</td>
<td>1200</td>
<td>775</td>
</tr>
<tr>
<td>Flow 5</td>
<td>3.4 ± 0.30</td>
<td>2.3</td>
<td>21.1</td>
<td>4.2</td>
<td>21.0</td>
<td>3.8</td>
<td>0</td>
<td>1200</td>
<td>475</td>
</tr>
<tr>
<td>Flow 6</td>
<td>2.5 ± 0.26</td>
<td>1.5</td>
<td>21.1</td>
<td>4.6</td>
<td>18.9</td>
<td>2.6</td>
<td>0</td>
<td>1000</td>
<td>232</td>
</tr>
<tr>
<td>Rumker</td>
<td>5.8 ± 0.72</td>
<td>3.4</td>
<td>---</td>
<td>---</td>
<td>19.1</td>
<td>1.0</td>
<td>600</td>
<td>1000</td>
<td>909</td>
</tr>
</tbody>
</table>

1 See Figure 1 for unit locations.
2 Age interpreted by comparing crater densities with those of sampled sites and compared with calculated dates of other areas [8, 9] See Figure 3.
3 Compositions in wt. % determined from Clementine images, using the method of Blewett et al. [2] and Lucey et al. [4,5], accurate to ± 0.5%.
4 Thickness and Exposed Area from pixel counts (see text). See Figure 2.

Figure 3. Crater densities v. age for flows in Oceanus Procellarum compared with radiometrically determined ages of rocks from sampled sites and data from the literature. See Table 1 for flow characteristics and Figure 1 for locations.
Measures of Extension and Strain at Alba Patera, Mars, from MOLA Topography

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Advisor: Patrick J. McGovern, Lunar and Planetary Institute, Houston, TX

Introduction
Alba Patera is a broad shield volcano 2000 km in diameter and 6 km tall situated in the Tharsis Plateau of Mars. The impressive size of Alba Patera and similar giants in the Tharsis region is attributed to the lack of plate movement over hot spots feeding the Tharsis Rise. Alba Patera, however, is particularly unique in the complex system of extensional features that dominate the flanks of the volcano. The extension, clearly identifiable from Viking Orbiter images, is expressed typically through normal faulting. Single normal faults can be seen, however more commonly faulting occurs in paired sets resulting in the formation of a topographically low, downfallen block known as a graben. Both simple and complex grabens, formed circumferentially to the center of Alba, surround the caldera complex beginning roughly half way down the volcano’s flanks, at least 200 km from the caldera [1]. Additional extensional features include N-S to NE-SW trending graben on the lower flanks formed radial to the Tharsis Rise.

New altimetry data from MOLA (Mars Orbital Laser Altimeter) aboard the Mars Global Surveyor spacecraft allow quantitative measurements of fault topography to be made. The purpose of this paper is to present extensional data of Alba Patera.

Theory
Given MOLA-derived measurements of the vertical throw of the fault, \( d \), and an assumed initial fault angle, \( \alpha \), the extension, \( E_s \), can be calculated as [2] 

\[ E_s = d / \tan(\alpha). \]  

From fault mechanics [3], normal faults form initially with a slope of \( 60 \pm 15^\circ \) and form striking perpendicular to the direction of extensional stress [1]. Measured faults carps are often considerably less than this assumed value due to post-faulting modification such as mass wasting and lava infilling.

The strain over an area is simply a measure of the change in length after deformation has occurred [3]. Here, instead of comparing initial and final length, the extensional deformation is found as a function of the extension experienced across a graben system. Strain, \( \varepsilon_s \), expressed as a percentage, is calculated as 

\[ \varepsilon_s = E_s / X \]  

where \( X \) is the distance between the first and last scarp coordinate. Since each orbit path is not oriented exactly normal to the graben trend, a correction must be made giving 

\[ \varepsilon_s = E_s / X \sin(\theta) \]  

where \( \theta \) is the angle at which the fault intersects the orbit path.

Methodology
Viking Orbiter images, which divide the entire volcanic feature into several smaller sub-areas covering approximately 5 degrees of longitude and 5 degrees of latitude, were used to digitally locate interpreted graben systems with Matlab 5.3. The regions studied are those where the circumferential graben system is the dominant extensional feature. This includes one previously examined area to the northeast of the caldera complex centered at 45°N, 112°W and three regions to the east centered at 35°N, 102°W; 40°N, 102°W and 40°N 97°W.

The location and angle of intersection of MOLA orbits and digitized faults were found; only orbit paths oriented closest to perpendicular to the faulting trend were studied. The angle of intersection was used both in the strain calculation (3) and in converting apparent faults carps slope to true slope measurements. The intersection of faults along each orbit path was used to identify the position of graben walls along the orbit profile and helped to avoid misinterpreting craters or other non-fault features as grabens (Figure 1). Again using Matlab, these faults carps were digitally traced and identified as either inward or outward facing relative to the center of the volcano. Measurements of the throw for each scarp face were found based on vertical displacement. Extension and maximum slope were found for every scarp, while regional extension was calculated for each orbit profile using equations (1) and (3).

For comparison with several axisymmetric finite element loading models, each orbit profile was projected to radial distance (km) from the volcanic center. Extension measurements along each profile were summed into 24 km running windows of radial distance. Extension values generated from the models were subjected to the same windowing process.

Results
The distribution of extension in the northwestern area (region 1) is shown in Figure 2. Extension vs. radial distance plots (Figure 3) show features concentrating at a radius between 200 and 400 km with most peaks at approximately 300 km. The average peak extension for the region was 600m. Extension vs. radial distance plots characteristic of the western flanks of the volcano are shown in Figures 3a and 3b. Orbit 10424, Figure 3a, shows one dominant peak in extension located at roughly 300 km with a magnitude of approximately 1000m that gradually tapers out with increasing radius from the caldera. Orbit 10663, Figure 3b, includes two clear extension peaks, at roughly 250 and 300 km, with a magnitude of approximately 400m. Other profile also show peaks typically separated by less than 100 km of radial distance. For the 26 orbit paths included in this region, the median strain value was 1.4% with a relatively low median faults slope of 15° and median extension per profile of 1300m and 77m per fault. Overall extension was 17877m and 14584m for outward and inward faults, respectively.

For the three areas to the east of the volcano’s center, the
faulting appears more complex and less localized. Extension vs. radius plots for region 2 show double extension peaks were concentrated at distances of approximately 300 and 475 km. The average profile peak was a high 1000m with several orbits showing extension peaks closer to 1500m. Orbit 10977, Figure 3c, shows a high magnitude single peak located at a radius of 350 km. Several other orbits crossing the region show extension typically peaks once or with a second peak of lower magnitude often well under 500m. This region includes the most pristine faults with a median strain value of 1.3%, median slope of 20°, and the highest profile median for extension at 1486m. Total extension for outward and inward facing faults was similar at 24788m and 24972m, respectively.

Lower average strain values were obtained for both remaining regions on the eastern flanks. Additionally, most of the graben systems for these areas are concentrated in the northwest and far southeastern regions resulting in lower overall extensional values. For the southeastern area (region 3) extensional features occur over a wide range of radius from 300 to 600 km with an average magnitude of 400m. Orbit 10386, Figure 3d, shows an extension vs. radius plot for one of 23 orbit paths that cross both region 3 and the region 2 directly to the north. This profile combines the lower magnitude peak of the more southerly region concentrated at a radius of 500 km, with the northern region's higher magnitude peak found further up the eastern flank. Median values for strain, extension per fault and extension were 0.642%, 34m and 715m, respectively. The median slope was a low 12°. Outward facing faults showed more extension overall (11798m) than inward facing faults (9846m). Overall extension for outward and inward facing faults was the lowest of all areas with values of 5910m and 6832m, respectively.

Discussion

There are currently three plausible models for the evolution of Alba Patera's circumferential graben system. The first concerns the possibility that these concentric grabens are the bounding faults of the large subsided caldera complex. Such a model would predict inward facing extension to be significantly greater than outward facing. However, total extension is greater on outward facing faults than inward facing faults in regions 1 and 3, and in region 2 there is no significant difference. Where outward facing faults on the mid-flanks of Alba Patera show greater extension this is suggestive of uplift of the summit relative to the flanks. Such a scenario supports sill inflation rather than an episode of caldera subsidence. However, for the heavily faulted eastern area (region 2) inward facing faults exhibit similar extension with outward facing faults. Many of the faults studied in these regions, particularly those east of the volcano oriented radial to the Tharsis Plateau, have formed well after the volcanic and tectonic events that formed Alba Patera [2]. Such faults would not be expected to show a preferred facing.

Alternative explanations discussed by McGovern et al. [4] include two subsurface loading models: sublithospheric buoyant (bottom) loading and intra-lithospheric sill loading. Previously, only bottom loading models have been compared to extension data for region 1 [5]. These models produce low magnitude, smooth, single peak extension curves (Figure 4a). Further extension measurements in region 1 (Figure 3a) also show the characteristic single extension peak interpreted to be consistent with bottom loading [5]. However, bottom loading models fall well short of the magnitude of extension required (compare Figure 3a with Figure 4a). Furthermore, the low lithospheric thickness required for bottom loading models to generate a mid-flank strain peak is inconsistent with gravity data [4]. Thus bottom loading cannot explain the extension observed at Alba Patera.

In contrast, sill loading models [4] can produce extension distributions broadly consistent with those observed on Alba Patera. In models 2 and 3 (Figures 4b and 4c) sill loads were applied relatively late in the volcano’s history. Late loading increases observed extension relative to models with loading applied evenly through the volcano’s history. Model 3 (Figure 4c) has a relatively shallow sill and low edifice stiffness, which also boost extension. Multiple sill models (Figure 4d) produce multiple extension peaks above the edge of each sill. These models suggest that single peaked MOLA extension curves result from a single episode of sill intrusion. Multiple-peaked MOLA curves suggest either multiple sill inflation episodes (with graben concentrated at the edges of each advancement), dike intrusions at edges of multiple sills or dike intrusions formed after the formation of the Alba Patera edifice.

The models in Figures 4a-d cannot generate extension magnitudes in the upper range of those observed, Figures 3a-d. Thus, purely tectonic processes are unlikely to produce the extensional features observed in regions where extension peaks at greater than several hundred meters. We therefore suggest that a large fraction of Alba's graben are a result of relatively young subsurface dike propagation. This is especially likely in region 2, which has the most pristine faults (implying youth).

