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Moscow, Russia
2-3 October 2007

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The 46th Vernadsky-Brown Microsymposium on Comparative Planetology

V.I. Vernadsky Institute of Geochemistry and Analytical Chemistry
Russian Academy of Sciences, Moscow, Russia
October 2-3, 2007
Organizers: Vernadsky Institute and Brown University
Sponsor: Russian Foundation of Basic Sciences

2 October

9:00-9:30 Introductory notes on the Sputnik-50 anniversary: J.W. Head and M.Ya. Marov
Technical notes: A.T. Basilevsky

9:30-13:00 MARS
Co-Chairs: V.-P. Kostama and R.O. Kuzmin
9:30-10:00 A.F. Chicarro. Mars Express – summary of latest scientific results. m46_10
10:00-10:30 J.W. Head, D.R. Marchant. Evidence for non-polar ice deposits in the past history of Mars. m46_22
10:30-11:00 G.A. Morgan. Characterization of intermediate units and layered deposits within the LVF/LDA deposits of the dichotomy boundary of Mars. m46_55
11:00-11:30 R.O. Kuzmin, et al. Mars: Observation of the water amount increasing in the surface layer during winter season within latitude belt ±50° based on the TES and HEND data analysis. m46_49
11:30-12:00 M.A. Kreslavsky, J.W. Head. Assessment of "wet" mechanism of slope streaks formation on Mars. m46_43
12:00-12:30 A.T. Basilevsky, et al. Geologic history of Mangala Valles, Mars, from geologic analysis and crater counts. m46_05
12:30-13:00 V.-P. Kostama, et al. Western Promethei Terra smooth plains region, Mars: A volcanic province? m46_41

13:00-14:00 Lunch break

14:00-16:00 VENUS AND SUPER EARTH PLANETS
Co-Chairs: N. Bondarenko and A.A. Ariskin
14:00-14:30 N.V. Bondarenko. Warm lava flows on Venus? m46_08
14:30-15:00 J. Helbert, et al. Exploring the surface of Venus with VIRTIS on Venus Express. m46_25
15:00-15:30 M.A Ivanov, J.W Head, A.T. Basilevsky. The history of topography on Venus. m46_29
15:30-15:50 T. Tormanen, et al. Coronae and arachnoids of Venus revisited: Sizes and topographic characteristics. m46_71
15:50-16:10 S. Franck, et al. Habitability of super-Earth planets. m46_18

16:10-16:30 Coffee break

16:30-18:00 POSTER SESSION
Conveners: J. Korteniemi, and A.A. Berezhnoi

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N.A. Evdokimova, et al. Seasonal dynamics of the water-ice on the surface of the northern polar cap of Mars based on the Omega data. m46_15
C.I. Fassett, et al. Sedimentary Fan Deposits in Jezero Crater Lake, in the Nili Fossae Region, Mars: Meter-scale layering and phyllosilicate-bearing sediments. m46_17
J. Korteniemi, et al. Possible dike-related features in the Hadriaca Patera region, Mars. m46_40
J. Korteniemi, et al. Morphological surface features in the West Promethei Terra region, Mars. m46_39
D. Mimoun, et al. The SEIS-EXOMARS experiment: A planetary seismometer for Mars. m46_54
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Yu.N. Bratkov. Concentric family of coronae around Great Russian Plane: Comparing with Mars and with Artemis Corona (Venus). m46_09
E.N. Guseva. Topography and extension estimates for rift zones of Beta and Alpha regions. m46_20
M.A. Ivanov. Global geological map of Venus: Preliminary results. m46_28
V.P. Kryuchkov, J. Raitala, T. Tormanen. Distribution of coronae on surface of Venus in compliance with morphological parameters of inside depression of these structures. m46_45

18:00-20:00 American Buffet and Slide Session. Includes talk by Louis Friedman about The Planetary Society and set of short slide presentations on planetary and everyday life issues supported by foods and drinks.

3 October

9:00-13:00 THE MOON, TITAN, MERCURY AND RADIATION ISSUES.
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10:00-10:30 V. Kaydash, et al. Surface variations of phase function steepness for two lunar sites from SMART-1 AMIE data. m46_78
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11:00-11:30 V. Kaydash, et al. Topography of selected lunar areas from SMART-1 AMIE data. m46_34
11:30-12:00 A.V. Rodin, Yu.V. Skorov, H.U. Keller. Microphysics of atmospheric aerosol of Titan. m46_60
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12:30-13:00 G. DeAngelis, et al. Time-dependent models for the radiation environment of planet Mars. m46_11
13:00-14:00 Lunch break

14:00-16:00 EXTRATERRESTRIAL MATERIALS
Co-Chairs: G.J. Flynn and V.A. Alexeev
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POLYGONAL IMPACT CRATERS ON VENUS AND THEIR ASSOCIATIONS WITH SURROUNDING TECTONIC FEATURES. M. Aittola¹, T. Öhman¹,², J.J. Leitner¹,³, V.-P. Kostama¹, J. Raitala¹, and T. Törmänen¹. ¹Astronomy division, Department of Physical Sciences, University of Oulu, Finland, ²Department of Geosciences, University of Oulu, Finland, ³Institute for Astronomy, University of Vienna, Austria. (marko.aittola@oulu.fi).

Introduction: There are impact craters whose shape in plan view is more or less angular instead of being circular or ellipsoidal on all terrestrial planets and also on many icy moons as well as asteroids [1,2]. These craters are called Polygonal Impact Craters (PICs) and they are rather common on bodies that have craters and fractured crusts [e.g. 3,4,5], including Venus [6; see example in Fig. 1]. The available data sets limit the interpretation of the Venusian PICs, but our updated catalogue includes 121 impact craters (Fig. 2), which show at least two straight rim segments. This number includes only the craters larger than 12 kilometres due to the uncertainties caused by the resolution of the Magellan data.

Fig.1. An example of impact craters with more than one straight rim segment. The image pair shows the Behn crater that is situated at 32.5S/142E (Magellan left-looking radar image), with the measured straight rim segments.

Fig. 2 The distribution of polygonal impact craters (diameter > 12 km) on Venus.

Background: The comparison of PICs and “normal”-shaped craters by using different characteristics (diameter, altitude, geologic setting, morphologic class, floor reflectance, degradation state, and wall terracing) showed that the only characteristic, which makes difference between normal- and polygonal impact craters, is the diameter [6]. It was obvious that relative abundance of polygonal impact craters increases among the smaller craters compared to larger craters. Thus, it seems to be evident that the smaller crater sizes favor the formation of straight rim segments, but otherwise these craters show similar characteristics to other craters.

Our study also showed that there are regions where the straight segments of the crater rims most clearly follow the orientations of the dominant tectonic features of the area. However, there is not a large population of impact craters on Venus, like there is on Mars, so we cannot use a statistically reliable population of PICs on Venus to determine their correlation with the local tectonics. However, the regions where the orientations of straight rim segments follow the tectonic alignments close to the craters indicate a possible correlation between crater formation and tectonics around the crater. Thus, the orientations of crater walls may reflect – at least in some places - the local tectonics and zones of weakness also on Venus and could thus tell us about the directions and distributions of fractures or other zones of weakness in the crust, just as in the case of Mars [4].

The purpose of this study is to explore if there can be found some real correlations between the straight walls of the PICs and the surrounding tectonics by analyzing all the tectonic features and their orientations around the craters. Furthermore, we wanted to find out if the type of the tectonics or their distance from the craters affect to these correlations. We also declared if the size of the impact craters would have some influence to possible correlations between the tectonics and straight walls of the PICs.

Data and methods: The study was performed by using the Magellan SAR images, which cover 98% of the surface [7]. For the analysis and measurements we used the full-resolution mosaicked image data with 75 m/pixel size. For the mapping we used compressed once mosaicked image data records (C1-MIDRs) with a resolution of 225 m/pixel. We have used also the altimetry data together with the SAR data, especially when mapping the example areas.

To measure the straight walls of the PICs, we selected the most obvious cases. Therefore, the original amount of PICs (131; [6]) was reduced to 121 for this study. For this catalogue we studied only the impact craters larger than 12 km in diameter due to the limitation of the Magellan data [6]. From the polygonal impact craters we measured all the straight walls and from the surroundings we measured the orientations of the tectonic structures. We
divided tectonic features to several different categories and also made the division based on the distance from the crater. Thus we gathered following information from the surrounding tectonic features: Type, orientation and the distance from the crater. Due to the nature of the Magellan radar data (left- and right-looking directions), we ignored all the East-West oriented features (±15°). That is because they are so difficult to define and it would be impossible to avoid uncertainties. When comparing the measurements of the orientations, we interpreted parallel those which difference is less than ±7.5°.

After considering the limitations mentioned above, we measured orientations of the PICs straight walls as well as all the surrounding tectonic features. The measured tectonic features were divided into classes: (1) young rift zones, (2) old rift zones, (3) wrinkle ridges on plains, (4) mountains, (5) lineaments (not identifiable, single features), (6) structures associated to tessera terrains, and (7) tectonic features associated to volcano-tectonic features - usually annulus of corona or arachnoids (we measured both concentric and radial orientations of the structures compared to PICs). Moreover, all these categories were subdivided on the basis of the distance from the crater in to two classes: Those which are situated less than 2 crater diameter from the PIC and those which are 2-10 crater diameter away from the PIC.

**Results:** Our preliminary results show that especially young rift zones, tessera terrains, and the concentric component of the corona/ arachnoid show strong correlations (parallel orientations) with the straight walls of the polygonal impact craters (Fig. 3). Other observation was that the correlations are systematically better when the tectonic features are close to the craters. The third finding was that the wrinkle ridges in volcanic plains show no clear correlation with PICs, even tough we can measure several different orientations for the ridges. Furthermore, although the smaller crater sizes seem to favor the formation of straight rim segments, the crater sizes do not seem affect to these correlations mentioned above.

Thus, previous studies have shown that there are polygonal impact craters on Venus and the smaller crater sizes do favor the formation of straight rim segments for some reason. The results of this study - the straight rim segments of the craters show clear correlations with certain tectonic features, especially when they are situated close to the crater – indicate that these craters somehow reflect the existing planes of weakness and/or fractures in the target material, just like in the case of Mars. Therefore, these craters could actually tell something about the tectonics hidden beneath the surface.


![Fig. 3. Image pair, which shows the tectonic surroundings of the Austen crater (25S/168E; 43.9 km in diameter). The sketch map in the lower image display the orientations of he straight walls of the impact crater (arrows) and their parallel orientations with the rifts (white lines) and the concentric structures of the corona feature (grey lines).](image-url)
COMPLEX EXPOSURE HISTORIES OF THE CHONDRITES FROM COSMOGENIC NOBLE GASES AND RADIONUCLIDES. V.A. Alexeev, Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow 119991 Russia; e-mail: aval@icp.ac.ru

In calculating cosmic-ray exposure ages of meteorites it is generally assumed that the meteoroids were expelled from a shielding position within their parent body and then experienced a single stage exposure before colliding with Earth. However, a comparison of the radiation ages based on the stable and radioactive cosmogenic nuclides makes possible to reveal the meteorites with a complex exposure history.

At calculation the radiation ages on cosmogenic isotopes of neon, the production rate of a $^{21}\text{Ne}$ ($P_{21}$, in $10^{-8}$ ccSTP/g Ma) is usually calculated according to the formula of Eugster [1]:

$$P_{21} = 1.61 \times F/(21.77 \times (22\text{Ne}/21\text{Ne})_c - 19.32) \quad (1),$$

where $F = 1$; 1 and 0.93 for L-, LL- and H-chondrites accordingly. However, analysis of the Leya et al. [2] data of the production rate of cosmogenic nuclides has shown that the production rates of $^{21}\text{Ne}_c$ in chondrites with a radius $R \leq 65$ cm at low shielding (high values of the $(22\text{Ne}/21\text{Ne})_c$ ratio) are significantly lower than values counted on the equation (1): down to 30%. The Fig. 1 shows the changes of the $^{21}\text{Ne}_c$ production rate at calculation on the equation (1) and according to Leya et al. [2] data. For approximating the calculated data by a functional dependence, we have transformed the equation (1) to identical but simpler type:

$$P_{21} = F/(13.52 \times (22\text{Ne}/21\text{Ne})_c - 12.00) \quad (2),$$

and then have determined the values of constants in equations of the "best" fitted curves for L-, LL- and H-chondrites. In that way we obtained the next equation of the curve:

$$P_{21} = F/(23.5 \times (22\text{Ne}/21\text{Ne})_c - 22.8) \quad (3),$$

where $F = 1$; 1.02 and 0.94 for L-, LL- and H-chondrites accordingly. This equation was used by us below at calculation of the $^{21}\text{Ne}_c$ radiation ages ($T_{21}$).

We used measured contents of $^{10}\text{Be}$ ($T_{1/2} = 1.51$ Ma) and $^{26}\text{Al}$ ($T_{1/2} = 0.705$ Ma) in chondrites for calculation of the radiation ages of $T_{10}$ and $T_{26}$ respectively. The age values are calculated according to the equation:

$$T = (1 – \exp(-\lambda T)) = (a+b \times (22\text{Ne}/21\text{Ne})_c) \times 21\text{Ne}_c/A. \quad (3)$$

Here $A$ and $^{21}\text{Ne}_c$ are contents of cosmogenic $^{10}\text{Be}$ (or $^{26}\text{Al}$) and $^{21}\text{Ne}_c$ accordingly; $\lambda$ is decay constant of $^{10}\text{Be}$ or $^{26}\text{Al}$; the parameters $a$ and $b$ were obtained from a correlation dependences of $P_{10}/P_{21}$ and $P_{26}/P_{21}$ vs. $(22\text{Ne}/21\text{Ne})_c$: see Fig. 2. (The similar equation was suggested in [2] for calculation of the $^{10}\text{Be}$ ages.)

A difference of the ages calculated according to equations (2) and (3) will indicate a complex exposure history of meteorite. The results of calculations for some investigated before meteorites are shown in the Table and in Fig. 3. The $^{10}\text{Be}$, $^{26}\text{Al}$ and Ne data are taken from [3-7]. The coincidence (within of the standard errors of calculations) of the $T_{21}$, $T_{10}$ and $T_{26}$ values for chondrites of Knyahinya and Sena is evidenced their simple exposure histories, whereas the chondrites of Cullison and especially Jilin, Tsarev and Shaw had most probable the complex exposure histories. Let's mark, that the comparison the ages obtained on the radioactive nuclides with different half-lifes allows investigating in more details a radiation history of meteorites.

References:

Table. Chondrites with different exposure histories ($T$ in Ma).