---

**TABLE 1. Parameters For Individual Regions**

<table>
<thead>
<tr>
<th>Map region</th>
<th>No. of orbits</th>
<th>Median strain</th>
<th>Median fault slope</th>
<th>Total outward extension (m)</th>
<th>Total inward extension (m)</th>
<th>Profile median extension per fault (m)</th>
<th>Profile median extension overall (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. 45°N, 112°W</td>
<td>29</td>
<td>1.4%</td>
<td>15°</td>
<td>17877</td>
<td>14584</td>
<td>77</td>
<td>1300</td>
</tr>
<tr>
<td>2. 40°N, 102°W</td>
<td>30</td>
<td>1.3%</td>
<td>20°</td>
<td>24788</td>
<td>24972</td>
<td>60</td>
<td>1486</td>
</tr>
<tr>
<td>3. 35°N, 102°W</td>
<td>27</td>
<td>0.64%</td>
<td>12°</td>
<td>11798</td>
<td>9846</td>
<td>34</td>
<td>715</td>
</tr>
<tr>
<td>4. 40°N, 107°W</td>
<td>29</td>
<td>0.7%</td>
<td>12°</td>
<td>5910</td>
<td>6832</td>
<td>40</td>
<td>450</td>
</tr>
</tbody>
</table>

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**References**

1. Dickie: Extension at Alba Patera, Mars
3. McLaughlin et al.: Constraints on Sill Inflation at Alba Patera
Conclusion

Continued investigation of MOLA data have shown that extensional features on the mid-flanks of Alba Patera are characteristic of both single and multiple distribution peaks. Single peak distributions may be the result of one intrusive episode during the evolution of the volcano while a more complex distribution of extension results from multiple episodes of intrusion, some of which may have formed later than the evolution of the Alba Patera edifice itself.

REFERENCES


Figure 1. MOLA profile with digitized fault scarps, both inward and outward facing, and impact crater (far right).

Figure 2. Distribution and magnitude of extension for region 1.

Figure 3. MOLA-derived extension for 24-km running windows for region 1 (a and b), regions 2 and 3 combined (c and d).

Figure 4. Model-derived extension for 24-km running window for sublithospheric bottom loading model (a) and intra-lithospheric sill loading models (b-d).
Ammonia and Phosphine Photochemistry of the Jovian Troposphere
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We investigate ammonia and phosphine photochemistry on Jupiter by developing a one-dimensional steady-state model for the stratosphere and troposphere that accounts for hydrogen, carbon, nitrogen, oxygen, and phosphorus chemistry, vertical eddy diffusion, molecular diffusion, condensation, and radiative transfer (including multiple Rayleigh scattering). This model is unique in that it includes numerous organo-nitrogen reaction schemes and simultaneously models the stratosphere and troposphere. We compare the results of three cases (baseline model, high-acetylene model, and high-eddy-diffusion-coefficient model) with infrared and ultraviolet observations. We find that organo-nitrogen reactions are not very important in Jupiter’s troposphere, and the acetylene mole fraction must be less than $10^{-7}$ in the troposphere in order for HCN to be within observational limits. The tropospheric eddy diffusion coefficient must be less than $10^7$ cm$^2$/s$^1$. The implications for chromophores on Jupiter and Saturn are also discussed.

Introduction: A variety of instruments including remote sensing devices and ground-based telescopes have been used to observe the chemical composition of Jupiter’s troposphere. In addition to H$_2$ and He, the Jovian troposphere is observed to contain CH$_4$, NH$_3$, PH$_3$, H$_2$O, CO, GeH$_4$, AsH$_3$, and some isotopes of the above species [2, 27]. Water is present but condenses deep in Jupiter’s cold troposphere, thereby becoming incapable of participating in gas-phase atmospheric chemistry. Ammonia (NH$_3$) is more volatile than water, and although it also condenses on Jupiter, NH$_3$ vapor survives to high enough altitudes to interact with ultraviolet radiation from the sun. Ultraviolet photons with sufficient energy can break apart an atmospheric molecule. The photolysis products can react with themselves or other atmospheric constituents. This whole process, termed “photochemistry”, drastically impacts the composition of the Jovian atmosphere.

There have been several models that use the above observations to predict the extent of photochemical processing on Jupiter. In 1973, Strobel [25] was the first to publish a realistic model for the photochemical recycling of NH$_3$ on Jupiter and to identify hydrazine (N$_2$H$_4$) as a major photochemical product. Although hydrazine condenses in Jupiter’s troposphere, its vapor abundance should be high enough that its photolysis can lead to the production of significant amounts of molecular nitrogen (N$_2$) [1]. Further modeling by Kaye and Strobel [17,18,19] suggests that ammonia photolysis products could interact with hydrocarbons, such as CH$_4$ or C$_2$H$_2$, to form organo-nitrogen molecules including hydrogen cyanide (HCN) or methylamine (CH$_3$NH$_2$). Ammonia photolysis also significantly affects the vertical profile of phosphine (PH$_3$).

The formation of hydrogen cyanide has been of particular interest since it has been hypothesized that the coloration seen in the atmosphere of Jupiter may be due to polymers of HCN and its photoproducts. The initial discovery by Tokunaga et al. [26] of HCN bands in infrared spectra of Jupiter sparked considerable interest in the photochemical production of organo-nitrogen compounds. Further investigation revealed that such compounds do not form readily in a hydrogen-dominated atmosphere. Laboratory experiments involving ultraviolet (UV) irradiation of CH$_4$-NH$_3$-H$_2$ mixtures show little interaction between nitrogen and carbon bearing species [12,24]. Photolysis of ammonia in the presence of unsaturated hydrocarbons, such as C$_2$H$_2$ or CH$_3$C$_2$H, may be a more promising way to form HCN and other organo-nitrogen molecules [9,10,11,18]. Recent observations suggest that HCN may not be as abundant as previously believed [4,6], and the original report may have been erroneous.

To determine the extent organo-nitrogen compound formation and the details of coupled ammonia-phosphine photochemistry on Jupiter, we have developed a one-dimensional model of photochemistry and vertical transport in the Jovian atmosphere. For the first time ever we include the C$_2$H$_2$-NH$_3$ schemes proposed by Ferris and Ishikara [10,11] in a photochemical model. Our model is also unique in that we consider stratospheric and tropospheric chemistry simultaneously. Infrared and ultraviolet observations of hydrocarbons, ammonia, and phosphine are used to constrain free parameters, such as the eddy diffusion coefficient or reaction rate coefficients. Our results are compared with previous models and with observed upper limits for HCN and HCP. The sensitivity of ammonia and phosphine to the eddy diffusion coefficient is discussed, as are implications for chromophores on Jupiter and implications for photochemistry on Saturn.

Model: Our one-dimensional steady-state model for the stratosphere and troposphere of Jupiter accounts for hydrogen, carbon, nitrogen, oxygen, and phosphorus chemistry, vertical eddy diffusion, molecular diffusion, condensation, and radiative transfer (including multiple Rayleigh scattering). Using the Caltech/JPL chemical kinetics and diffusion code [14], we solve the coupled one-dimensional continuity equations as a function of time and altitude. This partial differential equation is evaluated by finite-difference techniques, using Newton’s method to solve nonlinear chemistry until a steady state is reached.
The calculations are applied to the stratosphere and troposphere of Jupiter at ten degrees latitude, not including 5-micron hot spots, during average to low solar conditions. The vertical profile is divided into 111 levels for calculation purposes with the bottom level at a pressure of 6.7 bar and temperature of 300.0 K and a top level at a pressure of $1.0 \times 10^7$ mbar and temperature of 930.9 K. There are 120 species accounted for in the code, including condensed species. A total of 193 photolysis reactions and 600 other reactions are considered (including condensation reactions). The reaction list of Kaye and Strobel [17, 18, 19] was the starting point for developing a list of reactions for the model. The list was modified by adding organo-nitrogen reactions from Ferris and Ishikawa [10, 11] and Keane et al. [20], organo-phosphorus reactions from Guillemin et al. [16], and using updated reaction rates. The updated rate constants were largely obtained from information in the extensive databases of Mallard et al. [23].

Once the reaction list and several other input files, including lists of boundary conditions for all molecules, were developed, the kinetics code was run with all species varying for the following three cases: a baseline model with a tropospheric eddy diffusion coefficient of $\sim 10^4$ cm$^2$s$^{-1}$ and with acetylene produced from stratospheric photochemistry, a high acetylene model with an increased C$_2$H$_2$ mole fraction of $10^{-7}$ at the base of our model, and a high eddy diffusion coefficient model with an increased tropospheric value of $10^3$ cm$^2$s$^{-1}$. The results of the three models were compared with observational data.

**Results:** There are several important production and loss reactions for ammonia and phosphine photochemical products in the troposphere. Ammonia vapor is mostly destroyed through condensation and photolysis to NH$_3$ and H. The H attacks phosphine to form PH$_2$ and H$_2$. The NH$_3$ can then react with PH$_2$ to reform ammonia or the NH$_2$ can react with itself to form hydrazine. Thus, it is apparent that ammonia and phosphine photochemistry in the troposphere are interdependent and predominantly separate from hydrocarbon chemistry (as was first determined by Strobel [25] and by Kaye and Strobel [19]). As mentioned earlier the formation of hydrogen cyanide requires reaction of ammonia photolysis products in the presence of unsaturated hydrocarbons. The model considers numerous schemes to form HCN [see 9,10,11,12]; however, we find that the maximum hydrogen cyanide mole fraction is only approximately $5 \times 10^{-12}$ at 300 mbar for the baseline model. This value is well within the upper limit of $10^{-7}$ HCN mole fraction established by Davis et al. [6]. The HCN mole fraction for the high-acetylene model goes above the upper limit prescribed by Davis et al. [6], suggesting that C$_2$H$_2$ is not abundant in the Jovian troposphere.

Figures 1 and 2 show respectively the ammonia and phosphine mole fraction. The three model cases are plotted along with observed mole fraction values from ultraviolet (UV), 5-micron infrared (IR), and 10-micron IR. Ultraviolet data are taken from [7,8] and infrared from [5,13,15,21,22]. The observations from 10-micron IR are most relevant to our non-hot-spot model. The UV data are not as diagnostic since many molecules (such as PH$_3$) have broad and featureless absorption bands in the ultraviolet. This fact combined with the presence of aerosol extinction often makes unique identification of molecules from UV spectra difficult. The 5-micron IR is not entirely applicable since it penetrates further into the atmosphere, thereby being influenced by cloud-free regions of hot spots, which are not considered by this model. Therefore, we compare our results predominantly with 10-micron observations.

Figure 1 displays a $3 \times 10^{-4}$ ammonia mole fraction in the 1 to 2 bar region. This value is in agreement with Fouchet et al. [13] and Davis et al. [6]. The mole fraction then closely follows the saturation vapor pressure line until 100 mbar where it continues to decrease with decreasing pressure. The three models do not show much variance.

![Figure 1](image1.png)

**Figure 1:** Model results of ammonia mole fraction vs. pressure for baseline model (dotted line), high-acetylene model (dashed line), and high-eddy-diffusion-coefficient model (solid line) plotted with ammonia saturation vapor pressure (dash-dotted line) and observational data.

Figure 2 displays the mole fraction profile of phosphine for the three models. The baseline model and high-acetylene model are in close agreement. Increasing the eddy diffusion coefficient causes a large increase in the phosphine mole fraction while it has little effect on the ammonia mole fraction. There is a correspondence between the observed Edgington et al. [7,8] values and the increased eddy diffusion coefficient model, but the Edgington et al. data points were was obtained from UV observations that are not totally conclusive. The baseline model seems consistent with the 10-um phosphine observations.
Hydrocarbon chemistry occurs predominantly in the stratosphere of Jupiter, while ammonia photolysis occurs in the troposphere. The physical separation of ammonia and methane photolysis regions and the low abundance of unsaturated hydrocarbons in the troposphere result in minimal carbon-nitrogen reactions on Jupiter. Some hydrogen cyanide and complex organo-nitrogen compounds are formed, but not in observable quantities. For our standard model, hydrogen cyanide is within the observable limits.

There have been controversial reports of acetylene (C2H2) in the troposphere of Jupiter [3]. We have tested this possibility by including one model with a low acetylene concentration and one with a higher acetylene concentration proposed by Betremieux and Yelle [3]. Betremieux and Yelle suggest an acetylene mole fraction of 1.5x10^-7 in the troposphere. They propose that the abundance of acetylene in the upper troposphere is most likely created through lightning-induced chemistry deep in the troposphere. However, this large C2H2 mixing ratio violates infrared observations [B. Bézard, personal communication 1999].

Also, as seen by our high-acetylene model, a high tropospheric abundance of C2H2 causes organo-nitrogen reactions to become more important and the mole fraction of hydrogen cyanide to increase above the mole fraction limit of 10^-9 imposed by Davis et al. [6]. Therefore, it is unlikely that acetylene is produced in the troposphere.

The modeling presented here also allows us to constrain the eddy diffusion coefficient. The ammonia profile does not change significantly with the different eddy diffusion profiles. On the other hand, the phosphine abundance is very sensitive to assumptions about the tropospheric eddy diffusion coefficient. The phosphine profile with an eddy diffusion coefficient of ~10^3 cm^2 s^-1 is consistent with infrared observations in the upper troposphere. The phosphine curve with a higher eddy diffusion of 10^5 cm^2 s^-1 is consistent with ultraviolet observations. Therefore, the eddy diffusion coefficient is in the range of 10^4-10^5 cm^2 s^-1.

These models also allow us to draw conclusions regarding chromophores on Jupiter and implications for Saturn’s photochemistry. The model results show that organic molecules in the troposphere of Jupiter are sparse. Coloring agents for the clouds are therefore most likely due to inorganic molecules.

The results for running a similar code for Saturn would be quite different. In our Jupiter model, phosphine is eliminated mostly by reacting with ammonia-derived products. On Saturn ammonia condenses further down in the atmosphere and is less abundant at lower pressures. Thus, phosphine photolysis would initiate the photochemistry in Saturn’s troposphere [15].

References:
INTRODUCTION

This preliminary investigation is to determine how abiotic versus biotic processes affect the weathering of olivine. Two experiments were performed to weather small (<2 mm) crystals of olivine, from San Carlos (in central Arizona). The abiotic experiment involves exposing unweathered olivine grains to hot sterile water and the biotic experiment involves exposing sterile unweathered grains to natural ground water with biota known to affect Columbia River Basalts (in Washington). Perhaps the differences between biologically and inorganically weathered surface textures could be used as biosignatures for future application to extraterrestrial materials such as Martian meteorites [6].

MATERIALS AND METHODS

The samples used were olivine crystals from Peridot Mesa located on the Nation of the San Carlos Apache Reservation in central Arizona [6]. Peridot Mesa is a remnant of a xenolith, which was rapidly extruded material from the lower mantle. San Carlos olivine was used for all the experiments in this survey because of its high quality and consistency of composition. In this study the olivine crystals were crushed, sorted to <2 mm in size and ultrasonically cleaned in distilled water and ethanol. All control and experiment grains were documented by a JEOL 6340F field emission gun scanning electron microscope (FE-SEM) equipped with an IXRF energy-dispersive X-ray spectrometry (EDS) system.

To determine the effects on surface textures from inorganic water reactions, the olivine was heated in double-distilled sterile water at temperatures of 200°C and 98°C, with samples heated at each temperature for 24 and 48 hours. The trials at 98°C contained 6 grains each and were placed in a 1 L glass beaker filled with double-distilled-water, partially covered and placed on a hot plate. For trials at 200°C, the samples were placed in a closed system apparatus known as a “bomb”. For each trial 15.1 ml of triple-distilled-water and 3 grains <2 mm were placed in an inner container composed of Teflon.

For the biological experiments a flow through device was loaded with San Carlos olivine, and sterilized (Fig. 2). This device was attached to a ground water well head to collect the natural subsurface microbes, which live in the aquifer of Columbia River Basalts in Washington. The olivines were placed in enrichment mediums for both aerobic and sulfate reducing bacteria (SRB) weathering. After removal of the olivine from the media at regular intervals, the organic matter was fixed with 1% glutaraldehyde. Iron and magnesium were not added to any of the media. The samples were removed from the groundwater media after 24 and 48 hours. All samples were documented with a FE-SEM and EDS for analysis. A control sample, exposed to ground water for 24 hours with no enrichment media added, was also examined.

RESULTS

FE-SEM/EDS images documented pristine olivine grains before the weathered experiment. The 200°C for 48 hours sample in the “bomb” apparatus had precipitated a rust-colored rind. Qualitative analysis by EDS revealed the rind was composed of F and Mg. Small white crystals, >0.5 to <10 µm. that were also mixed within the rind, were composed of Ag and Cl. The EDS confirmed that AgCl was derived from the small amount of silver paste used to secure the grains to the SEM mounts. The Teflon sample holder was the source for the fluorine. Surface textures underneath the rind consisted of two types, “flute”-shaped scour pits and triangular fins (Fig. 1). Both features, when present together, were always perpendicular to each other. Fin structures were associated with a higher relative abundance of Fe than flute structures. Flute structures had a slightly higher relative abundance of Mg than the fin structures. In spite of the contamination rind, fins and flutes, the composition of the olivine remained consistent. Denticulate margins were also present on the sides of the grain samples at 200°C for 48 hours [5].

The inorganic experiments done at 98°C, which were not contaminated, displayed rough and bladed surface structures (Fig. 3). The blades and rough lobes had a higher relative abundance of Fe than the surrounding olivine material.

For naturally weathered samples, olivine grains, which had slightly cloudy surfaces were chosen. The edges of some of the grains were worn and exhibited etched-pits along or near fractures and edges (Fig. 3). Irregular etching on the scale of hundreds to tens of nanometers (nanoetching) was observed along ridges and rough surfaces, often near where organic...
material was found. Nanoetching was not observed in any of the experimental inorganically weathered samples. The organic material ranged in size 0.5-3µm with the largest and isolated organism at 90µm in length. The morphology the material was thick rods, sinuous filaments, transparent films and white-globs with increased relative abundance of Cu. At the bases between substrate and the organic matter was occasionally Ca.

Discussion
The process of weathering silicates by biologic versus inorganic processes is not well understood. The inherent problem investigating weathering are the time scale and chemistry of fluids in nature versus control experiments. The hot water experiments were to alter the surface texture and mimic weathering conditions under laminar to no flow. The elevated temperature of the water was meant to speed up the time scale of alteration by aqueous fluids in nature. In this upper temperature range, the olivine would lose Mg, Fe and Si into solution. Losing cations seems to produce distinct morphologies, such as etching, dissolution, pit scours, denticular margins and fin-shapes on the mineral surface. The contamination trials at 200°C also confirmed this by displaying the fin-shapes that were produced when Fe was pulled into solution and flute scours were produced when Mg was diffused and reacted with the F to precipitate the rind [3]. In the 98°C samples, the extremely rough texture areas began to chemically alter and exhibited increased relative abundance of Fe, some Al and Ca. This likely suggests the alteration into clays. The Al and Ca layers come from minor or traces in the olivine.

The biologic experiment was designed to see the effects of biota in nature on surface textures [6]. Although many of those samples are currently in study, the preliminary observations suggest that organisms do utilize whatever they need from their surrounding environment. For example the growth media were not enriched with Fe or Mg, but Fe is essential for the growth of bacteria [4]. It is possible that the bacteria extracting iron from the olivine produced the nanoetching. However, the nanoetching was only observed on the T=0 + bacteria control sample, which was given no enrichments of any kind. The aerobic and SRB samples did not differ much from each other, but differed from the T=0 sample by lacking nanoetching and having a light coating of carbon and copper in localized areas of rough surface textures. All the organic material was always associated with rough surfaces. Perhaps the rough surface texture was preferred by organisms to readily attach themselves to a substrate or be near areas where Fe and or Mg sites are more easily available.

Summary
Controlled weathering experiments can never accurately recreate natural weathering features, but they can offer insight into basic surface textural changes under controlled circumstances. Biota in nature and the chemistry of water vary so much, that to determine if the surface textures are created by the biota or the water is very difficult. This survey of how biota affect olivine is preliminary. It is necessary to continue observations of samples from the Columbia River Basalt ground water, and determine the species of organisms in the water and confirm their presence. Also, the surface textures after the removal of organic material need to be documented.

For the inorganic experiments, more time and lower temperature ranges need to be tried and documented at the range of textural and compositional changes in the olivine grains.

References

Acknowledgements
Thank you Simon Clemett, Penny Morris-Smith, Gordon Southam, for all your hard work, discussions and insight. Thank you Craig Schwandt and Susan Wentworth for all your good advice and technical SEM expertise. Thank you Todd Stevens for collecting the microbes from the CRB subsurface water in the flow through collector.
Figure 1. Comparative surface textures of the contaminated samples at 200°C, between "flutes" and "fins". The image on the left is a backscatter image displaying the rind (the dark and smaller patches) with the flutes-scour underneath the rinds. The image on right is in SEI displaying the association of the rind with the fin structures.

Figure 2. LEFT: flow-through device constructed by S. Clemett, G. Southam and S. Wentworth to expose olivine grains (in the center valve) to groundwater biota. RIGHT: the storage device for the olivine grains that fits inside the valve (shown left).

Figure 3. Comparative surface textures between naturally weathered, inorganic experiment and biologic experiment. LEFT: naturally weathered dissolution texture, MIDDLE: bladed, rough dissolution texture, RIGHT: crescent-shaped biotic material nestled on rough texture ridge.
REE and Sr Partition Coefficients for Nakhla Pyroxenes
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INTRODUCTION
Nakhla, which is generally believed to have come from Mars, is a medium-grained augite-olivine cumulate [1][2]. Although the composition of its intercumulus melt must preserve important information about petrogenetic processes on Mars, we cannot measure it directly. Various estimates of the major element composition of this melt ranged widely (Fig. I). Our group had used the approach of finding a melt composition that would produce clinopyroxenes of the same major and minor element composition as those found in Nakhla and resulted in determination of a composition that produce pyroxenes that were an excellent match for the ones in Nakhla [3]. This melt composition is shown as NJ4 in Fig. I. In addition to major and minor elements, it is also important determine the abundance of key trace element in the Nakhla melt. In this project, we proposed to experimentally measure pyroxene/melt partition coefficients for Sr and REE that can be confidently used toinvert the composition of the Nakhla pyroxenes. We used a synthetic glass of the revised Nakhla melt composition doped with per cent levels of REE and Sr as starting material for experiments.