<table>
<thead>
<tr>
<th>Meteorites</th>
<th>$T_{21}$</th>
<th>$T_{10}$</th>
<th>$T_{26}$</th>
<th>$D_{10}$, % *</th>
<th>$D_{26}$, % *</th>
<th>Exposure</th>
</tr>
</thead>
<tbody>
<tr>
<td>Knyahinya L/LL5</td>
<td>40 ± 3</td>
<td>32 ± 4</td>
<td>37 ± 2</td>
<td>-20 ± 12</td>
<td>-8 ± 9</td>
<td>Simple</td>
</tr>
<tr>
<td>Sena H4</td>
<td>1.7 ± 0.2</td>
<td>2.2 ± 0.4</td>
<td>1.8 ± 0.2</td>
<td>30 ± 30</td>
<td>6 ± 17</td>
<td>Simple</td>
</tr>
<tr>
<td>Cullison H4</td>
<td>1.7 ± 0.2</td>
<td>2.5 ± 0.4</td>
<td>2.6 ± 0.3</td>
<td>50 ± 30</td>
<td>55 ± 25</td>
<td>Complex</td>
</tr>
<tr>
<td>Jilin H5</td>
<td>4.8 ± 0.7</td>
<td>11.0 ± 1.3</td>
<td>10.3 ± 1.2</td>
<td>130 ± 40</td>
<td>115 ± 40</td>
<td>Complex</td>
</tr>
<tr>
<td>Tsarev L5</td>
<td>2.1 ± 0.3</td>
<td>6.6 ± 0.6</td>
<td>7.4 ± 0.6</td>
<td>210 ± 50</td>
<td>250 ± 60</td>
<td>Complex</td>
</tr>
<tr>
<td>Shaw L6/7</td>
<td>1.00 ± 0.15</td>
<td>6.8 ± 0.7</td>
<td>3.5 ± 0.3</td>
<td>580 ± 130</td>
<td>250 ± 60</td>
<td>Complex</td>
</tr>
</tbody>
</table>

* $D_{10(26)} = (T_{10(26)}/T_{21}) \times 100$, %
The rate formation of $^{21}\text{Ne}$ ($P_{21}$) vs. ($^{22}\text{Ne}/^{21}\text{Ne}_c$) in L-chondrites with $R \leq 65$ cm (according to the Leya et al. [2] data). 1 – Approximation of the calculated data by the "best" curve of $y = 1/(a + bx)$; 2 – dependence according to the Eugster et al. [1] data.

Fig. 2. Ratios of the rate formations of $^{10}\text{Be}$, $^{26}\text{Al}$ and $^{21}\text{Ne}_c$ ($P_{10}/P_{21}$ and $P_{26}/P_{21}$) vs. ($^{22}\text{Ne}/^{21}\text{Ne}_c$) in L-chondrites with $R \leq 65$ cm (according to the Leya et al. [2] data). $P_{10}$ and $P_{26}$ in dpm/kg; $P_{21}$ in $10^{-10}$ ccSTP/g Ma.

Fig. 3. Comparison of the $^{10}\text{Be}$ (1) and $^{26}\text{Al}$ (2) exposure ages with $^{21}\text{Ne}_c$ ages of chondrites. Data are given in the Table. Meteorites: Sh – Shaw, Cu – Cullison, Se – Sena, Ts – Tsarev, Ji – Jilin, Kn – Knyahinya.
THE MAIN BELT ASTEROID 3628 BOZNEMCOVA AS POSSIBLE SOURCE OF THE LL6-CHONDrites.

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Introduction: Meteorites are the last link in the hierarchy chain of collisions of cosmic bodies. The initial parent bodies of ordinary chondrites should be searched for among ~7-8 large (100-300 km in diameter) S(IV)-asteroids of the main belt being similar to 6 Hebe, and, most probably, among those of them that are situated near its inner boundary (at the semi-major axis $a~2-2.5$ AU) [1,2]. The subsequent collisions produced the intermediate parent bodies of different scale, and those of them with diameter of about ~10 km are considered to be the last parent bodies of chondrites. The collisions, secular resonances with planets and the Kirkwood ports are efficient mechanisms for transferring the fragments, i.e. meteorites, into the more eccentric orbits [3,4]. There is a coincidence of spectral types of some chondrites and asteroids, in particular, the reflectance spectrum of the Manbhoom LL6-chondrite is similar to that of the main belt asteroid 3628 Boznemcova [5]. In this connection, two LL6-chondrites fallen in Africa in 2002 attract attention [6,7]: the chondrite Bensour of total mass of ~45 kg fell on February 11th along the Algerian/Moroccan border, and the chondrite Kilabo of total mass of ~19 kg fell on July 21st in northern Nigeria. In particular, their similar petrography and fayalite composition: (Fa30.7) and (Fa30.9), respectively, could indicate a common source of origin of both the chondrites. It is natural to suppose that just 3628 Boznemcova is the parent body of these LL6-chondrites. Below we adduce some evidence in support of that idea.

Pre-atmospheric sizes of the Kilabo and Bensour chondrites: We have investigated the sample N16703 of mass of 152 g of the Kilabo chondrite. The density of tracks of VH-nuclei and the contents of radionuclides of $^{54}$Mn, $^{22}$Na, $^{60}$Co and $^{26}$Al have been measured [8]. According to the density of tracks of VH-nuclei, the shielding depth of that sample from the pre-atmospheric surface of the chondrite was $d = 6 \pm 3$ cm [9]. The most sensitive indicator of the pre-atmospheric size of chondrites is cosmogenic radionuclide $^{60}$Co produced in the reaction of $^{59}$Co(n,$\gamma$)$^{60}$Co with thermal and resonance neutrons, the accumulation of which is very sensitive to the size of the bodies [10,11]. The modeling of $^{60}$Co depth distribution in the Kilabo chondrite is presented in Fig.1 [12]. The cross corresponds to the average pre-atmospheric radius of the Kilabo chondrite of $R = 34 \pm 4$ cm. Using the density of LL-chondrites of 3.21 g cm$^{-3}$ [13], one may obtain that the pre-atmospheric mass of the Kilabo chondrite, average for the last ~ 8 years before the fall to the earth, is equal to ~529 kg, and the ablation through the passage of the earth atmosphere constitutes ~96.4%.

The contents of the noble gases and some radionuclides in the Bensour and Kilabo chondrites were measured in [7]. If the Bensour chondrite is formed in the same process, its orbit must be in close proximity to that of the Kilabo chondrite (as orbits of swarm of fragments in the case of total destruction of cosmic bodies). The orbit proximity amounts to the proximity of the chondrite velocities at the entrance to the earth atmosphere, and, therefore, the ablation of the chondrites may be comparable. At the ablation of 96.4%, which was undergone by the Kilabo chondrite, the pre-atmospheric mass of the Bensour chondrite was 1250 kg, and its pre-atmospheric radius $R \sim 45$ cm. The ratio of neon isotopes in the Bensour chondrite is measured to be $^{22}$Ne/$^{21}$Ne = 1.123 [7]. That ratio, similar to the tracks of VH-nuclei, can be an indicator of shielding depth of the investigated samples in chondrites of different size [14, et al.]. It is demonstrated in Fig.2. One can see that the ratio of $^{22}$Ne/$^{21}$Ne = 1.123 in the Bensour chondrite sample correspond to its screening depth of $d \sim 4-7$ cm [12].
be solved without the knowledge of their orbits. Just the orbits can identify the belonging of meteorites to some family of the celestial bodies, among which the sources of meteorites, their parent bodies, should primarily be looked for. The knowledge of the pre-atmospheric size and screening of the samples of the Kilabo and Bensour chondrites allow us to estimate the orbits of the chondrites using the “isotopic” approach [1,10,15] and measured values of $^{26}$Al in the samples: $68\pm7$ [8] and $62\pm1.2$ [7] dpm/kg, respectively [16]. The method provides for modeling the minimal and maximum production rates of $^{26}$Al in the samples at definite depth in the chondrites of definite radii, depending on the extent of their orbits, in accordance with the measured growth of the $^{26}$Al content in the chondrites of known orbits (see, please, [16]). The results are presented in Fig.3 together with the orbit of the 3628 Boznemcova [17]. The orbit of the Bensour chondrite (aphelion $q’ = 3.51$ AU; semimajor axis $a = 2.255$ AU; eccentricity $e = 0.557$; orbital period $P = 1236$ days) is rather smaller than that of the Kilabo chondrite ($q’ = 3.6$ AU; $a = 2.3$ AU; $e = 0.565$; $P = 1273.2$ days): seeing the Bensour chondrite had greater mass, it had received a smaller velocity at the explosive pulse. The orbits of both the chondrites intersect that of the asteroid at about 2.15-2.16 AU, i.e. practically at the same point, in the most densely populated range of the interplanetary space near the inner boundary of the asteroid belt, where the collisions are most probable.

complex radiation history of the kilabo chondrite: The cosmic ray exposure ages of the Bensour and Kilabo chondrites are different: their exposure ages $T_21$ from the content of cosmogenic $^{37}$Ne equal 19 and 33 My, respectively [7]. Basing only on these data, one may suppose that the Kilabo chondrite was ejected due to a catastrophic collision of the 3628 Boznemcova asteroid with a cosmic object near the inner boundary of the asteroid belt 33 My ago, and after 14 My the Bensour chondrite was ejected in the recurrent catastrophic collision of the 3628 Boznemcova in that region. However, study of tracks of VH-nuclei in the Kilabo Chondrite has revealed the bimodal dependence of the pyroxene crystal distribution on the density of tracks (see Fig.4), which indicates the complex radiation history of the chondrite [9]: it was exposed to cosmic rays at the surface of the parent body before ejection. Therefore, taking into account the similarity of structure and composition of both the chondrites, the vicinity of their orbits, the intersection of the orbits with that of the asteroid practically at the same point, as well as the complex exposure history of the Kilabo chondrite, the much more preferable scenario is that suggested in [7]: ejection of both the chondrites from the 3628 Boznemcova near ~2 AU from the Sun 19 My ago, the chondrite Bensour being ejected from the deep layers of the asteroid, which were completely screened from the galactic cosmic rays, and the Kilabo chondrite being, probably, exposed to the galactic cosmic rays on the surface of the asteroid during 14 My. The contemporary size of the 3628 Boznemcova asteroid amounts to ~7 km in diameter [17].

PRE-ATMOSPHERIC SIZES AND ABLATION OF THE KILABO AND BENSOUR LL6-CHONDRITES.

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Introduction: Apart from the aerodynamical approach to determination of the pre-atmospheric mass of the meteorites (see, e.g. [1-3]), there are many methods for estimation of the pre-atmospheric size and ablation of meteorites, using cosmogenic stable isotopes of the noble gases, tracks of VH-nuclei of cosmic rays and cosmogenic nuclides [4-10 and their ref.]. All the methods and their combinations are mutually complementary providing us with available information about the catastrophic events in history of the meteorites falling to earth. Indeed, because of collisions in cosmic space and due to erosion, the sizes of meteorite bodies vary continually. Stable isotopes and tracks of the VH-nuclei are accumulated during the whole period of cosmic ray irradiation and, therefore, they testify to the average sizes of meteorites over all their exposure age, whereas the radionuclides with various half-lives provide information on the average sizes of meteorites during \( \sim 1.5T_{1/2} \) of the radionuclide before the entry to the earth atmosphere. Hence, studying meteorites through the use of tracks, cosmogenic stable isotopes and radionuclides with different \( T_{1/2} \), one may retrace the evolution of their pre-atmospheric sizes from the moment of their ejection from the parent bodies up to their entrance to the earth atmosphere [4,11]. For instance, the average radius \( R \sim 30-40 \) cm of the Sen-Severin chondrite was conserved invariable during its whole exposure age of 11.2 My [12], whereas the average radius of the Ehole chondrite was already 3-5 times smaller for the last 8 years before its fall to the earth than for the last 6 My on the average.

Pre-atmospheric size of the Kilabo chondrite: The ordinary LL6-chondrite Kilabo fell in Nigeria on July 21\(^{st}\) 2002 as several fragments of total mass of 19 kg; as judged from the recorded detonation, it had underwent destruction in the atmosphere [13]. We investigated the sample N 16703 of mass of 152 g. According to the density of tracks of VH-nuclei, the shielding depth of that sample from the pre-atmospheric surface of the chondrite was \( d = 6 \pm 3 \) cm [14]. The most sensitive indicator of the pre-atmospheric size of chondrites is cosmogenic radionuclide \( ^{60}\text{Co} \) produced in the reaction of \( ^{59}\text{Co}(n, \gamma)^{60}\text{Co} \) with thermal and resonance neutrons, the accumulation of which is very sensitive to the size of the bodies [4, 15]. Our measurement of the \( ^{60}\text{Co} \) content in the sample gives \( 6.0 \pm 2.5 \) dpm/kg. It is obvious that the \( ^{60}\text{Co} \) generation increases directly with the content of Co, which varies over a sufficiently wide range in the chondrites [16,17]. For modeling the \( ^{60}\text{Co} \) depth distribution in the Kilabo chondrite, we used the average content of Co of 0.04% in LL6-chondrites (in particular, in the Sen-Severin LL6-chondrite) [17]. The \( ^{60}\text{Co} \) content, measured at the time of fall of the chondrite Kilabo to the earth, has been produced under the average galactic cosmic ray intensity \( I_0 \sim 0.2586 \text{ cm}^2 \text{s}^{-1} \text{sr}^{-1} \) for \( 8 \) years (i.e. about \( \sim 1.5T_{1/2} \) of \( ^{60}\text{Co} \) before the fall [18,19]. It is clear that, using \( ^{60}\text{Co} \), we obtain the estimate of the average pre-atmospheric size of the Kilabo chondrite for the last \( \sim 8 \) years before its entrance to the earth atmosphere. The results of the modeling are presented in Fig.1.

Fig.1 Dependence of the \( ^{60}\text{Co} \) depth distribution in chondrites of different pre-atmospheric radii \( R \) on the distance \( d=R-r \) from the surface (upper plot), and dependence of the \( ^{60}\text{Co} \) distribution at different distances from the surface on the radius of the chondrites (lower plot). Crosses show the measured content of \( ^{60}\text{Co} \) in the sample of the Kilabo chondrite at the depth of \( d=6 \pm 3 \) cm, as fixed by the track method.

The upper plot shows the \( ^{60}\text{Co} \) distribution in spherical chondrites of radii \( R \), depending on the shielding depth \( d=R-r \) of samples from the surface. The cross is the measured \( ^{60}\text{Co} \) content in the Kilabo chondrite (6.0±2.5 dpm/kg) at the depth of the investigated sample of \( d = 6 \pm 3 \) cm, as identified by the track evidence. The cross corresponds to the average pre-atmospheric radius of the Kilabo chondrite of \( R = 34^{+6}_{-4} \) cm, which is shown on the lower plot, describing the dependence of \( ^{60}\text{Co} \) distribution at various depths \( d=R-r \) from the surface on radius \( R \). Using the density of LL-chondrites of 3.21 g cm\(^{-3}\) [20], one may obtain that the pre-atmospheric mass of the Ki-
labo chondrite average for the last ~ 8 years before the fall to the earth equaled ~529 kg, and the ablation through the passage of the earth atmosphere amounts to ~96.4%. Within the limits of errors, it is in accordance with the statistical estimates of the chondrite ablation evidenced by the stable isotopes of the noble gases, in particular, with the average ablation of 94.1% of the LL-chondrites [21].

Pre-atmospheric size of the Bensour chondrite:

In [22,23] attention is called to the remarkable similarity of the Kilabo LL6-chondrite with the Bensour LL6-chondrite, fallen on February 11th 2002. In particular, their similar petrography and fayalite composition: (Fa30.9) and (Fa30.7), respectively, could indicate a common source of origin of both the chondrites. If they are formed in the same process, their orbits must be in close proximity (as ones of the swarm of fragments in the case of total destruction of cosmic bodies). The orbit proximity amounts to the proximity of the chondrite velocities at the entrance into the earth atmosphere, and, therefore, the ablation of the chondrites may be comparable. The total fallen mass of the Bensour chondrite is ~45 kg [22]. At the ablation of 96.4%, which was undergone by the Kilabo chondrite, and which agrees, within the errors with the average ablation of the LL-chondrites, the pre-atmospheric mass of the Bensour chondrite was 1250 kg, and its pre-atmospheric radius $R \sim 45$ cm. The ratio of neon isotopes in the Bensour chondrite is $^{22}\text{Ne}/^{21}\text{Ne} = 1.123$ [23]. That ratio, as well as the tracks of VH-nuclei, can be an indicator of shielding depth of the investigated samples in chondrites of different size [6,24-28, et al.].

The experimental profile of that dependence in the Sen-Severin LL-chondrite is measured in [26], and only its slight variations are registered in the chondrites of different chemical groups. That dependence for ordinary chondrites is presented in detail in [21], where for its construction the results of the model experiments of [27,28] are used. It is demonstrated in Fig.2. One can see that the ratio of $^{22}\text{Ne}/^{21}\text{Ne} = 1.123$ in the Bensour chondrite correspond to the distance of the investigated sample from the pre-atmospheric surface of $d = R - r \sim 4-7$ cm.

The obtained pre-atmospheric radii and depth locations of the samples of the Kilabo and Bensour chondrites are used for modeling $^{26}\text{Al}$ depth distributions in the chondrites to estimate the extent of their orbits [29]. Besides, they are used for estimation of radiation conditions in the heliosphere in the phase of maximum of the 23rd solar cycle [30].

References:
GEOLOGIC HISTORY OF MANGALA VALLES, MARS, FROM GEOLOGIC ANALYSIS AND CRATER COUNTS. A.T. Basilevsky1,2, G. Neukum1, S. Werner1*, S. van Gasselt1, A. Dumke, T. Kneissl1, D. Rommel1, L. Wendi1, U. Wolf1, W. Zuschneid1, and J. W. Head3; 1-Institute for Geosciences, Freie Universitaet Berlin, Germany, 2-Vernadsky Institute, RAN, Moscow, Russia (atbas@geokhi.ru), Department of Geological Sciences, Brown University, Providence, RI USA. * - now at Geological Survey of Norway.

Introduction. Our study of the Mangala region attempted to answer the following questions: 1) Was there only one, or more than one source for the valley-carving water, and 2) Was there only one, or were there more than one periods of flooding. We addressed these questions through photogeologic analysis and mapping of the upper-middle reaches of Mangala Valles using high-resolution stereo images taken by the HRSC camera (Mars Express) as well as MOC and THEMIS images. The mapping was accompanied by impact crater counts in more than 60 localities. The large number of crater counts and the high-resolution (12.5 m/px for HRSC, 3-6 m/px for MOC and 18 m/px for THEMIS) images used for the geologic analysis and crater counts provided the possibility to find new evidence for the water sources, and to obtain robust crater statistics and thus reliable age estimates. Early results are published in [1].

The mapped area is between 12° to 19.6° S and 148.7° and 150.7° W (Figure 1). We identified 19 morphologic units, 13 of them represent different parts of Mangala Valles, one unit represent recently identified glacial deposits [6], and five units represent highland terrains and (relatively) large craters.

Observations and analyses. Figure 1 shows that Mangala Valles begin in the opening of the graben trough, which is considered as a source of artesian water that carved the valley [e.g., 2-5, 7, 8]. The valley consists of two components: a broad high-standing erosional terrace (unit ps) and a set of deeper, incised valley elements. The terrace shows mostly large-scale traces of water erosion (downstream oriented grooves and a streamlined crater) while small scale traces of water erosion are generally not preserved here. The majority of elements of the deeper-incised part of Mangala valley show features indicative of large-scale violent erosion: downstream-oriented sets of grooves and streamlined islands (Figure 2).

Figure 2. Middle reaches of Mangala Valles. Outlined is area of crater counts on MOC image M08-04127 (see Figure 5b).

Morphologically smooth valley floor facies have also been observed, locally with meandering channels. The surface material of the smooth floor embays the floor-and-slope grooves, suggesting that it formed during the waning stage(s) of the flood(s). In several places we found features suggesting artesian water release, in addition to the major release from the valley head graben. These include a locality of sudden change of orientation of the valley slope grooves (Figure 3) and a locality of small troughs coalescing into a tributary to the main channel, both in association with tops of magmatic dikes (arrows in Figure 3) as well as heads of clusters of small sinuous channels cutting “normal” grooved slopes, and relatively small subareas of chaos-like terrain (Figure 4).

Representative areas for different elements of the valley were selected and more than 60 crater counts, have been made on them mostly using HRSC and MOC images. We found a large range of crater age estimates even for the same morphologic unit; this, however, does not discredit the results because the geologic analysis and mapping showed that most units had been repeatedly resurfaced after their initial formation. This is very applicable to the broad terrace part of the valley. Of 16 crater counts, 5 showed ages around 3.5 Ga with episodes of resurfacing (Figure 5a), while others show a broad spectrum of younger ages which, however, essentially coincide with episodes of formation and resurfacing recorded in the elements of the deeply incised part of the valley.

The latter have been characterized by counts in 30 localities. They also show a wide range of ages. We did not find any systematic age differences between the units representing these elements. This shows that these units represent the hydrological environments which could occur in the same or in different places at different times. It is important to note that for the ten localities from ~30 km from the valley beginning and down to the northern end of the mapped area were estimated ages of 0.8 to 1.2 Ga (Figure 5b). Taking in mind the estimate uncertainties this may be considered as one age~ 1 Ga.

Figure 5. Selected examples of crater count plots; a) broad valley terrace, count area on HRSC image is outlined in Figure 3; b) valley floor, count area on MOC image M08-04127 is outlined in Figure 2; c) valley head trough (MOC R22-01484 + R23-00048).

We also find significant that closer to the valley head, and within it, younger ages are typical with only the 0.2 Ga ages at the head trough and its opening to the north (Figure 5c). Similarly young ages are also found for the downstream parts of the valley. If we examine the histogram showing all ages determined for the Mangala study area (Figure 6), three peaks centered at about 3.5, 1 and 0.2 Ga are very noticeable with an additional peak around 0.4-0.5 Ga.

Figure 6. Histogram of crater age estimates.

Summary. We interpret the above mentioned observations and analysis as evidence that the Mangala Vallis channel system formed as a result of at least three distinct flooding episodes. In all three episodes the major water source was in the valley head graben trough. In addition to the latter there were several more sources that also released artesian water, but they played a subordinate role in the formation and evolution of Mangala Valles.

Acknowledgements. This work forms part of the HRSC Experiment of the ESA Mars Express Mission and has been supported by the German Space Agency (DLR) on behalf of the German Federal Ministry of Education and Research (BMBF) and by the Deutsche Forschungsgemeinschaft (DFG). We thank M. Chapman, R. Greeley and R. Kuzmin for fruitful discussions and Sebastian Walter for computer help.

GEOLOGIC INTERPRETATION OF THE SURFACE THERMAL EMISSION IMAGES TAKEN BY THE VENUS MONITORING CAMERA, VENUS EXPRESS: THE APPROACH AND INITIAL RESULTS. A.T. Basilevsky1,2, E.V. Shalygin3, D.V. Titov2, W.J. Markiewicz2, F. Scholten4, M.A. Kreslavsky5. 1- Vernadsky Institute, Moscow, Russia (atbas@geokhi.ru); 2 - Max-Planck Institut fuer Sonnensystemforschung, Katlenburg-Lindau, Germany; 3 – Astronomical Institute of Kharkov National University, Kharkov, Ukraine; 4 – DLR Institut fur Planetenforschung, Berlin, Deutschland; 5 - University of California, Santa Cruz.

Introduction. The Venus Monitoring Camera (VMC) is a part of Venus Express payload. It takes images in four channels, one of which is centered at 1.01 µm. When the camera looks at the night-side of Venus, the channel registers thermal emission from surface at the equatorial-to-low northern latitudes [6]. Spatial resolution of these images at the working distances (2,000 – 8,000 km) is 1 to 5 km, but because the surface radiation passes through the scattering atmosphere and cloud layer on its way to the camera, the effective spatial resolution in the surface images is about 50 km. Modelling of the atmospheric blurring is an essential part of this work. The camera takes sequences of images, which are mosaicized by orbits and by regions of study. We report results of initial analysis of the first study region which covers the latitude range of 35°S to 45°N and longitudes from 240° to 315°E. Complementary to VMC, observations of thermal irradiation of Venus surface at its southern hemisphere are being done by the Venera Express mapping spectrometer VIRTIS [3].

Intensity of the surface thermal emission at 1 μm strongly depends on its temperature thus giving a hope to register ongoing volcanic eruptions, and on the emissivity of the surface material which is a function of a number of parameters including surface texture and mineralogy. The latter gives a principal possibility for search for terrains which composition is different from dominant basalts. On Venus, surface temperature is a function of surface elevation. So in the search for fresh lava flows and in attempts to find mineralogical differences, it is necessary to take into account the altitude of the given place.

Modelling surface black body emission at the top of the atmosphere. The 1 µm spectral “window” is free from atmospheric absorption bands [7]. Thus only scattering on clouds particles and atmospheric gases produce atmospheric blurring. We use Monte-Carlo-based code to model light scattering in flat multi-layer atmosphere. The vertical structure of clouds as well as their optical properties have been taken from [13] and Rayleigh scattering coefficient of the lower gaseous atmosphere from [8]. We assume that such thick atmosphere (together with not very oblong phase function) will give us orthotropic radiation field on the top of the atmosphere. That is why we can obtain point spread function of atmosphere and use it to blur images of surface brightness in azimuthal (orthographic) projection. Such modelling gave us point spread functions with half-width ~50 km, which is in agreement with VMC data.

We used topographic maps (GTDR data set) derived from Magellan Radar Altimeter [1] The surface topography has been converted into the temperature distribution maps assuming constant temperature gradient (~8.1 K/km) and then to the images of black body brightness by calculating values of Plank function at 1.01 μm wavelength in each surface point. The surface emissivity of 1.0 was assumed. As the final step these images were blurred by the mentioned point spread function. This gives us modeled images of surface blackbody emission which we compare with the VMC images.

Geologic analysis. As it was mentioned above, the VMC image analysis can potentially progress in two directions: 1) to search for the ongoing or very recent (solidified but still “warm”) lava flows and 2) to look for areas whose surface emissivities at 1 µm differs from surroundings. Strategy for the first direction is to look attentively at areas of geologically young rifts and volcanoes comparing VMC images with images of model black body of the same areas. Simple estimations based on lava black body emission and atmospheric blur show that fresh lava surfaces with temperature of 1500, 1300, 1100, 1000, and 900 K can be detected if their surface is 0.5-1, 3-5, 20-30, ~100 and ~500 km² respectively. Similar assessment has been made by [2]. Typical lava flows in the recent volcanic eruptions on Earth are not larger than a few km² (www.soest.hawaii.edu/ GG/HCV/eruption.html; vulcan.wr.usgs.gov/Glossary/LavaFlows). If it would be the case for Venus, only lava fields with surface temperature 1300-1500 K could be detected. However, not so often but larger flows as long as a few tens of kilometers are also observed in terrestrial volcanic eruptions and this gives more hope to find evidence of ongoing eruption in analysis of VMS images especially taking in mind that lava flows from the past eruptions on Venus are often up to several hundreds km long. Now we are in a process of such search in the young rifts and volcanoes of the study region.

Strategy for the second direction of the analysis is to pay special attention to the terrains which mineralogical composition could differ from that of basalt. Obvious type of targets for this are mountain tops that showed anomalously low radar emissivity interpreted to be due to presence of conductive, semiconductive, ferroelectric or ferromagnetic materials [e.g., 4, 10, 11, 12, 14] which might have 1 µm emissivity noticeably different from that of basalts. In the study area these are tops of Theya Mons and Rhea Mons where the anomalous areas are rather extensive. In VMC images these high-standing areas show low brightness closely resembling what is seen in the images of model blackbody emissions. Hence, the potential effect of difference in 1 µm emissivity, if exists, is masked by the elevation effect. Further quantitative analysis is needed which is complicated by low signal-to-noise ratio of VMC night side images and absence of accurate atmospheric model.

Second type of the targets is represented by massifs of tessera terrain which are elevated above the surrounding basaltic plains. Nikolaeva et al. [9] compiled several evidences that tessera might be composed of the material geochemically more differentiated than basalts, e.g. silicic rocks, whose mineralogy is different from that of basalts. Such rocks should have lower 1 µm emissivity, than basalts. Later, joint analysis of the gravity field and topography of Ishtar Terra allowed [5] to conclude that some parts of Maxwell Montes consisting of material structurally similar to tessera, could be composed of material less dense than basalt and possibly silicic.
of the VMC night-side images. Typical for them is the presence of rather large radar-bright flows on the moderately dark background. The flows and the background around them are typically on the same elevation. Higher radar brightness of these flows is usually considered as a result of their higher surface centimeter- to meter-scale roughness, but some difference in mineralogy can not be excluded and we are testing this through analysis of VMC images. Within the study region we selected two areas with radar-bright lava flows on a darker background. Figure 3 shows one of them.

Figure 2. 400 x 400 km area of radar-dark and bright regional plains centered at 20°N, 300°E. a) Magellan SAR, b) model black body emission, c-d) VMC images, orbits 0365, 0367.

It is seen in Figure 3 that radar bright lava flow does not differ from its radar-dark background neither in model black body emission nor in VMC images. This supports earlier suggestions that radar brightness of lava flows is due to their surface roughness rather than due to composition differences.

Summary. At this initial stage of the study we developed method of producing images of modelled black body emission, formulated goals and determined targets of the geologic analysis and acquired first results of the analysis showing that the altitude/temperature effect is dominating over potential effect of differences in surface material mineralogy.

We acknowledge help of M.A. Ivanov and L.V. Moroz.

**IMPACT-PRODUCED EXOSPHERE OF MERCURY.**

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**Introduction:** Meteoroid impacts are considered to be one of the main sources of calcium in the exosphere of Mercury [1]. Killen and co-workers proposed that calcium is delivered in the form of calcium oxide. Other sources of exospheric Ca discussed in the literature include ionic sputtering and solar wind. Meteoroid bombardment is a more important source for the night side of the planet because ionic sputtering produces atoms in the exosphere only at the day side. Study of the composition of the impact-produced exosphere of Mercury is useful for estimating the elemental composition of the surface of the planet, which is still poorly known. For detailed study of the exosphere of Mercury, optical and ultraviolet spectrometers on board the NASA Messenger spacecraft will be used [2].

**Physics and chemistry of meteoroid impacts:** Hypervelocity impacts lead to vaporization of the impactor, vaporization, melting, and fragmentation of the regolith, and then to the ejection of multi-planetary soil. We use in our calculations $T_0 = 10^4$ K and $P_0 = 10^4$ bars. The time scales of reactions in which Na-, K-, Ca-containing species participate are comparable with hydrodynamic time scale for $T_{\text{quench}} ~ 3000$ K and $P_{\text{quench}} ~ 30$ bars. The time scales of reactions in which Na-, K-, Ca-containing species participate are comparable with hydrodynamic time scale for $T_{\text{quench}} ~ 3000$ K and $P_{\text{quench}} ~ 30$ bars. The time scales of reactions in which Na-, K-, Ca-containing species participate are comparable with hydrodynamic time scale for $T_{\text{quench}} ~ 3000$ K and $P_{\text{quench}} ~ 30$ bars. The time scales of reactions in which Na-, K-, Ca-containing species participate are comparable with hydrodynamic time scale for $T_{\text{quench}} ~ 3000$ K and $P_{\text{quench}} ~ 30$ bars. The time scales of reactions in which Na-, K-, Ca-containing species participate are comparable with hydrodynamic time scale for $T_{\text{quench}} ~ 3000$ K and $P_{\text{quench}} ~ 30$ bars.