EXPERIMENT
Experimental starting materials (in NJ4-724~NJ4-734) were mixture of our synthetic composition (NJ4) and REE oxides. NJ4’s composition is proposed by Kentaro Kaneka [3]. This melt would produce pyroxenes of same composition as those found in Nakhla. A series of experimental charges were doped with 0.5~2.0% REE oxide or SrO, placed on Pt wire loops, fused at 1300°C for 48 hours to be homogenized in gas mixing (CO2/CO) furnaces and then quenched to room temperature. Homogenized charges were placed back in the furnace at a temperature just below the pyroxene liquidus to be crystallized at 1160°C-1167°C. These liquidus temperatures are about 10 degrees lower than that of NJ4 at (1173°C) with no trace elements. Fig.2 shows a partition coefficient pattern for these experiments, where each point represents the average of 10-20 analyses, together with patterns derived from other experiments. The abundance of Sm2O3, CeO2, Eu2O3 and Gd2O3 in pyroxenes seems to have a positive correlation with that of Al2O3 such as those in Fig.3. The other REE oxides and SrO abundance in crystals have no correlation with Al2O3 abundance (Fig.4). Only a charge in NJ4-733 (with Sr and La) have iron zoning (Fig.5) and the other charges don’t have

RESULTS
The charges from a series of experiments contain >99% glass and a small number of euhedral crystals of pyroxenes which are over 50 microns across. The largest crystals are over 1500 microns. These charges doped with REE and Sr were quenched at 1160°C-1167°C. These liquidus temperatures are about 10 degrees lower than that of NJ4 at (1173°C) with no trace elements. Fig.2 shows a partition coefficient pattern for these experiments, where each point represents the average of 10-20 analyses, together with patterns derived from other experiments. The abundance of Sm2O3, CeO2, Eu2O3 and Gd2O3 in pyroxenes seems to have a positive correlation with that of Al2O3 such as those in Fig.3. The other REE oxides and SrO abundance in crystals have no correlation with Al2O3 abundance (Fig.4). Only a charge in NJ4-733 (with Sr and La) have iron zoning (Fig.5) and the other charges don’t have

Fig. 2. REE and Sr pyroxene/liquid partition coefficients. Shergotty values are extrapolated from values observed for less calcic pyroxenes [9]. Values for low-Al Nakhla melts are from [6][8]. Values for NJ2, "basaltic" Nakhla parent melt composition, are from [7]. Values for NJ4 are new data. REE partition coefficients for this "high-fidelity" Nakhla melt composition are intermediate between those for low-Al and high-Al melts as we would expect based on the intermediate Al composition of this melt.

Fig. 3. REE vs. Al2O3 in pyroxene. The abundance of Sm2O3, CeO2, Eu2O3, and Gd2O3 in pyroxenes seems to have a positive correlation with that of Al2O3.

Fig. 4. REE vs. Al2O3 in pyroxene. The other REE oxides and SrO abundance in crystals have no correlation with Al2O3 abundance.

are higher than those from the "ultramafic" Nakhla starting compositions like NT, NL and NJ and lower than the Shergotty partition coefficients and those from the NJ2 starting compositions. This new result is what we expect because we noticed earlier that Al content in melt is largely responsible for the difference in partition coefficient values [6][7][8]. And Al content in melt from NJ4 stating compositions is between that from Al-poor Nakhla starting compositions (NT, NL and NJ) and that from Shergotty or NJ2 starting compositions (Fig. 1). Note that data are lacking for elements between Gd and Yb. We drew in smooth curves but have no actual knowledge of how far up those curves might really go. We are not quite sure why some of the trace elements like Sm, Ce, Eu and Gd have positive correlation with Al2O3 abundance but other REE seem to not have such a correlation. We believe that all of trace elements should have positive correlation with Al2O3 abundance but we suspect that analytical uncertainty obscures this correlation for some elements, especially those in low abundance. That the range of Al2O3 abundance is small might also contribute. We are also not sure why in only one experiment (NJ4-733) synthetic pyroxenes have iron zoning and in the other experiments ones don’t have correlation with Wollastonite abundance.

DISCUSSION

These pyroxene/melt partition coefficients for Sr and REE must be confidently used to invert the composition of the Nakhla pyroxenes because NJ4 composition could produce pyroxenes that were an excellent match for the ones in Nakhla. It is apparent from Fig. 2 that REE partition coefficients from the NJ4 starting compositions

Content. This is also probably a result of analytical uncertainty because La pyroxene/melt partition coefficient is much lower than the others so La abundance in pyroxenes are quite low. Although such a strong positive correlation can't be seen in the other experiments, if in those experiments zoning of Fe, Mg or Ca had existed, trace elements should have strong correlation with Wollastonite content.

![Image](image1.png)

Fig. 5. Backscattered electron images of our synthetic pyroxenes doped with Sr and La. These pyroxenes have iron zoning. The bright areas are high-Fe pyroxenes, whereas the darker areas are low-Fe pyroxenes.

![Image](image2.png)

Fig. 6. Sr vs. Wollastonite in pyroxene. Sr abundance in pyroxenes with iron zoning has a strong positive correlation with Wollastonite content.

CONCLUSIONS
(1) REE and Sr partition coefficients were measured between pyroxenes and melts whose compositions of major and minor elements were same as those in Nakhla. Therefore these partition coefficients can be used to invert the composition of the Nakhla pyroxenes. (2) REE partition coefficients or this “high-fidelity” Nakhla melt composition are intermediate between those for low-Al and high-Al melts as we would expect based on the intermediate Al composition of this melt. (3) Some of REE had positive correlation with Al₂O₃ abundance but the other trace elements do not seem to have such a correlation. However, all the trace elements should essentially have a positive correlation with Al content. We believe that this was caused by analytical certainty. (4) In one of the experiments (NJ4-733), pyroxenes had iron zoning and in these crystals Sr had a strong correlation with Wollastonite content. This is consistent with the result proposed by G. Mckay. Although also because of analytical certainty such a correlation is hidden, each trace element should have one.

ACKNOWLEDGEMENT
I would thank Gordon Mckay for giving me the opportunity to work on this project and his guidance. Craig Schwandt helped keep equipment in the experimental petrology laboratory functioning smoothly. I am also grateful to Loan Le for her supporting all of the experiments and analyses.

REFERENCE
Gravity Studies of Coronae on Venus  
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Introduction

Despite having a similar size and composition, the surface geology of Venus differs greatly from that of Earth. Coronae are unique to Venus and are defined by a circular to elliptical annulus of fractures ranging in size from 60 to 2600 km across and 10 to 150 km wide. Coronae contain volcanic cones, shields and flows and form rises, plateaus, circular rims around depressions or more complex shapes [1,2].

The apparent topographic evolution and associated fracture patterns of coronae are consistent with formation by hot, rising mantle diapirs. As the diapir moves towards the surface, it creates a topographic high with radial extensional faults, creating corona-like features called novae. Close to the surface the diapir flattens out into an ellipsoid, which creates a plateau. When the diapir cools, the surface relaxes into a rim surrounding a depression [3,4]. Alternatively, some have argued that delamination and sinking of cool crustal material may contribute to corona formation. Previous gravity studies of coronae have considered only crust and lithosphere properties [5,6].

The highest concentration of coronae occurs between Beta, Atla, and Themis Regiones (BAT region) [1]. Hecate and Parga Chasmata connect Atla Regio to Beta and Themis Regiones, respectively. Geologic mapping has determined differing histories for three chains within this larger area. In the south, nine coronae on the western end of Parga Chasma appear to progressively age outwards along a three-armed rift. However, the twelve relatively old coronae in a chain stretched between Beta Regio and Kawelu Planitia appear have formed simultaneously. Situated between these chains, forty-three coronae along Hecate Chasma do not have a systematic age progression [7,8]. In this study, we use spherical harmonic representations of gravity and topography [9,10] to take a more detailed look at the subsurface structure associated with coronae in these regions.

Admittance Models

One way to use gravity and topography data to constrain the subsurface is to plot the ratio of gravity to topography as a function of wavelength (admittance). Admittance spectra for localized regions were calculated using the method of Simons et al. [11]. The model admittances were calculated using a top-loading elastic flexure model [12]. The crustal and elastic thicknesses were varied to find both the best fit and tradeoffs for the short wavelength end of the admittance spectrum in each of the three regions. These best fits were used to constrain the two-layer inversion model.

As shown in Figure 1, the admittance for the three curves differ, but roughly match top-loading flexure models at the large harmonic degrees. At Hecate, the observed admittance is consistent with top-loading for crustal and elastic thicknesses of 40 km beyond degree 40. Admittance models partially reproduce the steep rise in Parga’s admittance using reasonable elastic and crustal thicknesses. The curve centered on the chain along Kawelu is even more complex, with two different maxima. The curve at higher degrees fits a crustal thickness of 35 to 45 km, and elastic thickness from 10 to 20 km. The accepted value of crustal thickness for plains on Venus is around 30 km [13].

This discrepancy of crustal thicknesses and the complexity of the admittance spectra at low harmonic degrees indicates that the admittance curves do not describe the full story. However, they do provide constraints for further models.

Two-Layer Inversion Model

To explore the mantle structure beneath the coronae, we used a two-layer inversion model [14,15,16]. The shallow layer represents variations in crustal thickness, supported by a combination of Airy isostasy and elastic flexure, and creates gravity highs in elevated topography. The deep layer of the model represents mantle thermal anomalies, which are assumed to be vertically distributed through the mantle. Buoyant flow, driven by these thermal anomalies, contributes to the surface topography. The mantle thermal anomalies also contribute to the observed gravity anomalies both because of their effect on the mantle density and because of the surface topography that they produce. The relative strength of the two layers was adjusted to exactly match the observed gravity and topography for each term in the spherical harmonic expansion. The inversions are complete to harmonic degree 40, corresponding
to a half-wavelength resolution of 475 km. These inversions are calculated globally and mapped for the three regions of coronae of interest.

In Parga, two thermal anomalies are centered on Javine and Atete Coronae, instead of similarly sized coronae that are mapped as younger. The thermal map of Kawelu agrees with the assignment of old age to coronae here since all but one is over low temperature areas. However, Rauni Corona, embayed by the surrounding plains material [8], is on the edge of a thermal high, indicating a possible young age. Of the fifteen coronae over three hundred kilometers in diameter along Hecate, nine are over thermal highs. There does not seem to be a correlation between coronae topographic signature and thermal anomalies at Hecate. Also, there does not seem to be an overall structural pattern of thermal anomalies.

Further variations such as including bottom-loading flexure and testing different mantle viscosity models are necessary to explore the full range of the two-layer model. I'd like to acknowledge the patient help of my advisor for this project, Walter Kiefer.

References

Figure 1
Top - The admittance curves for Hecate Chasma, centered at 10 N and 240 E (line with error bars is observed admittance, solid line is a model with crustal thickness of 35 km and elastic thickness of 45 km, dashed line is a model of 40 km elastic and crustal thicknesses. If spherical harmonic degree increases to the left, then wavelength decreases to the left, i.e. spherical harmonic degree 10 corresponds to a wavelength of 3800 km on Venus and spherical harmonic degree 100 corresponds to a wavelength of 380 km.