At the high temperatures and pressures prevailing immediately after an impact, atoms are more abundant than molecules in the fireball. As the fireball expands and cools, complex species form. At quenching, the major components of a fireball are Na, NaOH, and O$_2$. The elements K, Al, Mg, Fe are present in the gas phase mainly as KOH, K, AlO, MgOH, Fe (see Table). At 2500-3500 K, Ca is present in the form of Ca(g), CaOH(g), Ca(OH)$_2$(g), and CaO(g).

**Photolysis of molecules in the exosphere:** The fraction of molecules destroyed during ballistic flight by photolysis may be estimated by comparing ballistic flight times and photolysis lifetimes. On Mercury, the ballistic flight time of species such as NaOH, KOH, AlO, MgO, CaO, FeO is about 700 s for gas temperature of about 3000 K. Photolysis lifetimes $\tau$ are determined from solar photo rates $J^1$. Only photons with energies exceeding the binding energy of a molecule cause dissociation. We expect that the photolysis lifetimes of molecules are proportional to the solar flux at the binding energy equivalent wavelength. We estimate photolysis lifetimes of major metal oxides based on the photolysis lifetimes and molecular constants of atmospheric diatomic species. The unattenuated solar flux, solar photo rates, binding energy equivalents and dissociation energies of atmospheric species were taken from [4]. The correlation between solar photo rates and the unattenuated solar flux $F$ at $\lambda_{\text{binding}}$ for atmospheric diatomic species is strong ($r^2 = 0.9$), the fit matches available data to within a factor of 10. Solar photo rates $J$ at 1 a.u. correlate with the solar flux $F(\lambda_{\text{binding}})$ for atmospheric diatomic species as $J = 3 \times 10^{13} \times F(\lambda_{\text{binding}})$, where $F$ is in ph$\times$cm$^{-2} \times$s$^{-1} \times$nm$^{-1}$ and $\lambda_{\text{binding}}$ (nm) $= 1240/E$(eV) is the binding energy equivalent wavelength.

From known data about dissociation energies of monoxides and described above obtained equation used for mean heliocentric distance of Mercury (0.39 a.u.) photolysis lifetimes on Mercury were estimated to be 80, 80, 200, 200, 600, 2000, and 30000 s for MgO, KO, FeO, CaO, SiO, AlO, TiO, respectively. Comparing ballistic flight times and photolysis lifetimes we see that the probability of dissociation of MgO, KO, CaO, FeO during its ballistic flight is higher than 90 %, and that the probabilities of SiO, AlO, and TiO dissociation are about 80, 40 and 5 %, respectively.
Properties of impact-produced atoms in the exosphere: NaOH and NaO will be destroyed by solar photons during its ballistic flight on Mercury because its photolysis lifetimes [5] are shorter than their ballistic flight times. Meteoroid bombardment leads to formation of two different Na components in the exosphere of Mercury, a low-energy component (0.2-0.4 eV) released directly during impact processes, and a high-energy component produced by the photodissociation of NaO and NaOH, is about 1-2 eV. The energy of the second component of Na atoms is about 1-2 eV because this value is typical excess energy of photolysis products of diatomic molecules. These same two components are represented in the atoms of other elements.

Our thermochemical results and comparison of ballistic and photolysis lifetime for CaO show that most of impact-associated Ca found in the exosphere of Mercury was not injected directly, but was produced by the photodissociation of CaO, CaOH, Ca(OH)2, which formed in the cooling fireball. These results are consistent with the detection of very hot Ca atoms in the exosphere of Mercury [1].

We estimate the concentrations of impact-produced atoms in the exosphere of Mercury based on measured Ca concentration. We do not take Na or K column density as reference because atoms of these elements have low temperature (1500 K) and meteorite bombardment is minor mechanism of its delivery to the exosphere [6]. We assume that atoms are captured by the surface after their first ballistic flight and column density of impact-produced atoms is proportional to its fireball abundance and probability of photodissociation of monoxides. Then \( \frac{[X]}{[Ca]} = \frac{f_x}{f_Ca} \cdot \frac{F_{\text{uncond}}(X)}{F_{\text{uncond}}(Ca)} \), where \([X], [Ca]\) are the column densities of considered element X and Ca in the exosphere, \(f_x\) and \(f_Ca\) are partial concentrations of considered element and Ca in the fireball, \(t_x(t_Ca)\) are ballistic flight times of considered atoms X and Ca atoms in the exosphere, \(f_x\) and \(f_Ca\) are ratios between abundance of atoms produced directly during impact and by photolysis and total abundance of species in the gas phase of the fireball for X and Ca, respectively, \(F_{\text{uncond}}(X)\) and \(F_{\text{uncond}}(Ca)\) are fractions of uncondensed species of element X and Ca in the gas phase, respectively.

The Ca column density on Mercury is \(1.3 \times 10^8\) cm\(^{-2}\), the temperature of Ca atoms is 12000 K [1]. If we do not consider condensation of refractory elements in the fireball we can use \(F_{\text{uncond}}(X) = 1\) and \(F_{\text{uncond}}(Ca) = 1\). Based on our modeling calculations, we estimate column densities of impact-produced atoms. These values are only upper limits because Ca may be delivered to the exosphere not only by meteoroid bombardment. The condensation of solid phases can significantly reduce the content of metal-containing species in the gas phase in the fireball. To take condensation into account we estimate fractions \(F_{\text{uncond}}(X)\) and \(F_{\text{uncond}}(Ca)\) assuming that fractions of uncondensed species correspond to the equilibrium fireball chemical composition at quenching parameters of chemical reactions (3000 K and 10 bars). Our estimated abundances of impact-produced atoms of main elements are much lower than those predicted by [7].

**Table.** Summary of behavior of metals and Si during meteoroid bombardment of Mercury. Models 1 and 2 give day-time abundances without and with condensation, respectively.

<table>
<thead>
<tr>
<th>Element</th>
<th>Main species</th>
<th>Abundance of impact-delivered during meteoroid impacts</th>
<th>Model 1</th>
<th>Model 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na</td>
<td>Na, NaOH, NaO</td>
<td>7 \times 10^7 , 2 \times 10^9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>K</td>
<td>K, KOH, KO</td>
<td>10^6 , 3 \times 10^7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ca</td>
<td>CaO, CaOH, Ca</td>
<td>1.3 \times 10^8 , 1.3 \times 10^8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Al</td>
<td>AlO, AlOH</td>
<td>2 \times 10^5 , 7 \times 10^5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fe</td>
<td>Fe, FeO, Fe(OH)_2</td>
<td>2 \times 10^5 , 3 \times 10^5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mg</td>
<td>MgO, Mg(OH)_2</td>
<td>3 \times 10^7 , 2 \times 10^8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Si</td>
<td>SiO_2, SiO</td>
<td>6 \times 10^4 , 6 \times 10^8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ti</td>
<td>TiO_2</td>
<td>5 \times 10^3 , 2 \times 10^5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Conclusions:** We have estimated the chemical composition of gas-phase species released to the Hermean atmosphere during meteoroid impacts (see Table). A significant fraction of impact-produced molecular species is destroyed by solar photons because for these compounds ballistic flight times are longer than photolysis lifetimes.

If we neglect the condensation of main elements in the fireball, the content of impact-produced atoms of Na, K, Ca, Fe, Mg, Si in the exosphere will be proportional to the content of these elements at the surface of the planet while Al and Ti are depleted in the exosphere due to the long photolysis lifetimes of their oxides. However, condensation of dust grains leads to significant decreases in the content of atoms with high condensation temperatures such as Al.

Meteoroid bombardment leads to the appearance of hot metal atoms (E~ 0.2-0.4 eV) produced directly by impacts and of very hot atoms (E~ 1-2 eV) formed during photodissociation of metal oxides and hydroxides. Atoms of Na, K, and Fe are produced mainly directly by impact while atoms of Ca, Al, Mg, Si, and Ti are delivered to the exosphere by photolysis of their oxides and hydroxides.

WARM LAVA FLOWS ON VENUS? N. V. Bondarenko Institute of Radiophysics and Electronics, National Academy of Science of the Ukraine, 12 Ak. Proskury, Kharkov, 61085, Ukraine. (bndr@kharkov.ua)

Introduction. Recent Earth-based radar studies of Venus surface have shown that some volcanic features in plains region (including fields of shield volcanoes and lava flow complexes) return radar echo with significant linearly polarized component, when illuminated by circularly polarized probing signal [1]. This can occur only when target surfaces are very smooth and rather transparent for radio waves, so that the waves scattered at internal interfaces or inclusions can reach the observer.

The flow transparency significantly increases the chance to observe thermal effect of recent volcanic eruptions on Venus through observations of surface thermal emission, because in this case the thermal emission is formed at some depth, where thermal effect of recent eruptions lasts orders of magnitude longer than at the surface.

In the present work I illustrate the possibility for hot flow detection using results of radiometry and altimetry experiments during Magellan mission to Venus. This study was based on the use of data archived in the GREDR, GEDR, and ARCDR data sets. SAR images were used also for the morphological analysis.

Source data and theoretical background. The values of surface emissivity (at the wavelength of $\lambda = 12.6$ cm) were calculated in the Magellan radiometry values of surface emissivity (at the wavelength of $\lambda$). After that, this value was fine through the surface reflectivity as stated by Fresnel equations. Thus, emissivity of the smooth and rather transparent for radio waves lava flow has to be lower then (1 - $R_f$). $R_f$ of the smooth interface depends only on dielectric permittivity of the flow material and on the incidence angle.

Magellan data allow independent estimates for surface dielectric permittivity through the so called “Fresnel reflectivity” $R_0$ obtained in radar altimetry experiment. $R_0$ is surface reflection coefficient at normal incidence derived using the approximation of received echo sequence by Hagfors law [4]. Spatial resolution for these data was about 15 km x 10 km at low latitudes but their accuracy is not high, the errors of individual measurements can reach ~30% [4].

Thus we can estimate some effective temperature of observed lava flow using both Magellan Fresnel reflectivity and emissivity data. We considered that apparent enhancement of emissivity in comparison with one expected from reflectivity-derived flow dielectric permittivity occurs due to increased temperature. Such approach allows estimation the low limit for the effective flow temperature. It can be only used for smooth terrains because rough upper flow interface easily leads to high emissivity values in comparison to those predicted with Fresnel formula.

"Warm" lava flow? The approach discussed above was applied to a dark lava flow located at about 28°E, 39°N in Bereghinia Planitia. This flow is recognized in the map obtained during earth-based radar polarimetric observations ($\lambda = 12.6$ cm) and has enhanced linear polarization (up to 12%) [5]. The flow is darker in comparison with surrounding surface as seen in Magellan 1st cycle SAR image shown in Fig. 1a. This observation was made from the East at incidence angle of about 39°. During 2nd Magellan cycle (observation from the West, incidence angle ~ 25°) the flow also exhibited darker appearance. This flow is distinctly dark in the earth-based radar map published in [5]. Observation was made here roughly from the South-South-West at oblique incidence angle.

Thus the flow is expected to be smooth at spatial scales of centimeters to decimeters, which control oblique backscattering at 12.6 cm wavelength. The flow is also very smooth at larger spatial scales of meters and to decameters, as seen from Magellan altimetry experiment results. The Hagfors’ roughness parameter map (GSDR data set) for the area under study is shown in Fig. 1b. The flow roughness is very low, ~0.4° (dark shades in Fig. 1b).
To obtain the distribution of possible flow temperature enhancement we referred to ARCDR data set as a source data for particular emissivity calculations. We used also map of $R_0$ distribution from the GREDR data set. Areas rougher than 2.0° as seen in the roughness map (Fig.1b, GSDR data set) were excluded from the analysis.

On the other hand thermal data presented in Fig.1c correspond to the lower limit for the flow thermal enhancement because of unaccounted scattering from the lower surface-flow interface. Spotty appearance of $R_0$ data used in calculations is responsible for the spotty appearance of the thermal enhancements shown in Fig. 1c.

Discussion and conclusions. The results presented in Fig. 1c raise the question, if the sample flow is really hot? Morphological analysis of this area using high resolution SAR mosaics (F-maps) showed that this radar-dark flow is younger in comparison with surrounding radar-brighter surface. The dark material clearly embays large wrinkle ridges, truncates narrow radar-bright lineaments and is free on any superposed tectonic structures.

Thus the flow is rather young. It has to be older 15 years because it is recognized in SAR map obtained in Pioneer-Venus mission in 1978. Unfortunately, PV SAR data have poor resolution and do not allow tracing small changes in size, shape, or fine structure of the flow.

It takes ~ 100 years for a 20-m thick slab of lava emplaced over "normal"-temperature surface to cool down to 50 K temperature excess. For thicker flows the thermal effect may last significantly longer.

Properties of the dark flow in Bereghinia Planitia are consistent with presently existing temperature excess at shallow depth and hence very recent flow emplacement. At the current state of knowledge, however, it is impossible to completely exclude the chance that these estimates are affected by strong systematic errors in the Magellan Fresnel reflectivity data.

CONCENTRIC FAMILY OF CORONAE AROUND GREAT RUSSIAN PLANE: COMPARING WITH MARS AND WITH ARTEMIS CORONA (VENUS).

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**Introduction:** We consider here a planet as an embedding of some nonspacelike object [1, 2]. A preimage of a planet is studied. Let the Earth be an exact sphere with the equator 40 000 km. Let mounts be points at the exact sphere, without heights.

Highest points of big mountain ranges form some concentric circles around Great Russian Plane. Let the center of the structure be the center of the double bridge over Volga in Yaroslavl. We give exact arc distances (lengths of arcs on the exact sphere) from this center to some selected points.

Map scales are given for evaluating data precision. Recall that, for example, scale 1:500 000 means 1 cm : 5 km.

**Ring 1, radius 1165 km:**
1) 322, the highest point of Pridneprovskaya Height, 1163.7 km, map 1:500 000.
2) Yudychvumchorr 1200, the highest point of the Khibines, Kolsky Peninsula, 1165.9 km, map 1:200 000.
3) 324, the highest point of Priazovskaya Height, 1167.8 km, map 1:500 000.

**Ring 2, radius 1345 km:**
1) Narodnaya 1895, the highest point of Ural Mountains, 1345.0 km, map 1:500 000.
2) 471, the highest point of Podolskaya Height (Gologory), 1346.2 km, map 1:3 000 000.
3) Paarkov-Sarlopy 173, the highest point of Kolguev Island, 1346.4 km, map 1:500 000.

**Ring 3, radius 1600 km:**
1) Bolshoy Bakhtobay 657, the highest point of the Mugodzhary (the most southern range of Ural Mountains), 1597.0 km, map 1:500 000.
2) Elbrus 5642, the highest point of the Caucasus, 1597.1 km, map 1:500 000.
3) Gerlakhovsky-Shitt 2654, the highest point of the Carpathians, 1602.8 km, map 1:200 000.
4) Paiyer 1472, the highest point of Polar Ural Mountains, 1603.7 km, map 1:500 000.

**Ring 4, radius 2600 km:**
**Subring 1, radius 2595 km:**
1) Mont Blanc 4810, the highest point of the Alps, 2595 km, map 1:2 500 000.
2) Slættaratindur 882, the highest point of the Faroe Islands, 2598 km, map 1:2 500 000.
3) Vesuvius 1281, a volcano in Italy, 2598 km, map 1:2 500 000.
4) Vatican, Italy, 2599 km, map 1:2 500 000.
**Subring 2, radius 2620 km (chord 2600 km):**
1) Sparta (the center), Peloponnesus, Greece, 2619.2 km, map 1:200 000.
2) London (the center), England, 2621.0 km, map 1:200 000.
3) Novosibirsk (the center), Siberia, 2623.3 km, map 1:500 000.

**Subring 3, radius 2630 km:**
1) Norilsk (the center), Siberia, 2629.2 km, map 1:500 000.
2) Profitis-Ilias 2407, the highest point of Peloponnesus (Sparta, Greece), 2633.8 km, map 1:200 000.
3) Demavend 5604, a volcano, the highest point of Iran, 2634.5 km, map 1:200 000.
4) 606, the highest point of Franz Josef Land, 2635.4 km, map 1:500 000.

**Subring 4, radius 2640 km:**
1) Paris (the center), France, 2638 km, map 1:2 500 000.
2) Berenberg 2277, a volcano, the highest point of Jan Mayen Island, 2642 km, map 1:2 500 000.
**Subring 5, radius 2650 km:**
1) Bukhara (the center), Uzbekistan, 2649.1 km, map 1:500 000.
2) Paikend (the barrow with famous ancient capital city), Uzbekistan, 2649.5 km, map 1:500 000.
3) Ben Nevis 1343, the highest point of the British Isles, 2651 km, map 1:2 500 000.

**Ring 5, radius 12 550 km:**
**Subring 1, radius 12 540 km:**
1) Yerupaja 6617, the highest point of Cordillera Huayhuash, Peru, 12 537.1 km, map 1:500 000.
2) Ausangate 6384, the highest point of Cordillera Carabaya and Cordillera Vilcanota, Peru, 12 541.8 km, map 1:500 000.
3) Bluff Knoll 1096, the highest point of Stirling Range and of South-Western Australia, 12 536 km, map 1:2 500 000.
4) Woodroffe 1440, the highest point of Musgrave Ranges and of South (not South-Eastern) Australia, 12 551 km, map 1:2 500 000.
**Subring 2, radius 12 557 km:**
1) Tiawanaco (cult center), Bolivia, 12 557.2 km, map 1:500 000.
2) Machu-Picchu (cult center), Peru, 12 557.4 km, map 1:500 000.
3) Sacsayhuaman (cult center in Cusco), the spiritual center of Andes (there is such inscription), Peru, 12 557.7 km, map 1:500 000.

**Subring 3, radius 12 576 km:**
1) Ancohuma (Illampu) 7014, the highest point of America, Bolivia, 12 575.7 km, map 1:500 000.
2) Salkantay 6271, the highest point of Cordillera Vilcabamba, Peru, 12 575.9 km, map 1:500 000.

**Comments:** 4 (four) mountains from the ring 3 are placed at one exact circle. The arc radius is 1602.2 km, and the center is 57°40′59″ N, 39°50′31″ E (this point is placed at the center line of
Volga in Yaroslavl). We recall that every 3 (three) points define some circle. However, 4 (four) points being at one circle is an extraordinary situation. This structure exists at the exact sphere, and it becomes approximated (unexact) at the geoid.

The Chomolungma-Chogory unit of length: Chomolungma (so-called Everest) and Chogory are the two highest points of Earth. Consider the distance between them. Let this distance be a span distance, i.e. length of a chord in \( R^3 \). Denote 1 span cc = 1312.54 km. Here cc means Chomolungma-Chogory. Thus we represent span ring radiiuses in cc units:

- the ring 1: 0.89 span cc,
- the ring 2: 1.02 span cc,
- the ring 3: 1.21 span cc,
- the ring 4: 2.00 span cc for the subrings 4, 5,
- the ring 5: 8.08 span cc for the subring 1.

We see here (approximately) integer-valued numbers for rings 2, 4, 5.

If we consider a non-exact ring 5 with the radius nearly 8 span cc, then there exist much more interesting mounts in this corona.

Global units: So-called global unit (gu) was introduced by the author after studying distances at the Easter Island (ostrov Paskhy in Russian) [3]. The distance between the two highest mounts of this island is a local unit of length. There are integer-valued distances (in this unit) at this island, and 1 km = \( \sqrt{3} \) units. So let 1 km be a local unit for Earth, thus denote 1 gu = \( \sqrt{3} \) km.

Consider the center of the Eastern hemisphere of the globe (Earth). The arc distance to Yaroslavl (Spassky monastery) is 4003 gu. Yaroslavl is the center of the coronae under consideration. Also span distances (lengths of chords) to Novgorod (Russia) and to Alice Springs (the center of Australia) are both equal to 4004 gu. Alice Springs is the center of some big corona around Australia, and Novgorod is the center of some small local corona.

It is well-known now ([4, p. 87], [3]) that centers of hemispheres of Earth, Mars, and Venus are preferred (selected, singular) points. It was incomprehensible and not properly understood from non-mathematical point of view. Reasonable explanation could be founded in classical algebra in topos theory [1, 2].

Comparing with Mars: Let Martian cc (mcc) be the distance (arc or span) between Olympus and Ascreaus, the two highest Martian mounts. Thus we have a (non-exact) ring with 1 mcc radius. Namely, Ascreaus 1.0 mcc, Arsia 1.04 mcc, and Alba 1.02 mcc have approximately the same distance from Olympus. Here the difference between arc and span distances isn’t essential. Data precision is rough: big flat mounts with undetermined highest points.

Also the arc distance from Olympus to Elysium 2.52 arc mcc is nearly integer-valued.

Consider the center of the Western hemisphere. The arc distance to Olympus is 4.5 arc mcc. Also we have nearly integer-valued distances to:
- a) the center of Argyre Planitia 2 span mcc,
- b) the center of Hellas Planitia 2.5 span mcc,
- c) the center of Isidis Planitia 3 arc mcc.

The orthodox version of the Impact Hypothesis falls.

Now consider the center of the Eastern hemisphere. We have nearly integer-valued distances to:
- Arsia 2.0 mcc, arc or span, or 6666 \( \approx \frac{8}{5} \) 10000 arc Martian km;
- Olympus 3000 span mcc (Martian km),
- Pavonis 4060 span mgu,
- Ascreaus 4500 span mgu,
- Alba 2.54 arc mcc,
- the center of Argyre Planitia 4 arc mcc.

Some additional conclusion could be made: hemispheres at Rodionova’s map of Mars (2004) are centered unscientifically: the centers by Rodionova are (0°, 90°), (0°, 270°).

Comparing with Venus:
1) Let Venusian cc (vcc) be the distance between Maxwell and Freyja, the two highest Venusian mounts.

Consider the center of Artemis Corona. It seems to be a Yaroslavl-like object. The (non-exact) ring with 8 span vcc radius is: Gula 7.95 vcc, Maxwell 7.96 vcc, Freyja 8.00 vcc, Atira 8.08 vcc, Sif 8.09 vcc.

2) The span distance from the center of Artemis Corona to the center of the Eastern hemisphere is nearly 4 vcc, and the span distance to the center of the Western hemisphere is 6500 Venusian gu (not vcc), it is integer-valued.

Moreover, the arc distance from the center of the Eastern hemisphere to Maxwell (the highest point of Venus) is 12 750 Venusian km (here Venus is an exact sphere with the equator 40 000 Venusian km). It is nearly equal to length of the diameter (12 732 Venusian km). Recall that data precision is approximately 15 km. USGS data for Venus were used.

References:
MARS EXPRESS – SUMMARY OF LATEST SCIENTIFIC RESULTS

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The ESA Mars Express mission, launched on 02 June 2003 from Baikonur, Kazakhstan, on-board a Russian Soyuz rocket, includes an orbiter spacecraft which was placed in a polar martian orbit. In addition to global studies of the surface, subsurface and atmosphere of Mars, with an unprecedented spatial and spectral resolution, the unifying theme of the mission is the search for water in its various states everywhere on the planet by all instruments using different techniques. A summary of scientific results from all experiments after more than three terrestrial years in orbit is given below.

The High-Resolution Stereo Colour Imager (HRSC) has shown breathtaking views of the planet from both hemispheres, pointing to very young ages for both glacial and volcanic processes, from hundreds of thousands to a few million years old, respectively. The IR Mineralogical Mapping Spectrometer (OMEGA) has provided unprecedented maps of H₂O ice and CO₂ ice in the polar regions, and determined that the alteration products (phylllosilicates) in the early history of Mars correspond to abundant liquid water, while the post-Noachian products (sulfates and iron oxides) suggest a colder, drier planet with only episodic water on the surface. The Planetary Fourier Spectrometer (PFS) has confirmed the presence of methane for the first time, which would indicate current volcanic activity and/or biological processes. The UV and IR Atmospheric Spectrometer (SPICAM) has provided the first complete vertical profile of CO₂ density and temperature, and has discovered the existence of nightglow, as well as that of auroras over mid-latitude regions with paleomagnetic signatures and very high-altitude CO₂ clouds. The Energetic Neutral Atoms Analyser (ASPERA) has identified solar wind scavenging of the upper atmosphere down to 270 km altitude as one of the main culprits of atmospheric degassing and determine the current rate of atmospheric escape. The Radio Science Experiment (MaRS) has studied the surface roughness by pointing the spacecraft high-gain antenna to the Martian surface. Also, the martian interior has been probed by studying the gravity anomalies affecting the orbit, and a transient ionospheric layer due to meteors burning in the atmosphere, was identified by MaRS. Finally, results of the subsurface sounding radar (MARSIS) following the late deployment of its antennas due to safety concerns, indicate strong echoes coming from the surface and the subsurface allowing to identify buried impact craters and tectonic structures, as well as the very fine structure of the polar caps. The Northern crust appears thus just as old as the Southern one, owing to the large number of impact basins being recognized. Also, probing of the ionosphere reveals a variety of echoes originating in areas of remnant magnetism.

Mars Express is already hinting at a quantum leap in our understanding of the planet’s geological evolution, to be complemented by the ground truth being provided by the American MER rovers. The nominal mission lifetime of one Martian year for the orbiter spacecraft has already been extended by another Martian year (687 days). During the extended mission, priority is being given to fulfill the remaining goals of the nominal mission (e.g., gravity measurements and seasonal coverage), to catch up with delayed MARSIS measurements during the nominal mission, to complete global coverage of high-resolution imaging and spectroscopy, as well as subsurface sounding with the radar, to observe atmospheric and variable phenomena, and to revisit areas where discoveries were made. Also, an effort to enlarge the scope of existing cooperation is being made, in particular with respect to other missions at Mars (such as MER and MRO) and also missions to other planets carrying the same instruments as Mars Express (i.e. Venus Express). Finally, Mars Express is providing valuable data for the preparation of ESA’s Aurora Exploration Programme first mission to Mars (called ExoMars and including a capable rover to perform astrobiological, geophysical and climatological investigations), in terms of helping identifying potential landing sites on Mars, establishing a useful surface/subsurface geological database, as well refining the existing atmospheric one, in order to assess potential risks for the exploration of Mars. For further details on the Mars Express mission and its science results:

http://sci.esa.int/marsexpress/
TIME-DEPENDENT MODELS FOR THE RADIATION ENVIRONMENT OF PLANET MARS.

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Introduction: The radiation protection is one of the two NASA highest concerns priorities [1]. In view of manned missions targeted to Mars [2], for which radiation exposure is one of the greatest challenges [3], it is fundamental to determine particle fluxes and doses at any time and at any location and elevation on and around Mars [4]. With this goal in mind, models of radiation environment induced by Galactic Cosmic Rays (GCR) and Solar Particle Events (SPE) on Mars have been developed [5]. The work is described [6] as models of incoming cosmic ray [7-9] and solar events [5-6] primary particles rescaled for Mars conditions then transported through the atmosphere down to the surface, with topography and backscattering taken into account, then through the subsurface layers, with volatile content and backscattering taken into account, eventually again through the atmosphere, and interacting with some targets described as material layers. The atmosphere structure has been modeled in a time-dependent way [10-11], the atmospheric chemical and isotopic composition over results from Viking Landers [12-13]. The surface topography has been reconstructed with a model based on Mars Orbiter Laser Altimeter (MOLA) data at various scales [14]. Mars regolith has been modeled based on orbiter and lander spacecraft data from which an average composition has been derived [4-6]. The subsurface volatile inventory (e.g. CO2 ice, H2O ice), both in regolith and in the seasonal and perennial polar caps, has been modeled vs. location and time [15-16]. Models for both incoming GCR and SPE particles are those used in previous analyses as well as in NASA radiation analysis engineering applications, rescaled at Mars conditions [4-6]. Preliminary models have been developed for the surfaces of the Martian satellites Phobos and Deimos.

Results: Particle transport computations were performed with a deterministic (HZETRN) code [17] adapted for planetary surfaces geometry and human body dose evaluations [4]. Fluxes and spectra for most kinds of particles, namely protons, neutrons, alpha particles, heavy ions, pions, muons etc., have been obtained. Neutrons show a much higher energy tail than for any atmosphereless bodies [4]. Results have been obtained for different surface compositions: only at the latitudes closer to the equator the soil is mostly silicatic regolith, whereas for northern or southern locations a suitable mix, with variable ice concentration with time, of ices of water and carbon dioxide needs to be used [4-6]. Results have been calculated for different locations and atmospheric properties models [4-6]. The results obtained with these models differ from those from other models obtained with a simplified model of the Martian atmosphere (single composition, single thickness, no time dependence) and with a regolith-only (no-volatiles) surface model [18]. This Mars Radiation Environment Model will be tested against spacecraft data (e.g. Co-Investigator in the LIULIN-PHOBOS onboard the PHOBOS-GRUNT spacecraft from the Russian Space Agency KKA).

Conclusions: Models for the radiation environment to be found on the planet Mars have been developed. Primary particles rescaled for Mars conditions are transported through the Martian atmosphere, with temporal properties modeled with variable timescales, down to the surface, with altitude and surface backscattering patterns taken into account. The work is being extended to the Phobos and Deimos surfaces. The Mars Radiation Environment Model will be tested with the data from spacecraft instruments in the future.

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TIME-DEPENDENT MODELS FOR THE RADIATION ENVIRONMENT OF THE MOON.

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Introduction: The radiation protection is one of the two NASA highest concerns priorities [1]. In view of manned missions targeted to the Moon [2], for which radiation exposure is one of the greatest challenges to be tackled [3], it is of fundamental importance to have available a tool, which allows the determination of radiation fluxes and doses at any time on, above and below the lunar surface [4]. With this goal in mind, models of radiation environment due to Galactic Cosmic Rays (GCR) and Solar Particle Events (SPE) on the Moon have been developed, and fluxes and spectra hereby computed [5]. The work is described [6] as models of incoming cosmic ray [7-9] and solar primary particles [6] impinging on the lunar surface, transported through the subsurface layers, with backscattering taken into account, and interacting with some targets described as material layers. Time dependent models for incoming particles for both GCR and SPE are those used in previous analyses as well as in NASA radiation analysis engineering applications [10]. The lunar surface and subsurface has been modeled as regolith and bedrock, with structure and composition taken from the results of the instruments of the Luna, Ranger, Lunar Surveyor and Apollo missions, as well as from groundbased radiophysical measurements (see discussion in [4-6], [10]). The lunar-like atmosphereless body surface models are used to develope radiation models for the surfaces of the Martian satellites Phobos and Deimos [11].

Results: In order to compare results from different transport techniques, particle transport computations have been performed with both deterministic (HZETRN) [12] and Monte Carlo (FLUKA) [13] codes with adaptations for planetary surfaces geometry for the soil composition and structure of the Apollo 12 Oceanus Procellarum landing site [14,15], with a good agreement between the results from the two techniques [6,10]: GCR-induced backscattered neutrons are present at least up to a depth of 5 m in the regolith, whereas after 80 cm depth within regolith there are no neutrons due to SPE [6,10]. Moreover, fluxes, spectra, LET and doses for many kinds of particles, namely protons, neutrons, alpha particles, heavy ions, pions, muons etc., for various other lunar soil and rock compositions have been obtained with the deterministic particle transport technique [6]. The results from this work can only be compared in literature with previous versions of the same models or with very simplistic models [4-6,10], as also mentioned in [16]. This Moon Radiation Environment Model will be tested against spacecraft instruments data (e.g. Co-1 in the RADOM investigation onboard the CHANDRAYAAN-1 spacecraft from Indian Space Research Organization ISRO) in the near future.