Middle - The admittance curves for Parga Chasma, centered at 15 S and 250 E (line with error bars is the observed admittance, all models have crustal thickness of 25 km, lower dashed line is a model with an elastic thickness of 15 km, solid line is a model with an elastic thickness of 20 km, and the upper dashed line is a model with an elastic thickness of 25 km).

Bottom - The admittance curves for Kawelu Planitia, centered at 33 N and 257 E (line with error bars is the observed admittance, the solid line is a model with a crustal thickness of 35 km and elastic thickness of 20 km, the dashed line is a model with a crustal thickness of 40 km and an elastic thickness of 15 km).
Figure 1

Hecate Chasma

Parga Chasma

Kawelu Planitia

Spherical Harmonic Degree
Occurrence of Fluid Inclusions in CM2 Chondrites

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Introduction: In the past fluid inclusions have been found in two H5 chondrites, Monahans and Zag, [1] and in one CI chondrite, Orgueil. These fluid inclusions presented the first glimpse into the possibility of obtaining samples of asteroidal fluid, possibly the water that effected the alteration of many of the C chondrites. For the first time the possibility of doing direct work on this fluid, and thereby placing some constraints on the theory surrounding asteroidal alteration processes, presented itself. In addition this water presents the first possibility of tracing the origins of water on earth and in our solar system.

With all this in mind our investigation had three primary goals. First, to locate fluid inclusions in carbonaceous chondrites. Second, to characterise the minerals and lithologies where the fluid inclusions occur and, finally to characterise the fluid inclusions themselves to determine whether they result from terrestrial contamination or to characterize the extraterrestrial fluid from which they originated. The first two goals were reached and represent the majority of the information presented here. The final goal could not be achieved in a satisfactory manner for several reasons yet several lines of reasoning point to a possible answer and indicate where further work is needed.

Samples & Methods: The investigation centered around carbonaceous chondrites which showed aqueous alteration effects but where no fluid inclusions had previously been noted. Twenty-eight thin sections from 16 different meteorites were studied using, primarily, a petrographic microscope. Moving vapour bubbles, concrete evidence of a fluid inclusion, [2] were located using high (100x) magnification.

A Cameca SX-100 microprobe was used to probe fluid inclusion bearing and associated minerals (primarily olivines and carbonates) and to obtain BSE images of the thin sections. The probe was run at 15kV and 40nA for the first scan and 20nA for the second scan.

Results and Discussion: Fluid inclusions were positively identified in 2 new meteorites, both CM2s. In addition the possibility of further inclusions remain for several other meteorites.

The fluid inclusions, whose diameter measured less than 10 micrometers, occurred within several minerals. In LON94101 these minerals were high Mg, low Fe olivine and Ca carbonate. In ALH84029 the fluid inclusions were found exclusively in Ca/Mg carbonates.

The fluid inclusions found occurred in a variety of lithologies, though in the same lithology within each meteorite. In LON94101 the fluid inclusion lithology consisted of a dark matrix composed primarily of indistinguishable phyllosilicates with a few fine-grained phyllosilicates which were distinguishable from the matrix. The latter set of phyllosilicates occurred primarily in altered ameboid olivine inclusions or remnant chondrules. The primary transparent phase was olivine of widely varying texture. Secondary anhedral granular carbonates were also present though they made up less than 5% by volume of the samples. A few rounded grains of metal and even fewer euhedral sulfide grains were seen in one sample. In ALH84029 the relative proportions of olivine and carbonate were almost reversed. It also contained more remnant chondrules and altered ameboid olivine inclusions. The remnant chondrules had largely been altered to phyllosilicates with some carbonate present. In one of the samples carbonate appears to be concentrated in, though not exclusive to, ameboid inclusions and remnant chondrules. The matrix was made up of approximately an equal mixture of fine-grained phyllosilicates, some of which had a variolitic texture, and dark, unidentifiable matter similar to the dark phyllosilicate matrix of LON94101. Other than this both olivine and carbonates are evenly distributed through the lithologies of all the samples.

In addition to samples where aqueous alteration was prominent, where fluid inclusions were expected, we studied several CV3s, both
falls and finds. In the 3 sections of CV3s that were studied no fluid inclusions were found.

The fact that fluid inclusions were found in the same lithology in several samples of the same meteorites yet no samples of other meteorites argues strongly in favour of the fluid being of extraterrestrial origin. Contamination during preparation would affect all meteorites prepared in that fashion in the same way. The extraterrestrial origins of this fluid will be further tested in future work where we propose to perform heating and freezing experiments on the inclusions. In addition laser RAMAN spectroscopy may prove useful in identifying possible contamination. Since we postulate that the fluid present is brine the presence of peaks in the wavelengths associated with either cutting oil or alcohol would indicate contamination. On the other hand a peak associated with brine would be substantial proof of the extraterrestrial origins of this fluid. One last procedure which could prove very useful but which is still in the developmental stages is laser ablation spectroscopy on micrometer sized objects.

Several ideas have been put forward to explain why fluid inclusions would not be present in the most altered meteorites. They all center around the softness of the mineral constituents of those meteorites and their behavior in impacting conditions.

It was found that upon probing, with the microprobe, a large number of the fluid inclusions were destroyed in both olivines and carbonates.

In addition to ALH84029 a meteorite which was thought to be paired with it, ALH83100 was looked at but no fluid inclusions were found.

Conclusions: The conclusions gathered from our results are: 1) Fluid inclusions do exist in carbonaceous chondrites. 2) Of the chondrites studied (CI, CM, CV) fluid inclusions were found mostly in CMs and by extrapolation are most likely to be found in CM’s in the future. 3) Fluid inclusions occur in at least two separate lithologies, yet these two lithologies are similar in the presence of olivines, the lack of metals or sulfides, the presence of ameboid inclusions and remnant chondrules and the presence of at least some secondary carbonates. 4) Fluid inclusions were found in only two CM meteorites both of which had been moderately aqueously altered and not in any that were altered to a greater or lesser degree. There therefore seems to be a narrow range of conditions under which a meteorite will retain trapped fluid. 5) The fact that probing with a microprobe or, presumably any other particle beam and possibly simply any significant heating, decrepitates the inclusions may explain why so few inclusions have been found to date. In particular it may afford a possible explanation as to why only one fluid inclusion has been found in CI’s as these are some of the most extensively probed meteorites. 6) The fact that no fluid inclusions were found in ALH83100 which is assumed to be paired with for ALH84029 may indicate that in fact these two are not paired.

References:

Table 1: Summary of Findings

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Fig. 4: Fluid inclusions in a carbonate grain in ALH8402926.

Fig. 3: Fluid inclusions in a carbonate grain in LON94101-36.
THE RELATIONSHIP BETWEEN HEMATITE AND BACTERIA IN THE GUNFLINT IRON FORMATION FROM ONTARIO, CANADA: IMPLICATIONS FOR THE HEMATITE DEPOSITS ON MARS

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INTRODUCTION
A fundamental requirement for life is water; thus the main theme in the search for life on Mars is linked with water. The recent discovery of what appears to be large hematite deposits on the equatorial surface of Mars [1] has excited great interest because in most cases, hematite can only be formed while in the presence of water. The hematite deposits may therefore be a potential site in the search for past Martian life. With this in view, the present study of the interactions between bacteria and hematite in the Gunflint Iron Formation from Ontario, Canada was undertaken.

The Gunflint is a 2.0 Ga formation that spans the United States/Canada boundary, through Minnesota and Ontario. These rocks are essentially unmetamorphosed [2]. Barghoorn and Tyler [3] made the first observations of Precambrian microbial fossils in this formation. Its diverse microbial population has been thoroughly investigated, primarily by optical microscopy.

The portion of the Gunflint Formation investigated in this study was from an area west of Mink Mountain, about 7.3 km west of Mackies, Ontario (PPRG 336) [2]. The Gunflint is a banded iron formation (BIF) which formed in a restricted, possibly lagoonal shallow water environment. It was deposited during a very unique transitional period in Earth's history when a stable oxic atmosphere was being established [4]. BIFs were common only during the Archean and Proterozoic when the Earth was in transition from an anoxic to an oxic atmosphere.

There are two hypotheses regarding the formation of BIFs. The physiochemical model is as follows: owing to continental weathering and high rates of volcanism, there were high amounts of iron and silica in solution in the ocean waters. Without planktonic organisms in the oceans to draw out silica, the ocean waters were supersaturated with respect to silica and thus there was a continual deposition of silica. Periodically, ferric ions in solution would become oxidized by a build-up of oxygen in surface oceans. The ferrous ion is insoluble and thus precipitated out onto the ocean floors. In the biological model, microbiological activity is presumed responsible for the concentration of the iron into layers. In this model, Fe ions in solution chelate to functional groups such as COH, POH, and OH on the envelopes of the individual cells within the microbial mats coating the shallow sea floors, in this way building up laminated layers of precipitated hematite.

The sample studied from the Gunflint Formation is an iron-mineralized banded iron stromatolite. A stromatolite is a mat-like community of living organisms that forms in shallow water environments. It has been suggested that there may have been hydrothermal activity associated with this deposit [5].

METHODS
A combination of methods was used to investigate the hematite-rich Gunflint sample. Three sections from the sample were analyzed: etched and unetched rock chips, and an etched thin section. Etched samples were held in the fumes of hydrofluoric acid (HF) for 15 and 30 minutes respectively and then thoroughly rinsed with filtered deionized water.

The chips were observed with a binocular microscope prior to preparation for the scanning electron microscope (SEM). An Olympus petrographic microscope with objectives up to 100x was used to analyze the thin section.

A JEOL JSM-6340F Scanning Electron Microscope was used to analyze both the etched and unetched rocks chips and thin section. The EDX analytical system had a detector sensitive to elements as light as carbon.

Microprobe analyses were conducted with a Cameca SX100 microprobe. X-ray diffraction analyses were carried out with a XDS 2000 Scintag Diffractometer.

RESULTS
The etched and unetched rock chips, as well as the thin section, show the characteristic stromatolite texture of wavy laminae. The laminae, as observed in thin section contain dense accumulations of micron scale hematite crystals.

Trapped between hematite-rich laminae are oval peloids, approximately 1 mm in size. The peloids are characterized by concentric banding of hematite and quartz. They have an outer finely laminated layer of hematite crystals. The quartz rich bands in some of the peloids contain 10 µm diameter, slightly irregular spheres, which may represent the degraded remains of the sheaths of colonial cyanobacteria.

Between the peloids runs a chlorite rich band containing filamentous structures approximately 1-2
µm in width and tens of microns in length. Numerous 1 to 2 µm wide filaments with overlapping orientations were observed (figure 1).

The stromatolite is cut by a quartz vein 2 mm thick. Clusters of micron-scale hematite crystals were observed along the vein.

Many fossil structures were observed under SEM, particularly in the etched rock chip. Degraded, hematite-rich, carbonaceous, coccoidal microstructures (1-2 µm in diameter), sometimes occurring in chains, were observed. Some of the carbonaceous filaments were hematite-rich and up to 200 µm in length.