Conclusions: Models for the radiation environment to be found on the Moon (on, above and below the surface) due to GCR, SPE and backscattering effects have been developed. A good agreement has been found between results from deterministic and Monte Carlo transport techniques. The quite large differences in the time and effort involved between the deterministic and Monte Carlo approaches deeply favor the use of the deterministic approach in computations for scientific and technological space radiation analysis. The work is being extended to the Phobos and Deimos surfaces. The Mars Radiation Environment Model will be tested with the data from spacecraft instruments in the future.

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LATE AMAZONIAN GLACIATION AT THE DICHOTOMY BOUNDARY ON MARS: EVIDENCE FOR GLACIAL THICKNESS MAXIMA AND MULTIPLE GLACIAL PHASES. J. L. Dickson¹, J. W. Head², D. R. Marchant¹. ¹Dept. of Geol. Sci., ²Dept. of Earth Sci., Brown Univ., Providence, RI 02912 USA, james_head@brown.edu

Introduction and Background: Recent exploration has revealed evidence for ice-related deposits in non-polar regions of Mars, including tropical mountain glaciers (e.g., [1]), mid-latitude ice-rich and glacial-like flows (e.g., [2]), and latitude-dependent dust-ice mantles (e.g., [3]). Deposition of significant quantities of ice down to tropical latitudes in the past history of Mars is consistent with predictions of spin-axis and orbital variations (e.g., [4]) and atmospheric general circulation models during such periods (e.g., [5]).

The most well-defined portion of the mid-latitude dichotomy boundary occurs as a complex set of scarps, reentrants, and isolated massifs where the boundary extends into northern mid-latitudes (~25-47°N), between 0° and 80°E. A considerable portion of the boundary (northern part of Arabia Terra) is comprised of fretted terrain and fretted channels and is the site of linedate valley fill (LVF) and lobate debris aprons (LDA) [6]. The LDA and LVF are characterized by lobate outlines and very low slopes and thus have historically been interpreted to be related to viscous flow of debris lubricated by groundwater or ice, with ice thought to have been supplied by vapor diffusion and condensation in talus (e.g., [6]).

Recent studies [7-10] interpret LVF/LDA as representing integrated glacial landsystems (e.g., [11], implying that: 1) the muted patterns that we see today are relict patterns and represent only a fraction of the original system; 2) the surface has experienced modification since the time of maximum glaciation (e.g., retreat, sublimation, periglacial, elolian, mass wasting, etc.); 3) exposed ice no longer remains on the surface, but in places may remain protected under thick layers of sublimation till; 4) convolved in the observed patterns is the record of glacial retreat and downwasting (trimlines, flow reversals, moraines, high stands, etc.); and 5) superposed on these patterns is a record of a) subsequent crater impact and erosion, and b) recurrent glacial phases. Major outstanding questions include: 1) What was the original extent of glacier ice, both areally and vertically? 2) On the basis of this information, what are the original volumes of ice transported to, and deposited in, these areas? 3) Is there evidence for multiple, recurring phases of glaciation? We report here on an example from Protonilus Mensae, near Coloe Fossae, in which we observe evidence of both: 1) former high-stan elevations of LVF about 920 meters above the present level, and 2) distinct lobate valley glaciers clearly superposed on, and cross-cutting, linedate valley fill.

Evidence for the former extent of LVF: Detailed reconstructions of flow directions in Protonilus Mensae and their relationships to topography revealed an anomalous area in which a 5 km wide, loop-like lobe extends from the LVF on the valley floor up into an elevated box canyon [12] (Fig. 1a, c). This is opposite the situation usually encountered, in which flow commonly extends from high-elevation alcoves out onto low-lying areas in lobe-like configurations (Fig. 1d). On the basis of the map view of the lobe morphology, the lobe appears to have been an extension of flow from the central part of the valley into this alcove, compressing and undergoing folding as it abutted the wall (Fig. 1b, c). MOLA profiles show that the current distal edge of the LVF lobe surface at the highest point in the alcove is at about -800 m, about 920 m above LVF in the adjacent valley, which itself is ~1300 m above the surrounding plains to the west (Fig. 1b).

The potential implication of this geometry, and the ~920 m elevation difference between the lobe and adjacent LVF, is that the isolated lobe represents a remnant of the former lateral extent of LVF that would provide an estimate of the vertical downwasting (thickness of the ice implied by the elevation difference) during retreat. A test of this interpretation is to extrapolate the elevation of the inferred limit elsewhere in the area and look for additional evidence of an upper limit. To do so, we extrapolated the ~800 m contour line (orange contour in Fig. 1b) across the mapped area and found that it corresponded to linear ridges situated at the margins of the main plateau and outlying massifs. For example, the southern and western margins of massif A (Figs. 1a,b) contain a series of approximately horizontal ridges that we interpret as marginal moraines, formed by debris that accumulated along ice margins. The ridges all occur within a narrow band restricted to a few hundred meters above and below the -800 m contour.

Predicted Behavior During Glacial Recession: On the basis of these observations, our working hypothesis is that glaciation in this area was more extensive during its peak development than is currently recorded in the remnant LDA/LVF textures alone. The current configuration of the LDA/LVF (Figs. 1a,b) essentially represents a snapshot of the terminal stages of this particular glaciation, i.e., at a point where downwasting dominated.

What is the predicted behavior of ice flow between peak development and the terminal snapshot? Typically, glaciation is a dynamic process whereby advance occurs when accumulation exceeds ablation, and recession occurs when accumulation is less than ablation. On Earth, continental-scale glaciation during the last glacial maximum in the High Arctic gave way to plateau icefields and ultimately to valley glacial landsystems [11].

In the Coloe Fossae region, we predict a similar transition, with the valleys initially filled with flowing glacial ice (at or above the -800 m contour). During this phase, lobes of glacial ice extended from the accumulation zones into the surrounding valleys (see diverging arrows in Fig. 1b) and then out into the surrounding lowlands; one of these lobes flowed into the elevated box canyon, forming the isolated lobe (Fig. 1). As this phase of extensive glaciation drew to a close, accumulation decreased and ice retreated from the northern lowlands back toward the dichotomy boundary. Even though glacial flow continued, the glacier surface lowered, exposing more and more mountainous topography. The exposed topography began to alter glacial flow patterns, and with continued lowering, local topography (alcoves and small valleys) played an ever-increasing role in influencing flow patterns. Furthermore, newly exposed bedrock provided a source region for supraglacial debris, particularly along the steep alcoves. As debris falls on ice it contributes to a transition from ice-dominated glacier surfaces to debris-covered glaciers. We suggest that as the regional glacial landsystem evolved and alcove-focused debris-covered glaciers began to dominate, new patterns of glacial flow evolved, including single to multiple sources, and in some cases reversal of flow direction.

In the case of the loop-like lobe in the box canyon, such late-stage transitions are evident. A reconstruction of the original configuration near peak glaciation shows the lobe extending into the valley, compressing to form the ridges, and stagnating there, while the major part of the flow continued down the main trunk valley just to the north. As the glacial system began to recede, the level...
of ice in the main valleys lowered and left the upper part of the loop-like lobe at the -800 m elevation level. This is likely to have encouraged reverse flow, for which there is evidence in the flow patterns of the loop-like lobe. As the glacial surface continued to recede in elevation, local sources of ice and debris-covered glaciers began to dominate. This is most clearly seen at the eastern part of the loop-like lobe, where a series of late stage lobes descend from the valley walls and converge with the main valley glacial flow, deforming the loop-like lobe in the process. Lobes become increasingly compressed into chevrons until they are completely integrated into the lineated valley fill. This current configuration is a snapshot of the final stage of glacial flow, when the glacier geometry no longer provided the forces necessary for glacial movement.

Evidence for Post-LDA/LVF Glacial Activity: Evidence for renewed glacial conditions subsequent to LDA/LVF formation is seen in the form of features interpreted to be lobate debris-covered glacial remnants (Fig. 2). In several cases, lobate flow-like features form within valley tributaries in alcoves, extend downslope, and emerge out onto the valley floor. Clear relative age relationships are seen (Fig. 2): the lobes are superposed on and cross-cut the linear trends in the LVF. Evidence for downwasting and decay of this lobe is seen in the position of marginal moraines and/or trimlines (inset in Fig. 2). In summary, the clear superposition of this and several other lobes (and their lateral and distal moraines) emerging from alcoves out onto the current surface of the lineated valley fill indicates that a separate, but less extensive phase of glacialiation followed the lowering of the LDA/LVF valley glacial landsystem.

Conclusions: A distinctive lobe of lineated valley fill in a box canyon is interpreted to represent a former glacial highstand, providing evidence that the current surface of adjacent LVF, about 1500 m above the plains, was at least 920 m higher during peak glacialiation. The present flow patterns provide insight into the latest stages of an originally much thicker and more areally extensive glacial landsystem; the peak thickness of the glacial system was possibly as much as 2.5 km. Evidence for a renewed, but much less extensive phase of glacialiation is seen in superposed lobes interpreted to be debris-covered glacier remnants.


Figure 1. Plateau and adjacent isolated massifs (east-central part of Fig. 1). LDA/LVF form distinctive patterns in tributaries and on valley floor. a) MRO CTX image POI_001570_2213_XI_41N305W (same in all figures). b) Topographic map; 100 m contours. Arrows represent flow patterns mapped in LDA-LVF. Orange line is the -800 m contour (the estimated uppermost position of the loop-shaped lobe). c) Loop-shaped lobe entering into the box canyon, with ice sourced from the main trunk valley (see Data Repository, Fig 3, for detailed patterns). d) Tributary debris-covered valley glacier emerging from the northern slope of massif A and joining a main trunk valley. Note the evidence for flow patterns emerging from the upper part of the valley, coalescing downslope, and producing a lobate flow feature as it emerges from the tributary out onto the valley floor. Note also the distinctive differences between the patterns associated with flow into the box canyon (a) and out of the tributary (b).

Figure 2. Debris-covered, glacier-like lobe emerging from a tributary on the southern wall of massif A (Fig. 1a; also see image in 2a). a) MRO CTX image. b) Sketch map showing main features and relationships. Note that the lobe extends out onto the floor, cross-cutting linear trends in the LVF, and appears topographically superposed on the LVF. The inset shows evidence for lateral moraines along the tributary wall, suggesting the progressive loss of ice down to the present topographic level.
ESTIMATION OF SURFACE ALBEDO VARIATIONS FOR THE PURPOSES OF RELIEF RECONSTRUCTION. I. A. Dulova, S. I. Skuratovsky, Yu. V. Kornienko, N. V. Bondarenko, Institute of Radiophysics and Electronics, National Academy of Science of the Ukraine, 12 Ak. Proskury, Kharkov, 61085, Ukraine. (id@ire.kharkov.ua)

**Introduction.** Images of planetary surfaces can be used as a source for high-resolution topographic information (limited by the resolution of images under study) since apparent surface brightness is controlled by surface slopes. Some methods were developed to derive elevation data from images, for example, traditional photoclinometry (e.g., [1]) and, the most rigorous, photometric method proposed in [2].

Correct relief reconstruction requires at least two images of surface obtained at different illumination directions. The most preferable for reconstruction of relief is a pair of illumination directions orthogonal to each other [3]. Such a pair of observations is sufficient for relief reconstruction, if the surface is photometrically uniform.

Real surface usually contains terrains with different photometrical properties. Nonuniform photometric properties like surface albedo and photometric function do not allow true surface topography reconstruction even under the preferable geometry of experiments.

In the present work we propose an approach for determination of surface albedo that can be used as complementary procedure for correct relief reconstruction.

**Optimal observational conditions for surface albedo determination.** Generally, observed surface brightness in location with coordinates \((x, y)\) (components of a 2D vector \(r\)) in the \(j\)-th image from the set under study depends on surface photometric properties and surface normal vector \(n\) as

\[
I_j(r) = F_j(n)A(r), \tag{1}
\]

where \(F_j\) – photometric function, and \(A\) – albedo of surface which depends on surface coordinates and does not depend on surface slopes.

For small surface slopes, the surface brightness \(I_j(r)\) can be expanded into a Taylor series on \(n\) in the vicinity of the surface mean normal vector \(n_0\). With two series terms it can be presented as

\[
I_j(r) = [I_{0j} + c_j(n - n_0)]A(r) = [I_{0j} + c_j\nabla H(r)]A(r), \tag{2}
\]

where \(c_j\) is the 2D constant vector controlled by surface properties and viewing geometry, \(\nabla H(r)\) is a 2D elevation gradient, and \(I_{0j}\) - brightness components corresponded to the general surface slope.

If the elevation gradient is constant over the surface presented in the image, observed variations of the brightness are related to true variations of surface albedo. Usually, surface topography causes deviations of particular elevation gradients from the general surface slope. It leads to variations of observed brightness over the areas with the constant albedo.

Considering small surface slopes deviations, unknown surface albedo can be derived using two observations of the surface illuminated in the opposite directions. In these cases changes of observed brightness caused by surface slopes \(I(r) = c_j\nabla H(r)\) are expected to have opposite signs.

Thus brightness of two images can be presented, respectively, as:

\[
\begin{align*}
I_1(r) &= (I_{01} + I_j(r))A(r), \\
I_2(r) &= (I_{02} - I_j(r))A(r), \\
\end{align*}
\tag{3}
\]

where \(I_j(r)\) is the brightness component depending on slope changes, \(I_{01}\) and \(I_{02}\) are the brightness components related to the general slope in the images.

Using Eq. (3) surface albedo can be derived as

\[
A(r) = (I_1(r) + I_2(r))/(I_{01} + I_{02}). \tag{4}
\]

Unknown values of \(I_{0j}\) and \(I_{02}\) can be approximated through the average image brightness, respectively, with expressions

\[
\begin{align*}
<I_1> &= I_{01} - <A>, \\
<I_2> &= I_{02} - <A>, \\
\end{align*}
\tag{5}
\]

where \(< >\) denotes an average over the surface.

Finally, distribution of surface albedo can be calculated as

\[
A(r) = (I_1(r) + I_2(r))/(<I_1> + <I_2>). \tag{6}
\]

\(<A>\) used in Eq. (6) is unknown value. So albedo variations can be only calculated with accuracy to an unknown multiplicative constant.

This unknown multiplier has no effect on relief reconstruction procedure with photometric method. For other applications some additional albedo information has to be involved. It could be, for example, estimates of albedo in individual surface locations using other techniques or an albedo map for the same surface with lower resolution.

**Test for the surface albedo determination.** To illustrate the approach for surface albedo determination
we simulated Lambertian surface with lunar-like cratered topography. Slopes of the test surface shown in Fig. 1 vary between 0 and 34° as for real planetary surfaces. We simulated surface albedo values by product of sine functions:

\[ A(x, y) = 0.5(1 + \sin(kx)\sin(ky)). \] (7)

Since Lambertian surface means that behavior of surface photometric function describes by Lambert law, only illumination geometry has to be taken into account. An example of initial image of left-illuminated surface with model surface albedo is presented in Fig. 2. Direction of illumination is shown with white arrow. Incidence angles for both images under study (left- and right illuminated) were chosen to be 35.5°.

The image presented in Fig. 2 has no noise component. With the absence of the noise in the source images, the surface albedo can be found rather precisely. Results including calculated albedo along with corrected left-illuminated initial image are shown in Fig. 3a, b, respectively. Corrected image in Fig. 3b corresponds to an ideal photometrically uniform surface.

Similar results obtained using source images with signal-to-noise ratio of 8 are presented in Fig. 4a, b. In this case elimination of albedo-related component of the image brightness is worse.

**Discussion and conclusions.** Computer experiments described above show that proposed method for determination of surface albedo from images is promising. Photometric behavior of many planetary surfaces like lunar or Mars surfaces can be approximated by Lambert law. In these cases for albedo determination two images of surface taken under opposite illuminations directions and at similar incidence angles of radiation need to be used. Such a pair of observations differs from the pair of surface observations commonly used in photogrammetry practice when two images have to be taken from different viewing directions under similar illumination.

Strong restriction for the use of proposed method is considered small surface slopes. Therefore the method cannot be used in the vicinity of features characterized by steep slopes. For surfaces with steep slope-features the proposed method for albedo determination can be used together with, for example, photogrammetry.

Observations of the surface illuminated in three directions different from each other by 90° at similar incidence angles are enough for rather correct reconstruction of the both surface topography and surface albedo variations.

SEASONAL DYNAMICS OF THE WATER ICE ON THE SURFACE OF THE NORTHERN POLAR CAP OF MARS BASED ON THE OMEGA DATA
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Introduction: First results obtained from OMEGA [1] data analysis [2, 3, 4] show that this data are very powerful tool for studies of the seasonal water cycle in the atmosphere-ice-regolith system and for understanding of the details of the current Martian climate. In the work we focused on the variations of the water ice spectral indexes on the surface of the Northern polar cap during the aphelion season. With the goal to study the dynamics of the structural changes on the ice cover of the polar cap during different intervals of the summer season we have conducted the mapping of the water ice spectral signatures corresponding to the band 1.2 μm and 1.5 μm.

Data analysis: Since the working range of the instrument includes series of notable absorption bands of CO 2 and other minor atmospheric components (H 2O, CO etc.), for the analysis of the surface spectra it is necessary first of all to eliminate the atmospheric contribution from observed spectra. For this we have calculated spectral transmittances taking into account observation geometry for each point of view. Temperature profiles and surface pressure values from the European Mars Climate Database [5], a comprehensive climate model taking into account atmospheric circulation and MOLA topography [6], were adopted. We took into account the non-uniform water vapor distribution in the atmosphere column. While zonal-averaged water vapor content from Smith et al. [6] has been adopted, zonally variable scale height of water vapor was simulated using GFDL’s, MGCM [8]. Calculated atmospheric transmission was convolved with the point spread function of the instrument and was used for removing atmosphere contribution from observed spectra.

For mapping of the absorption strength of the water ice on the Martian surface we have calculated the spectral indexes, characterizing the relative depth and shape of spectral bands specific for water ice in the each pixel of OMEGA surface footprint. For broad bands of water ice adsorption (at 1.25, 1.51, and 2.0 μm) we used the ratio of squares inside of the bands for more efficient measure of ice content than relative reflectivity.

Results: It is well known that the water ice (or snow and frost) spectra very sensitive to the grain sizes of the ice [7, 9]. At increasing of the grain size the broad band near 1.25 μm becomes deeper and 1.5 and 2.0 μm bands becoming much wider and less deeeper. The older snow is more coarse-grained in contrast to freshly deposited fine-grain frost. So, the bands are useful for study of the dynamics of the ice structure during summer season. At the fig.1 one can see evolution of the spectrum of the small region of the northern polar cap during different intervals of the summer period (mark by red pointer at the figure 2b). For the spectrum we can approximately estimate the mean grain size of the ice grains by comparison of the observing spectra with the theoretical spectra from [7, 9]. Such comparison show that the blue spectrum on the fig.1 may be corresponding to the relatively small sizes of ice grains (~ 50 μm), the brown spectrum – to greater ice grain size range (100-150 μm), and green – to the biggest ones (~ >200 μm).

Fig. 1. OMEGA’s spectra of one small fixed Northern polar cap region during different intervals of the summer season. From top to bottom: Ls~94, 114, 130.

The results of the selected water ice bands mapping are shown on the fig. 2 and 3. Figures were compiled for 3 periods of the summer: Ls~93-96 (using 886-922 orbits of MEX), Ls~113-115 (1048-1060 orbit) and Ls~122-134 (1177-1212 orbits). On the figures one can see distinctive changes of the water ice bands intensity during different sub-seasons of the first summer of Mars-Express operation. All these pictures were compiled in the same scale. It was found that the band depths for 1.25 μm and 1.5 μm, being shallow in the beginning of the summer, are becoming deeper in the later sub-season of the summer. This increasing of the water ice indexes during summer season is indicative of the ice grain size changes. Also we can see the evident zonal wave-3 and wave-4 features in the frost deposition and it’s evolution during the aphelion season. This phenomenon is the result of wave character of water vapor redistribution around the northern pole.
during summer. The received results support the manifestation of the noticeable contribution of stationary and quasi-stationary planetary atmospheric waves and are consistent with simulations of the Mars water cycle.

**Conclusions:** Our mapping of two selected spectral water ice bands in the different sub-seasons of the summer season shows very distinctive dynamics of the water ice microstructure changes on the surface of the Northern polar cap. This dynamics may be resulted by such factors as ice grains size change due to sublimation of the smaller ice grains, and ice’s metamorphism and increasing of the dust component at summer-time sublimation of the surface ice.

Moreover, these results are in good accordance with the GSM model.

**Acknowledgments:** This study was supported by the Russian Foundation for Basic Research (project N 06-02-16920a)


![Fig. 2. Maps of the water ice spectral index for 1.2 µm band during different intervals of the Northern summer. a) Ls~93-96 (886-922 orbits of MEX); b) Ls~113-115 (1048-1060 orbits of MEX); c) Ls~122-134 (1177-1212 orbits of MEX).](image)

![Fig. 3. Maps of the water ice spectral index for 1.5 µm band during different intervals of the Northern summer. a) Ls~93-96 (886-922 orbits of MEX); b) Ls~113-115 (1048-1060 orbits of MEX); c) Ls~122-134 (1177-1212 orbits of MEX).](image)
SEDIMENTARY FAN DEPOSITS IN JEZERO CRATER LAKE, IN THE NILI FOSSAE REGION, MARS: METER-SCALE LAYERING AND PHYLLOSILICATE-BEARING SEDIMENTS. C. I. Fassett1, B. L. Ehlmann1, J.W. Head2, J. F. Mustard2, S. C. Schon3 and S. L. Murchie2, 1Dept. of Geological Sciences, Brown Univ., Providence, RI 02912, 2JHU/Applied Physics Laboratory, Laurel, MD 20723. (Caleb_Fassett@brown.edu).

Introduction: Valley networks on Mars are generally believed to have formed via fluvial erosion [e.g., 1], and are thus commonly invoked as geomorphological evidence for an active water cycle on Mars early in its history [e.g., 2]. It has long been assumed that when valley networks were active, they must have transported significant amounts of sediment; however, terminal sedimentary deposits related to valley network activity are unusual and have only been recognized in a few locations [see 3 for a recent catalog].

An exceptional example of sedimentary deposits which were transported by valley networks are found in a ~40-km diameter crater northwest of the Isidis Basin, near the Nili Fossae, which has recently been assigned the name "Jezero crater" (centered at 77°40' E and 18°25' N) (Fig. 1). As originally described by Fassett and Head [4], two terminal fans debouch into the crater from the west and north, and on the crater's eastern end, an outlet valley is observed. Based on the present elevation of the outlet channel, for water to spill out of the crater it must have been filled with a sizeable lake, with a volume of at least ~250 km³, comparable to terrestrial Lake Tahoe (V~160 km³) or Erie (V~480 km³). Moreover, since the outlet valley has a minimum elevation comparable to or higher than the top of the present fan deposit [4], the fans found in Jezero were interpreted to have been deposited as deltas into this lake.

The Mars Reconnaissance Orbiter (MRO) instrument suite provides valuable new information about the fan deposits and Jezero crater, since it allows for further examination of the geologic setting of the crater with CTX, its sub-meter scale geology with HiRISE, and its mineralogical/spectroscopic signatures with CRISM. Using this new data, as well as information derived from instruments on MGS, Mars Odyssey, and Mars Express, we have begun a new examination of Jezero crater to test the hypothesis that the fans were deposited into a standing body of water, as well as (1) develop a stratigraphic model for the fan deposits and their surroundings; (2) learn about the mineralogical characteristics of the putative delta sediments; and (3) to characterize the Jezero crater deposits and environment as a potential landing site for the Mars Science Laboratory (MSL) or for a reference human mission as part of HEM-SAG.

Characteristics of the fan deposit and its surroundings: Finely Layered Material and Depositional Relationships. Initial observations of layering in the fan deposits focused on the margin of the present structure, where decimeter-scale layering was clearly evident in MOC data on both the northern and western deposits [4]. However, HiRISE allows observation of layered materials at the sub-meter-to-meter scale; such deposits are common in portions of the fan deposits and their surroundings.

In specific locations, the finely-layered material is richly structured (Fig. 2). Key observations include (1) layers appear to dip gently (<10-15°); (2) layering is locally-continuous over 500-m length scales, often outcropping in a sinuous pattern; (3) unconformities are discernable between packets of layered material. These observations are consistent with the emplacement of all the finely-layered material by lateral accretion during meander migration. The unconformities in the deposit imply that deposition was laterally discontinuous across the fan surface, and suggest that the fan surface underwent numerous channel switching events. The meandering patterns that are observed on the fan deposit imply that they were deposited on the subaqueous portion of a delta, into a lake with a relatively-stable base level, based on terrestrial analogues (see companion abstract, [5]).

Massive Elongate Deposits. Along with outcrops of fine-layering on the fan, elongate deposits are found on the fan surfaces. These massive materials were originally interpreted as topographically-inverted, remnant channel pathways on the basis of MOC data [4]. This more massive unit tends to be locally high and have steep margins and sometimes appears lenticular in form. HiRISE observations reveal that this unit is commonly weathering into large boulders or blocks. The elongate ridge deposits appear to be stratigraphically higher than the finely-layered materials described above. It seems likely that the reason the elongate material is locally high is that it was more resistant to billions of years of aeolian erosion than the weaker, finely-layered sediment which is finer-grained ("survival of the coarsest"). The relationship between individual elongate deposits on the fan surface is complex, but visible cross-cutting relationships suggest that they were emplaced independently in a series of sequential depositional episodes (Fig. 3).

Multiple hypotheses continue to be considered for the origin of this massive material that is the upper unit of the fan. The hypothesis suggested by earlier observations [4] that these massive materials were channel bed and levee deposits composed of comparatively coarse-grained material continues to be viable. In this view, the depositional fan aggraded and prograded over time, consistent with sediment in the fan coarsening upward. However, an alternative hypothesis is that the resistant massive upper units of the fan are volcanic and related to the resistant, smooth unit on the crater floor, described below.

Smooth Crater Floor Unit. Across much of the floor of Jezero crater is a resistant smooth unit, with arcuate, lobate margins. It is consistently stronger than material in its surroundings based on preservation of small craters, morphology of craters where impacts occurred into the contact with its surrounding, and its comparative resistance against erosion. It seems probable that this material is volcanic, potentially emplaced in a mare-like, low-viscosity flow, similar to Hesperian Ridged Plains (HR) found elsewhere in this region. At present, the source for this smooth unit is not apparent, though difficulty pinpointing a clear source for mare-like volcanic flows in craters is not unusual.

Mineralogy: A diverse mineralogical assemblage was discovered in the Nili Fossae region using the OMEGA instrument aboard Mars Express, including phyllosilicates within the watershed of Jezero crater [e.g., 6]. Recent studies have detailed the stratigraphic context for the phyllosilicates and other distinct units based on mineralogy that are found in the Nili Fossae region [7,8].
Targeted observations with CRISM build upon these OMEGA observations, and provide a high-resolution, visible/near-infrared hyperspectral picture of Jezero crater, the fan deposits, and its surroundings [9]. Although a full examination of these data are beyond the scope of this abstract, key observations include: (1) the finely-layered sediments of the fan (Fig. 2) have an iron-magnesium phyllosilicate component; (2) both the massive upper unit of the fan and the smooth unit on the crater floor are essentially spectrally neutral, an observation which motivates the suggestion that these materials may be of similar origin; (3) where dunes are visible, their VN-IR spectra often match olivine; (4) phyllosilicates are also found on the Jezero crater interior and floor off the margin of the fan itself.

Landing Site Analyses: Sedimentary fan deposits were suggested as a class of landing site for MSL in several independent abstracts, two of which included the Jezero fan deposits [10]. This is motivated by the fact that sediments have the potential to preserve an excellent record of the surface environment at the time of their emplacement, including information about habitability that lies at the core of the goals of the MSL mission and Mars program as a whole. The discovery of phyllosilicates in the Jezero crater fan deposit only increases the attractiveness of this location as a potential landing site. Thus, we have begun initial examination of whether a site inside Jezero meets the MSL engineering requirements. In a 20-km diameter circle on the smooth terrain east of the fan deposits, we have extracted MOLA point-to-point slopes at relevant baselines for MSL; results of this examination are summarized in Table 1.

Conclusions: New data from MRO and other missions continue to provide support for the model that the sedimentary fan deposits in Jezero crater are valley network sediments, emplaced into a standing body of water in the crater, as originally suggested by [4]. Indeed, recognition of clear sedimentary signatures such as the finely-layered material that we interpret as lateral accretion deposits increase our confidence in this model. Further analysis is necessary to understand the origin of the upper, coarse-grained elongate units of the fan material.

Both the geological observations which suggest that Jezero crater was once a deep lake and the recognition by CRISM that Jezero sedimentary fan deposits contain phyllosilicates makes it an important potential target for continuing study and in situ exploration. Phyllosilicates are thought to have formed only early in Mars’ history when water was comparatively more abundant at the surface or in the near-surface [11]; as such, study of phyllosilicate-bearing materials should inform us about the habitability of the early surface environment of Mars and are therefore a key target for MSL and other future missions.


<table>
<thead>
<tr>
<th>Baseline</th>
<th>2-5 km</th>
<th>200-500 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Constraint</td>
<td>&lt; 3°</td>
<td>&lt; 5°</td>
</tr>
<tr>
<td>Observation</td>
<td>99.84% &lt; 3° (along track)</td>
<td>99.91% &lt; 3° (between tracks)</td>
</tr>
<tr>
<td>300-m scale:</td>
<td>99.57% &lt; 5°</td>
<td>99.57% &lt; 5°</td>
</tr>
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Fig. 1. Map of Jezero crater + fan deposits, after [4].

Fig. 2. Portion of HiRISE image PSP_002387_1985 of layering on fan; this outcrop is phyllosilicate-bearing (see text).

Fig. 3. Portion of CTX image p02_001820_1984 (left) and map of elongate lobes (right).
THE STARUST COLLECTION OF COMET 81P/WILD 2 PARTICLES: A COMPARISON WITH VEGA AND GIOTTO RESULTS FOR 1P/HALLEY. G. J. Flynn, Department of Physics, State University of New York – Plattsburgh, 101 Broad St., Plattsburgh, NY 12901 USA (george.Flynn@plattsburgh.edu).