The sample contains numerous 5 to 10 µm euhedral hematite crystals concentrated in the iron-rich layers. The crystals are embedded and coated by mineralized polymeric material, i.e. fossilized polymeric substances that together with bacteria form bacterial biofilm. Some of the large, hexagonal hematite crystals contain concentrated carbon-rich material (1-2 µm in diameter).

Microprobe analyses showed a dominance of hematite and quartz in the sample. X-ray diffraction analysis confirmed the microprobe findings, and in addition indicated a 7-angstrom phyllosilicate, tentatively identified as chlorite.

Most of the fossils were fairly degraded before their fossilization but the polymeric material has been well preserved. The bacteria and polymeric substances were fossilized either by a process of permineralization or hematite mineralization [6]. The observations clearly document the association of polymeric substances and microorganisms, which form the microbial mat layers of the stromatolite.

DISCUSSION

Three relationships were observed within the sample: 1) an intimate association between hematite crystals and polymeric substances (figure 2), 2) hematite-mineralized bacterial remnants (figure 3), and 3) hematite crystals with carbon inclusions (figure 4).

Microfossils observed in the Gunflint sample were badly degraded before their fossilization, and fossilized polymeric substances were found to be ubiquitous in both the rock chips and in thin section.

Numerous scientists have identified many coccoidal and filamentous structures within Gunflint Formation samples. However, hematite mineralization of biogenic structures has not previously been investigated. The bacterial cell envelopes may have attracted and concentrated the iron needed to catalyze the crystallization of hematite.

Large crystals of hematite have been found to contain 1 to 2 µm sized pockets of carbonaceous material within their structures (figure 4). These structures may represent the degraded remains of bacteria trapped during the precipitation of the large hematite crystals. This observation suggests that the formation of hematite may have been microbially mediated. The close association of the hematite with the microbial structures shows that the deposition of hematite was a very early diagenetic occurrence, contemporaneous with the permineralization of the microbial mats [7].

RELEVANCE FOR MARS

The Gunflint Iron Formation sample analyzed in this study is similar to what may be found on Mars, but not identical. The Martian deposits differ from banded iron formations like the Gunflint in that they show no evidence for other minerals, such as silica, along with the hematite [1].

This study has documented the close association between the hematite in the Gunflint and the microfossils observed within it. If bacteria contributed to the formation of the iron layers in the Gunflint sample, it can be hypothesized that if bacteria ever occurred on Mars, they could have served the same purpose.

This study was not able to establish whether there was any hydrothermal influence on these sediments, but hydrothermal activity cannot be excluded. Certainly hydrothermal deposits are to be expected in the early (and present) Martian environment.

CONCLUSION

The use of a multi-method approach including scanning electron microscopy has been invaluable in documenting fossil microbial structures which have not previously been observed. These include observations of the filaments, chains of 2 µm coccoidal cells, the fossilized polymeric substances, and the intimate association between hematite deposition and organic matter.

Three associations between bacteria and hematite were documented in the hematite-rich Gunflint Formation sample from Ontario, Canada: 1) an intimate relationship between polymeric substances and hematite crystals (figure 2), 2) iron-mineralized remnants of bacteria, filaments, and polymeric substances (figure 3), and 3) hematite crystals with carbonaceous material within their structures (figure 4).

The microfossils observed in the Gunflint sample were badly degraded before their fossilization, and fossilized polymeric substances were found to be ubiquitous in both the rock chips and in thin section.

The Gunflint, while not identical to the Martian hematite deposits, may provide clues as to the formation of iron deposits on Mars. The findings of this study will provide a database for a sample return mission to Mars in coming years.
ACKNOWLEDGEMENTS
Thank you to Everett Gibson for the sample, and also J. William Schopf, Craig Schwandt and Sue Wentworth for SEM and microprobe help. Charlie Galindo for XRD help. Dave McKay, Penny Morris Smith, Jan Toporski, and Andrew Steele for help with interpretation. The LPI summer intern program for allowing me to spend my summer researching the Gunflint.

REFERENCES

Figure 1: Back-scattered scanning electron image of hematite mineralized crossing filaments with quartz background. Note the 10 µm scale bar.

Figure 2: Scanning electron microscope image of euhedral hematite crystals covered with fossilized polymeric substances. Note the 1 µm scale bar.

Figure 3: Scanning electron microscope image of a chain of four hematite mineralized coccoidal bacteria. Beneath the chain are two collapsed, hematite mineralized cell rims. Note the 1 µm scale bar.

Figure 4: Back-scattered scanning electron microscope image of irregular hematite crystals with carbonaceous inclusions with a quartz background. Note the 1 µm scale bar.
The Effects of Impact Angles on Martian Craters using MOLA data
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Advisor: Dr. Robert R. Herrick, Lunar and Planetary Institute, Houston, Texas

Introduction

The effects of impact angle on impact crater morphology and topography have been studied extensively with experimental work, but little work has been done comparing that work with observations of planetary impact craters. The following study compares experimental work by Gault and Wedekind [1] with the topography and morphology of Martian craters. Gault and Wedekind [1] found that no significant elongation in craters occurred for impact angles > 15°. In reference to the ejecta blanket, they found no distinct difference in the blanket's axial symmetry until an angle < 45°. As the impact angle decreases with respect to horizontal, a forbidden zone appears uprange and then spreads to downrange. A butterfly wing ejecta pattern occurs from near grazing impacts.

The purpose of this study is to compare these experimental results to topography and images of Martian craters. From there comparisons can be made with a study from Bottke et al. [2] which uses both topographic data and the lab experiments, a study such as Garvin et al. [3] using topographic data and morphology studies such as Strom et al. [4].

Methods

Two different sources were used in obtaining the data needed for analyzing the craters for the purpose of this study. The imagery data was provided by the Viking Orbiter at a resolution of 233 m [5]. Another instrument, the Mars Orbiter Laser Altimeter (MOLA) via the Mars Global Surveyor provided the necessary topography data at a vertical resolution of approximately 30 cm and along-track spatial resolution of 300-400 m [6]. The image data was overlain by the topography data giving detailed profiles through many features of the selected craters.

Initially, craters were chosen by lack of erosion to the crater and ejecta, ejecta blanket morphology, crater circularity, and size. The presence or lack of ejecta in two distinct regions, uprange and downrange, and two obvious morphologies, perpendicularly symmetrical and butterfly ejecta patterns, were used as an initial classification method. Profiles in various directions from 53 different craters were then used to reduce the data set to craters that were well observed in the MOLA data. Effort was taken to ensure the range of size and ejecta morphologies would be preserved. The profiles of the remaining 43 craters containing characteristics such as ejecta extent and thickness, crossrange, downrange, and uprange rim height, downrange and uprange crater wall slope, crater-floor morphology, and depth comparisons were used in the final comparison of Martian oblique crater morphology to laboratory experiments performed by Gault and Wedekind [1].

Results

Differences in Ejecta Patterns

As distinct choices, both symmetrical and asymmetrical craters/ejecta patterns were chosen as two of the groups. The other two classifications of ejecta were determined by the surplus or absence in the uprange and downrange areas of the ejecta blankets. These four categories are shown in Figures 1-4 with respective profiles. These four categories will further be referred to as higher angle (round ejecta), mid-angle (surplus downrange or missing uprange), low angle (missing up-and downrange), and grazing (butterfly ejecta and asymmetrical craters.)

Higher angle ejecta patterns can further be described as having a symmetrical ejecta blanket completely around the crater. Higher angle craters can also be easily identified by similar uprange/downrange and crossrange profiles of the crater also.

Mid-angle craters typically either have missing uprange ejecta or excess downrange ejecta blankets. In these craters a slight higher ratio for the crossrange rim height to uprange/downrange rim heights of the crater can be expected.

Low angle craters can be typically be made distinct from mid-angle craters by the increase in lack of ejecta in both the up- and downrange directions in addition to a higher ratio between the crossrange and up/downrange rim heights.

Grazing craters are distinctly different from low angle craters by the presence of elliptical craters and distinct floor morphology patterns when present. The morphology difference is evident in the two images of figure 2 and 4. The difference in crossrange and up/downrange profiles is distinctly evident in figure 7.
Another distinctive quality in ejecta blankets is the presence of rays. Rayed craters have a generally thin ejecta thickness and tend to demonstrate a lower impact angle as evident by their ejecta shape.

An interesting characteristic of all the profiles shown (figures 5-7), is the similar terrain elevation to crater depth ratio. This suggests that there is no relationship between crater diameter, crater morphology, and crater depth. Furthermore, the profile data suggests a similar ratio for all four size distinctions. For the smallest craters the ratio was \(0.056\), the 18-24 km range depth/diameter ratio was \(0.037\), the 26-30 km ratio was \(0.037\), and the 32-39 km ratio was \(0.029\). These numbers possibly suggest a small inverse size to depth ratio but more sampling is needed.

**Differences in Sizes**

Other distinct differences are seen in crater morphology with varying sizes. This study consists of craters from 8-39 km in diameter and were broken up into two different categories - craters with no floor morphology and with a central peak. The distinction is between a lack in central peak and the presence of a central peak. This boundary has been established by Melosh [7] to be between 10-20 km. Craters were found to contain peaks with diameters as low as 8 km and were not found in some craters as large as 33.9 km.

The craters were divided into four groups - 8-14 km, 18-24 km, 26-30 km, and 32-39 km. This distinction allowed comparison between each of the four categories of ejecta blankets.

When comparing ejecta thickness to diameter, there was no obvious correlation. Three of the size groups, 8-15, 18-24, 32-39, all have the same average for ejecta depth while the fourth group, 26-30, has an average quite high. This suggests no correlation.

On the other hand, the ejecta extent shows indication of a possible correlation. For Example, the extent amounts for sampled craters between 8-15 km ranges from 19,000 to 45,250 m while the measurements of the 18-24 km groups range from 41,750-65,750 m.

Another variable that does not appear to be related to size is peak heighth. The values span the range for all diameters.

**Conclusions**

Overall, the general morphology and topography of Martian craters corresponds well with what is expected from the laboratory experiments. In particular, with decreasing impact angle the uprange rim first becomes depressed and for lower angles the downrange rim also becomes depressed relative to the crossrange direction. The depth-diameter ratio does not appear significantly different for oblique versus near-vertical impacts. The ejecta extent-diameter ratio suggests a correlation but the ejecta thickness-diameter ratio does not.

There are few indications of slight difference between true planetary and lab conditions. These could be eliminated or further studied with more sampling.

**References**


**Acknowledgements**

I would like to thank Dr. Robert Herrick for all of his support and confidence in me during the course of this study. I would also like to thank all of the computer staff at the LPI for helping me in my times of panic.
Figure 1. (top left) A crater image showing evidence of probable high angle impact. The ejecta is symmetrical in all directions and the rim heights are all roughly equivalent (see Fig. 5 [top right]).

Figure 2. (second down on left) An image of a crater exhibiting a lack of uprange ejecta.

Figure 3. (second from bottom on left) A crater image showing an ejecta pattern approaching a butterfly shape. As evident in Figure 6 (its profile counterpart) the downrange rim heights is a little lower than the rest and the slope is more gradual than the steeper uprange.

Figure 4. (bottom left) An image of a crater demonstrating definite low angle impact. This is evident in its profiles (Figure 7) and by its asymmetrical shape and distinct difference in rim heights.

Figure 5. (upper right) Profiles extending from north to south and west to east for the crater in Figure 1.

Figure 6. (second from bottom on right) Profiles extending in the crossrange and north to south directions for the crater in Figure 3.

Figure 7. (bottom right) Profiles extending in the crossrange and downrange directions for the crater in Figure 4.
Surface Composition of Mars: Examination of Syrtis Major Using Thermal Infrared Data

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Introduction. An important goal of Martian infrared studies is to determine surface mineralogy, enabling scientists to locate deposits of interesting minerals (e.g., carbonates, sulfates) for landing site selection. Currently, the Global Surveyor Thermal Emission Spectrometer (TES, 6-50 µm) is returning infrared data from Mars that can be used to search for such minerals.

Syrtis Major is one of the prominent low-albedo regions of Mars, and is believed to be a low relief shield volcano [e.g. 1]. However, it has an unexpectedly moderate thermal inertia, implying the presence of coarse-grained material [2,3]. It was chosen because of its reasonably low relief and the likelihood of a relatively strong spectral signature commonly associated with low-albedo TES data ([4] and our findings).

Christensen et al. [4] have found calcium-rich pyroxene to be the best match to TES spectra of the low-albedo region, Cimmeria. Mustard et al. [5,6] assert that both low and high calcium pyroxenes compose the dark region of Syrtis Major. Bandfield et al. [7] cite two likely compositions for the low-albedo regions of Mars: (1) a clinopyroxene and plagioclase feldspar dominated basalt and (2) a volcanic glass and plagioclase feldspar dominated andesite. They propose that NW Syrtis Major has an andesitic composition, while the majority of Syrtis is in the basaltic regime [7]. Many believe that the dark regions are composed of more coarse particles than the bright regions [e.g. 8,9], with Syrtis in particular containing sand-sized particles [3]. The purpose of this project is to test the assertions described above.

Background. Rocks emit as blackbodies, except where there are molecular absorptions or scattering effects. Absorptions occur at characteristic wavelengths due to variations in the chemistry and configuration of the elements present. These factors dictate the regions that do not exhibit blackbody behavior. The signal received will thus be weakened and show absorption bands in characteristic regions. Even minerals composed of similar elements are distinctive due to the various bending and stretching vibrations of their constituents. Silicates, for example, have intense absorption bands that occur between 8 and 12 µm, which are the result of asymmetric Si-O-Si stretching vibrations [10]. Figure 1 shows a typical Mars signature plotted with a Planck curve, which exhibits characteristic atmospheric and surface absorptions.

Before infrared radiation is detected by TES, it passes through a layer of atmospheric dust and gases. The radiance measured will have absorption features caused by the atmosphere as well as the mineral(s) that characterize the source region. Thus the spectra contain features of this aerosol, which must be removed in order to view the true surface spectra. With some reasonable simplifications [11], the measured radiance \( L_{\text{obs}} \) can therefore be defined as:

\[
L_{\text{obs}} = \tau L_s + L_{\text{up}} \tag{1}
\]

where \( \tau \) is the atmospheric transmission, \( L_s \) is the radiance of the source, and \( L_{\text{up}} \) is the re-emission of the dust, which exists because the aerosol has a temperature [11]. Solving the above equation for \( L_s \) gives information about the surface.

![Figure 1](image1.png)

Figure 1: Negative x-axis signs are artifacts of the plotting program. Radiance is in watts/m²·sr·cm⁻¹.

Figure 2 shows the calculated atmospheric transmission spectrum. The uncorrected TES spectra most resemble this spectrum, which shows that a strong aerosol signature must be removed to see the surface spectrum. The result is a spectrum of emitted radiation from the surface, which is extremely subtle when compared to the transmission spectrum (Figure 2). For a more detailed explanation of atmospheric removal techniques, see [11].

![Figure 2](image2.png)

Figure 2: The Mars surface spectrum shows the full spectral range measured by TES. The dark shading in the Mars surface curve corresponds to the region looked at in this study. Arrows point to CO₂ absorption features, and double circles indicate water vapor. Double arrows show the aerosol band.

Procedure. The region of Syrtis sampled was located between -10 - 30° latitude and 285 - 320° longitude. This incorporates both the western extent of Syrtis and the eastern portion of Arabia, providing for data comparison.

The data used were acquired by TES, which is a Michelson Interferometer with an unapodized spectral resolution of 10 or 20 cm⁻¹. Most of the data were taken at...
20 cm$^{-1}$, used in this study. The spatial resolution of the instrument is ~3 km [12], with a peak-to-peak signal-to-noise ratio (SNR) of 66 at 10 µm and a 270K blackbody.

TES produced 143 points of radiance data in the wavelength range from 1600 to 200 cm$^{-1}$, which is in the thermal infrared [13]. We used data from orbits 3888-3888 through 4207. After examining all wavelengths, we limited the wavelength range used from ~1400 to 750 cm$^{-1}$, using 61 points. We dropped 750 - 550 cm$^{-1}$ because strong CO$_2$ gas absorptions inhibit viewing of the surface at these wavelengths. The 1971 IRIS mid-day spectra typically show a 20 µm band accompanying the 10 µm band, which was not observed in the TES data. We removed wavenumbers <550 cm$^{-1}$ due to this possible calibration problem. Further, we limited data used to a peak-to-peak signal-to-noise ratio (SNR) above 20 (~100 r.m.s. SNR) at 270 K, which excluded wavenumbers >1400. The peak-to-peak SNR of the TES data was calculated using the method described in [14].

Under Department of Defense-sponsored programs, the Aerospace Corporation developed an In Scene Atmospheric Compensation (ISAC) technique to calculate the atmospheric transmission and re-emission (Equation 1), and we used this technique on the TES data [15].

**Observations.** Figure 3 shows two TES spectra from NE Syrtis with the atmospheric compensation applied. The top spectrum is from a region with an average TES albedo (0.3-2.7µm) of 0.09 (incidence angle (ina)=44°, emission angle (ema)=0.2°), and the bottom spectrum is from a 0.10 albedo region (ina=45°, ema=0.2°). A better SNR is achieved by averaging several spectra. The overall bowl-shape of these spectra is common in Syrtis's dark regions, with emissivity peaks occurring near 1300 cm$^{-1}$, followed by a flat shape at higher wavenumber (shorter wavelength).

**Figure 3:** The top curve shows the average of 6 Martian spectra taken from TES orbit 3962. The bottom, gray curve shows the average of 235 spectra taken from orbits 3888 to 4207 in a region approximately 470 by 170 km in size and centered at 12°N latitude, 289° longitude. Note the better signal-to-noise of the bottom curve, which is vertically offset by 0.02 for clarity. This and the following plots show only a subset of the TES data (see Figure 2).

Figure 4 shows another typical dark region spectrum (average albedo=0.10, ina=45°, ema=0.2°) from central Syrtis Major plotted with a bright region spectrum (average albedo=0.27, ina=46°, ema=0.2°). Bright regions are observed to be relatively flat in the 9-µm region, while there is a large dip in emissivity at higher wavenumbers and an emissivity trough near 850 cm$^{-1}$.

**Figure 4:** The light gray plot is an average of 746 spectra from orbits 3888 to 4207 in a typical bright region of Syrtis Major, which [7] designate as "Surface Type 1.

We compared our surface spectrum to laboratory spectra of andesite and basalt from [10,16], and the overall broad envelope is consistent with that of basalt or andesite. Figure 6 shows the surface spectrum plotted with basalt.

**Discussion.** We identified three sources of error and uncertainty. (1) SNR: This is straightforward to characterize, and we address it by excluding low SNR data. (2) Systematic radiance errors associated with the instrument calibration [17]: We partly address it by excluding data at wavenumbers <550. (3) Atmospheric compensation: There is no clear method to calculate the uncertainties because we have no independent knowledge of the true atmospheric spectral shape or variation with time. However, sources of uncertainty include (1) the assumption that the atmospheric profile is invariant with latitude and longitude and (2) the assumption that the calculation is fairly insensitive to topography. This was handled by choosing fairly level regions and excluding corrected spectra that exhibit residual atmospheric CO$_2$ bands.

It can be concluded that the signature obtained for the dark region of Syrtis is consistent with the TES team signature interpreted as basaltic [7]. The presence of a clear band indicates that this region is composed of well crystalline material with a particle size at least comparable to the wavelength [10], which means that rocks could exist. A
flattened emissivity is also observed at wavenumbers >1300, which is also consistent with larger particle sizes. An exact determination of particle size would require knowledge of the percentage of dust coverage and the vesicularity of the materials at the surface. These are unknowns.

The bright region spectrum from Arabia in Figure 4 shows an apparent emissivity near 1 from ~1260 to 900 cm⁻¹. Three observations can be made about this bright region spectrum: (1) there is a feature at ~850 cm⁻¹ with a depth of ~0.008 in apparent emissivity (Figure 4), which we interpret as a transparency feature [18]; (2) emissivity decreases sharply toward wavelengths >1260 cm⁻¹ (<7.9 µm); and (3) the region is brighter. All of these observations are consistent with finely particulate material [10, 18].

The TES team used two spectral ranges in their study described in [4,7]: ~1300 - 800 cm⁻¹, and ~500 - 200 cm⁻¹, and we used 1400 - 750 cm⁻¹ for our study. Over the region of overlap, the spectral shapes agree broadly with results from [4, 7] for dark regions. However, the TES team obtained a blackbody spectrum for bright regions [19], which is contrary to our results.

To best understand the results of this study and to accurately determine the composition of Syrtis Major, a landing site in this region would be highly advantageous [20]. Edgett et al. [20] note that it would be favorable to study a region with exposed rocks or large particles, and from our results it seems clear that there are large particles and possibly rocks exposed at the surface of Syrtis. There is no indication of carbonates or sulfates in our study, but if these materials are rough or fine-grained, they would be below the detection limit. This depends on grain size and surface roughness, and therefore "absence of evidence is not necessarily evidence of absence." Again, a sample of dustless rocks in this region would be highly beneficial [20].

One of the most important factors to note as a result of this study is that the atmospheric correction is severely critical: the answer is driven by this correction (note Figure 2). For example, we applied the correction to the Lunae Planum region of Mars, but because the surface signature was so subtle this region had to be abandoned. For Syrtis Major, Figure 5 shows that we have obtained a weaker band than that obtained by the TES team (~ 3% compared to ~5% for these spectra). There are three possible explanations for this discrepancy: (1) the TES team spectra still have residual atmospheric dust, (2) there is some scaling difference between the two data sets, and (3) one of the atmospheric correction techniques needs improvement. It should be realized that the strongest type region signature we found only had a 3% band depth. The importance of the atmospheric compensation is evident.

The main difficulty in making adequate corrections involves the SNR. Also, a higher spectral resolution would increase confidence in the result, but the spatial resolution was not an issue at this level, i.e. in examining the basaltic plains.

In the future, an unambiguous answer for landing site selection is desired. This leads to the need for an intense examination of the methods used to remove the effects of the atmosphere. The answer is only as good as the atmospheric correction.
Phase Relations of a Model Martian Mantle at 5 GPa.
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Advisor: Carl Agee, NASA Johnson Space Center, Houston, TX

Introduction: Studies of the Shergottite-Nakhlite-Chassignite (SNC) meteorites have determined many important constraints on the nature of the Martian mantle. Major element, trace element, and isotopic systematics indicate that the Martian mantle from which the SNC meteorites were derived was previously strongly depleted in rare earth elements and aluminum [1]. These depletions have been attributed to multiple large-scale differentiation events in early Martian history. An alternative hypothesis is that fractionation of garnet in an early Martian magma ocean produced these depletions. This is the first work of a larger study designed to experimentally address potential processes active during the early differentiation of Mars.

Relationships between major-element ratio pairs show that ordinary chondrite compositions like Homestead L5 [2] are good starting points from which the array of Martian compositions (SNC meteorites and Pathfinder rocks) can be generated via garnet fractionation from an early Martian magma ocean. The bulk silicate composition of Homestead is also similar to currently used model Martian compositions [3]. Accordingly, Homestead was chosen as the starting material for these experiments.

The two specific objectives pursued in this stage of the project were a) to begin exploring the melting relations of Homestead L5 and b) to test the garnet fractionation hypothesis for generating the Mars composition array. Good progress was made on the former objective, and one experiment was successfully completed that helps address the latter.

Procedure: The approach used here was analogous to a study by Agee et al. [4] of the Allende meteorite as a model primitive solar system composition. Experiments were performed at Johnson Space Center using standard techniques for a Walker type multianvil device [4]. 5 GPa pressures were achieved using an 8-mm truncation edge length on the comers of Toshiba tungsten carbide cubes. Matching cast, finned octahedra of Aremco Ceramaceast 584 ceramic were the pressure media. The starting material was a powdered natural sample of the Homestead L5 chondrite, stored in a drying oven between experiments. The powder was placed inside graphite capsules with lids and insulated by hard alumina sheaths. Samples were positioned in the center of a rhenium foil heaters and bracketed by crushable alumina spacers. Temperatures were measured by W/Re thermocouples wrapped around the middle of the heater near the capsule. Pressure was applied at approx. 5 GPa/hour and decreased at approx. 2.5 GPa/hour. An external controller connected to the thermocouple regulated the temperature. The assembly was heated at about 500°C/min and was quenched by cutting power to the heater. The sample temperature fell below the transition to glass in seconds and to room temperature in less than 5 minutes. Each 5 GPa experiment was kept at run conditions for 3 hours.

One experiment was completed at 15 GPa, using a 6-mm truncation edge length and correspondingly smaller cast octahedra. The experiment lasted about 30 minutes before the heater failed.

Run products were examined using Johnson Space Center's Cameca SX100 electron microprobe. It was used to identify phases and measure their compositions as well as to take pictures. The electron beam was operated with a 15 kV accelerating potential and beam currents ranging from 20 to 100 nA, and the beam sizes ranged from ~1 micron for crystals to 20 microns for quenched liquid areas.

Results: The run conditions and phase assemblages for each successful experiment are listed in Table 1. The ranges in composition for each silicate phase are listed in Table 2. The 15 GPa experiment has not yet been analyzed due to the difficulty of polishing the graphite capsule that partially transformed to diamond. Most supersolidus experiments also contained two immiscible metallic liquids, but this study focused only on the silicates.

Figure 1 shows backscattered electron images of AW8 (subsolidus), AW4 (supersolidus), and AW6 (superliquidus) run products. In the subsolidus run, the bright metallic phases, the light gray olivines, and the dark gray pyroxenes were spread homogenously throughout the sample. In AW4, 1800°C, the phases were clearly segregated. The left two-thirds of the sample were entirely liquid with the metal contained in one large area, and the rest was a fine-grained silicate quench texture. The remaining third of the sample was crystalline with one metallic area. The pyroxenes and the olivines were much larger than in the subsolidus experiment. The bright areas just below the boundary into the quench texture were holes caused by plucking during polishing, not metallic phases. In the supersolidus experiment, AW6 at 1850°C, almost all of the metal had gathered at the extreme right edge of the sample. The light gray area was fine-grained silicate quench with no crystalline phases present.
The six successful experiments at 5 GPa bracketed the solidus between 1575°C and 1650°C and the liquidus between 1800°C and 1850°C.

The crystalline run products were predominantly olivine and low-calcium pyroxene, with other phases present but less prevalent. The lowest temperature run, AW1 at 1500°C, contained only olivine (Mg# = 77) and pyroxene. At 1575°C to 1750°C, two pyroxenes were identified, one with relatively more calcium than the other, as well as olivine and liquid. The lower-Ca pyroxene ranged in composition from Wo3En7gFs19 at 1500°C to Wo1En90F5g at 1800°C, while the higher-Ca pyroxene ranged from Wo11En70Fs19 at 1575°C to Wo9En74Fs17 at 1750°C. Garnet was identified only in AW3 at 1750°C, and it was about 1% of the total silicate volume. In AW4, 1800°C, only one pyroxene phase was present, as well as olivine (Mg# = 88) and liquid. The highest temperature run, AW6 at 1850°C, was entirely liquid with an average silicate composition very close to that of the original Homestead meteorite [2].

Discussion and Future Study: The Homestead meteorite melting interval is approximately the same width as that of the Allende meteorite [4], 200± 50°C, however both the liquidus and the solidus are shifted 50-100°C higher, consistent with the fact that Allende is richer in FeO than is Homestead.

As expected, the Mg# of both the olivines and the pyroxenes increases with increasing temperature. It is yet to be determined why the runs at 1500°C and 1800°C have only one pyroxene phase and why garnet is present only at 1750°C. The crystal habit and mean Z of garnet in AW3 are very similar to that of pyroxene, so it is conceivable that garnet is present in other runs but has not yet been identified. To continue the first objective of this project – exploring the melting relations – experiments will be performed to determine the liquidus and solidus of this composition at higher pressures.

Figure 2 plots CaO/Al2O3 versus Al2O3/TiO2 and provides a preliminary test of the garnet fractionation hypothesis, the second objective of the project. On this graph, the SNC meteorites and the Pathfinder rocks lie along a coherent trend; Homestead L5 lies above it. A trend for progressive garnet fractionation calculated using the crystal-liquid D values for garnet produced in AW3 (TiO2 = 0.54, Al2O3 = 5.21, CaO = 0.74) progresses from Homestead L5 towards the trend of Mars compositions, intersecting it between 20 and 30% garnet crystallization. In comparison, using the Dreibus and Wänke (1985) [4] mantle composition, the garnet fractionation trend – approximated using the same AW3 garnet composition – does not intersect the Mars trend. The effect of pyroxene fractionation on this plot is small in comparison to garnet, and the removal of olivine has no effect.

Garnet in AW3 was not observed to be in direct contact with quenched liquid, so this result is provisional, but it is nevertheless promising as support for further investigation of the garnet fractionation hypothesis. To continue the study, more experiments at high pressure will be performed in an attempt to produce garnet and majorite in contact with the quenched liquid and in greater quantities (approaching 20 to 30% crystallization). After the conditions are found where garnet is unequivocally in contact with the melt, a starting material doped with trace elements will be constructed so that trace element partitioning relationships can be determined as well. This will allow more detailed comparisons to SNC compositions and more thorough test of the garnet fractionation hypothesis.


Acknowledgements: I would like to thank first the Lunar and Planetary Institute for providing me with the opportunity for this internship, second the NASA Johnson Space Center for allowing me to use space and equipment, and finally all the people who helped make this project a success – Carl Agee, Dave Draper, Nancy Chabot and Dimitris Xirouchakis in particular.
Table 2: Composition Ranges for Crystals

<table>
<thead>
<tr>
<th>Wt%</th>
<th>Olivine</th>
<th>Low Ca Pyroxene</th>
<th>High Ca pyroxene</th>
<th>Garnet</th>
</tr>
</thead>
<tbody>
<tr>
<td>K₂O</td>
<td>0.011’</td>
<td>0.015’</td>
<td>0.023’</td>
<td>0.005’</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.06 - 0.13</td>
<td>0.17 - 0.92</td>
<td>1.39 - 2.34</td>
<td>0.06 - 0.12</td>
</tr>
<tr>
<td>MgO</td>
<td>40.4 - 48.6</td>
<td>28.1 - 35.7</td>
<td>22.8 - 26.9</td>
<td>21.8 - 23.8</td>
</tr>
<tr>
<td>SiO₂</td>
<td>39.2 - 41.6</td>
<td>54.7 - 59.2</td>
<td>54.3 - 57.1</td>
<td>42.5 - 44.2</td>
</tr>
<tr>
<td>SO₂</td>
<td>0.019’</td>
<td>0.073’</td>
<td>0.063’</td>
<td>0.008’</td>
</tr>
<tr>
<td>CaO</td>
<td>0.10 - 0.21</td>
<td>0.52 - 1.68</td>
<td>3.93 - 5.59</td>
<td>2.48 - 2.98</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.02’</td>
<td>0.01 - 0.07</td>
<td>0.04 - 0.11</td>
<td>0.13 - 0.15</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.19 - 0.36</td>
<td>0.33 - 0.78</td>
<td>0.74 - 1.11</td>
<td>2.56 - 3.11</td>
</tr>
<tr>
<td>MnO</td>
<td>0.26 - 0.37</td>
<td>0.19 - 0.38</td>
<td>0.37 - 0.45</td>
<td>0.53 - 0.57</td>
</tr>
<tr>
<td>NiO</td>
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<td>0.01 - 0.07</td>
<td>0.02 - 0.04</td>
<td>0.03’</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.10 - 0.22</td>
<td>0.91 - 2.63</td>
<td>3.03 - 4.5</td>
<td>18.7 - 20.5</td>
</tr>
<tr>
<td>P₂O₅</td>
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<td>0.098’</td>
<td>0.03 - 0.08</td>
<td>0.11 - 0.14</td>
</tr>
<tr>
<td>FeO</td>
<td>12.2 - 21.9</td>
<td>6.3 - 12.4</td>
<td>9.9 - 11.4</td>
<td>10.3 - 12.0</td>
</tr>
</tbody>
</table>

Range of compositions of each crystalline phase in compound weight percent. The ranges are defined by the averages of the minerals in each run product.

† maximum value; some points are below detection limit.

Figure 1: Electron backscattered images of selected 5 GPa run products. Top: AW8, subsolidus run at 1575°C. Middle: AW4, supersolidus run at 1800°C. Bottom: AW6, superliquidus run at 1850°C. Scale bar is 500 μm.

Figure 2: Martian compositions compared to garnet fractionation trends. Solid symbols represent SNC and Pathfinder compositions. Open circles show the trend of increasing percentage of garnet fractionation, having the composition measured in AW3, from the bulk Homestead L5 (X). Open diamonds show the same calculated trend but starting from the model Martian mantle of Dreibus and Wänke [3].

0% = all liquid. 50% = 50% liquid, 50% garnet.