Introduction: Particles emitted by comets are believed to be the same dust that accreted, along with ices as the Solar System was forming, ~4.58 billion years ago. This dust has been held in “cold storage” in the comets since their formation. Thus, comets are generally thought to have preserved a record of the processes and conditions in the Solar Nebula at the time of Solar System formation. This record is not available in bodies such as meteorites that experienced aqueous or thermal metamorphism.

In 1986 two VEGA spacecraft and the Giotto spacecraft flew through the dust coma of Comet 1P/Halley. Instruments on these spacecraft provided information on the size-frequency distribution and elemental composition of small dust particles.

NASA’s Stardust spacecraft flew through the dust coma of Comet 81P/Wild 2 on Jan. 2, 2004, collecting samples of Wild 2 dust by impact at ~6.1 km/sec into low-density silica aerogel and Al-foil [1]. These samples were delivered to Earth on Jan. 15, 2006.

A small fraction of the Wild 2 samples collected by Stardust were studied in the Preliminary Examination, a 7-month period that ended on August 15, 2006. Results obtained during the Preliminary Examination were reported by six teams of investigators who focused on craters [2], organics [3], isotopes [4], infrared spectroscopy [5], elemental composition [6], and mineralogy [7] of the Wild 2 samples.

Wild 2 Particles Collected by Stardust: Long-exposure images taken by the Stardust spacecraft during the encounter (Figure 1) show that Wild 2 is an active comet, with at least 20 areas on the surface emitting highly-collimated jets of gas and dust [8]. These jets are believed to originate in pockets of gas and dust in the interior of the comet. The Dust Flux Monitor Instrument, an active dust counter carried on Stardust, indicated that the spacecraft passed through several intense swarms of particles [9], some of which Sekanina et al. [10] associated with specific jets seen in the images. These results indicate that Wild 2 particles collected in the aerogel and Al-foil flown on Stardust spacecraft sample several different source regions in the interior of the comet. Thus, the collected particles are believed to be representative of the non-volatile component of Wild 2 over the size range that was sampled during the encounter.

Sample Sizes: The Stardust aerogel capture cells have a total collection area of 1039 cm² and the Al-foils have a total collection area of 152 cm² [2]. Particles striking the foils produced “tracks,” or damaged regions of aerogel, as they decelerated. Particles impacting the Al-foil produced craters. The 6.1 km/sec collection velocity can be replicated using laboratory light-gas guns. Calibration experiments indicate that the smallest craters were produced by particles having masses less than 10⁻¹⁵ grams, while the largest track corresponds to a particle having a mass greater than 10⁻₄ grams [2].

The mass-frequency distribution of the particles that impacted the Stardust collectors can be derived from the plot of the cumulative number of craters or tracks versus particle mass given by Horz et al. [2]. For LDEF impacts the mass flux increased from the smallest mass measured (10⁻⁹ grams) up to a particle mass of ~10⁻⁵ grams then decreased for particles up to the largest mass they observed (10⁻⁴ grams) [11].

Sample Heterogeneity: Chemical heterogeneity is an indicator of the size-scale of the individual mineral grains of a sample. The Elemental Composition Preliminary Examination Team measured element abundances in 23 whole tracks extracted from Stardust aerogel capture cells, randomly selected to span the size and morphological diversity seen on the collectors, and the elemental composition of the residue in 7 craters in the Al-foils [6]. Figure 3 shows the diversity in composition for S and for Ca of the 21 whole tracks for which S and Ca data were obtained. Both S/Fe and Ca/Fe show four order-of-magnitude or more variation in element/Fe ratios of the tracks, with more than an order-of-magnitude variation in
Comparison of Wild 2 with Halley: G. J. Flynn

Figure 2: The mass per mass decade incident on the Stardust collector, scaled to a 1 m² area, derived from Figure 4 of Horz et al. [2]. The sharp increase in the mass contributed by the largest particles indicates that the five particles >100 μm contain most of the mass collected at Wild 2, and that, unless they have exceptional compositions, the smallest particles cannot significantly perturb the mean composition.

Figure 3: Element/Fe ratios for S (top) and Ca (bottom) for the 21 whole tracks from the Stardust aerogel collectors for which S and Ca were both measured. The analyses are ordered with the measured Fe mass increasing to the right. The blue symbols are the individual track element/Fe ratios, the green line is the cumulative average composition, the blue line is the 21 particle mean composition, and the red line is the CI meteorite mean composition.

Figure 4: Mean CI normalized element/Fe ratios for the 23 Wild 2 tracks measured during Preliminary Examination [6].

the 3 largest particles measured. This indicates particles from Wild 2 are heterogeneous at the largest size scale measured (~25 μm). Nonetheless, crater composition measurements indicate that composite particles dominate at the smallest size (~100 nm) [2].

Mean Elemental Composition: The mean elemental compositions of the 23 tracks (Figure 4) and the 7 craters are consistent with the CI meteorite composition, suggested to be the mean composition of the Solar System because of its similarity to the composition of the Solar photosphere [12], for the refractory elements, although a small depletion in the Fe/Si ratio is indicated by the crater analysis [6]. Most of the moderately-volatile elements are enriched relative to the CI composition, although S appears to significantly depleted relative to CI.

Statistical analysis suggests that if we examined 23 different Wild 2 particles in the same size range (~2 to 25 μm) we would obtain a mean composition within our 2σ error range 95% of the time [6].

Comparison of Wild 2 to Halley: The slope of the cumulative size frequency distribution of the dust in the Halley coma is similar to that measured at Wild 2 [2]. The elemental composition determined from the Wild 2 samples is consistent with that obtained for comet Halley dust, but extends the measurement of comet composition to include moderately-volatile minor elements. The major differences between the Wild 2 results and the CI meteorite composition are for elements not well determined in the Solar photosphere, suggesting that CI may not represent the mean Solar System composition [6].

HABITABILITY OF SUPER-EARTH PLANETS. S. Franck\textsuperscript{1}, C. Bounama\textsuperscript{1}, W. von Bloh\textsuperscript{1}, and M. Cuntz\textsuperscript{2},
\textsuperscript{1-Potsdam Institute for Climate Impact Research, PF 601203, 14412 Potsdam, Germany (franck@pik-potsdam.de), 2-University of Texas at Arlington, Box 19059, Arlington, TX 76019, USA.

Methodology: Our numerical model couples the stellar luminosity, the silicate-rock weathering rate and the global energy balance to obtain estimates of the partial pressure of atmospheric carbon dioxide, the mean global surface temperature, and the biological productivity as a function of time. The main point is the persistent balance between the CO\textsubscript{2} sink in the atmosphere-ocean system and the metamorphic (plate-tectonic) sources. This is expressed through the dimensionless quantities
\[
\tilde{f} \equiv f_{\text{wr}}(t) \cdot f_{A}(t) = f_{\text{sr}}(t),
\]
where \(f_{\text{wr}}(t)\) is the weathering rate, \(f_{A}(t)\) is the continental area, and \(f_{\text{sr}}(t)\) is the spreading rate, which are all normalized by their present values of Earth. The evolution of the surface temperature can be derived directly for known luminosity, distance to the central star and geophysical forcing ratio (GFR=\(f_{\text{sr}}/f_{A}\)). For the investigation of a super-Earth under external forcing, we adopt a model planet with a prescribed continental area. The fraction of continental area to the total planetary surface is varied between 0.1 and 0.9.

The thermal history and future of a super-Earth has to be determined to calculate the GFR values. Assuming conservation of energy, the average mantle temperature \(T_{m}\) can be obtained as shown in Fig. 1.

The photosynthesis-sustaining HZ (pHZ) is defined as the spatial domain of all distances \(R\) from the central star where the biological productivity is greater than zero, i.e.,
\[
\text{pHZ} := \left\{ R \mid \Pi(P_{\text{atm}}(R,t), T_{\text{surf}}(R,t)) > 0 \right\}.
\]

Results and discussion: Very recently, Udry et al.\textsuperscript{[2]} announced the detection of two super-Earth planets in this system, Gl 581c with 5.06 Earth masses with a semi-major axis of 0.073 AU, and Gl 581d with 8.3 \(M_{\text{E}}\) and 0.25 AU. Both mass estimates are minimum masses uncorrected for the inclination term \(\sin i\).

According to Valencia et al.\textsuperscript{[1]} super-Earths are rocky planets from one to ten Earth masses with the same chemical and mineral composition as the Earth. We use scaling laws to obtain the total radius, mantle thickness and average density as a function of planetary mass. The pHZ around Gl 581 for super-Earths with five and eight Earth masses has been calculated for \(L=0.013L_{\odot}\) (von Bloh et al.\textsuperscript{[3]}).

The results are shown in Fig. 2. The planet Gl 581c is clearly outside the habitable zone. A planet with eight Earth masses has more volatiles than an Earth size planet to build up such a dense atmosphere. This prevents the atmosphere from freezing out due to tidal locking. In case of an eccentric orbit of Gl 581d (\(e=0.2\)), the planet is habitable for the entire luminosity range considered in this study, even if the maximum CO\textsubscript{2} pressure is assumed as low as 5 bar. In conclusion, one might expect that life may have originated on Gl 581d. The appearance of complex life, however, is unlikely due to the rather adverse environmental conditions. To get an ultimate answer to the profound question of life on Gl 581d, we have to await future space missions such as the TPF/ Darwin. They will allow for the first time to attempt the detection of biomarkers in the atmospheres of the two super-Earths around Gl 581.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{fig1.png}
\caption{Sketch of the thermal evolution and the scaling laws for super-Earth planets with radius \(R\) and mass \(M\), where \(\rho\) is the density, \(c\) is the specific heat at constant pressure, \(q_m\) is the heat flow from the mantle, \(Q\) is the energy production rate by decay of radiogenic heat sources in the mantle per unit volume, and \(R_{\text{in}}\) and \(R_{\text{c}}\) are the outer and inner radii of the mantle, respectively.}
\end{figure}
HABITABILITY OF SUPER-EARTH PLANETS: S. Franck, C. Bounama, W. von Bloh and M. Cuntz

Figure 2: The pHZ of Gl 581 for super-Earth Gl 581c (a) and Gl 581d (b) with a relative continental area varied from 0.1 to 0.9 and a stellar luminosity of 0.013 L$_*$ as a function of planetary age. The light colours correspond to a maximum CO$_2$ pressure of 5 bar, whereas the dark colours correspond to 10 bar. For comparison, the positions of Venus, Earth and Mars are shown scaled to the luminosity of Gl 581. The light and dark grey shaded areas denote different approximations for the so-called stellar HZ. The vertical bar at 2 Gyr denotes the range of distances due to the (possibly) eccentric orbit. The area below the solid black curve is affected by tidal locking [3].

References:
**Introduction:** The Bukhara carbonaceous chondrite fell in Uzbekistan on July 9th 2001, one fragment of mass of 5.3 kg being only found [1]. It belongs to the carbonaceous chondrites of the chemical CV-group, i.e. to the group of the Vigarano carbonaceous chondrite; the most famous and well-studied representative of that group is the Allende chondrite [2-4 et al.]. We have investigated the specimen N 16697 of mass of 320 g. The exposure age of the Bukhara chondrite is not measured yet. According to the density of tracks of VH-nuclei, the shielding depth of that sample from the pre-atmospheric surface of the chondrite ranged within \(d \approx 5 - 17\) cm [5].

Using the complex of low-level gammaspectrometric instruments of the laboratory of cosmochemistry of GEOKHI RAS [6], the contents of the following radionuclides are measured non-destructively in the Bukhara chondrite: \(^{54}\text{Mn} - 120 \pm 40, \(^{22}\text{Na} - 88 \pm 10, \(^{60}\text{Co} - 6 \pm 2\) and \(^{26}\text{Al} - 56 \pm 6\) in dpm/kg, respectively [7].

By means of the analytical method developed beforehand [8,9], modeling of the production rates of the radionuclides measured in the Bukhara chondrite is performed using the stratospheric data [10] on the average galactic cosmic ray intensity for \(~1.5T_{1/2}\) of the radionuclides before the fall of the chondrite to the earth. The analysis of the experimental data and regularities of the theoretical modeling has allowed us to estimate the pre-atmospheric size and ablation of the Bukhara chondrite, as well as the extent of its orbit.

**Pre-atmospheric size and ablation:** The most efficient approach to estimation of pre-atmospheric sizes of chondrites is a combination of the data on VH-nuclei track density (which are the most sensitive to the shielding depth of the sample from the surface) with the data on the \(^{60}\text{Co}\) content (which are the most sensitive to the sizes of the chondrites) [8,11-13]. Indeed, radionuclide \(^{60}\text{Co}\) is produced in the reaction of \(^{59}\text{Co}(n,\gamma)^{60}\text{Co}\) with thermal and resonance neutrons, the accumulation of which is very sensitive to the size of the bodies [8, 14]. It is obvious that the \(^{60}\text{Co}\) generation increases directly with the content of Co, which varies over a sufficiently wide range in the chondrites [15,16]. When modeling \(^{60}\text{Co}\) depth distribution in the Bukhara chondrite, we used the average content of Co of 0.06% in carbonaceous chondrites (in particular, in the Efremovka CV3-chondrite) [16]. The \(^{60}\text{Co}\) content, measured at the time of the Bukhara chondrite fall to the earth, was produced under the average galactic cosmic ray intensity \(I_0 \approx 0.2705\) cm\(^2\) s\(^{-1}\) sr\(^{-1}\) for \(~8\) years (i.e. about \(~1.5T_{1/2}\) of \(^{60}\text{Co}\)) before the fall [10,17]. It is clear that, using \(^{60}\text{Co}\), we obtain the estimate of the average pre-atmospheric size of the Bukhara chondrite for the last \(~8\) years before its entrance to the earth atmosphere. The results of the modeling are presented in Fig.1.
The upper plot shows the $^{60}$Co distribution in spherical chondrites of radii $R$, depending on the shielding depth $d=R-r$ of samples from the surface. The cross is the measured $^{60}$Co content in the Bukhara chondrite (6±2 dpm/kg) at the depth of the investigated sample of $d = 11±6$ cm, as identified by the track evidence. The cross corresponds to the average pre-atmospheric radius of the Bukhara chondrite of $R = 24^{+6}_{-3}$ cm, which is shown on the lower plot describing the dependence of $^{60}$Co distribution at various depth $d=R-r$ from the surface on radius $R$. Using the density of CV-chondrites of 2.95 g cm$^{-3}$ [18], one may obtain that the pre-atmospheric mass of the Kilabo chondrite (6 cm corresponds to the orbit with aphelion of $q' = 24^{+6}_{-3}$ cm) varies between 170 kg, and the ablation through the passage of the earth atmosphere amounts ~96.9%. Within the limits of errors, it is in accordance with the statistic estimates of the ordinary chondrite ablation [19].

Orbit: The previously elaborated isotopic approach [8,13,20], based on the content of cosmogenic radionuclide $^{26}$Al, is used to estimate the position of aphelia $q'$ of the chondrite orbit. According to that approach, the measured level $r$ and orbital period $P = 670.5$ days. This orbit as the regularity $r(t)$ (where $t$ is time and $r$ is a heliocentric distance), calculated with the Kepler formulae [21], is shown in Fig.2. It corresponds to the orbit of the majority of the chondrites, the maxima of the aphelion distributions of which fall on the range of $q' \sim 1.9-2.0$ AU [8,13]. Apparently, due to the increase of the population density of cosmic bodies near the inner boundary of the main asteroid belt, and, hence, due to their ever-growing fragmentation [22,23], the population peak of bodies of the ‘chondrite’ size (~1 m [13]) falls just on that range. It testifies to the universality of the mechanisms of evolution and selection of cosmic bodies in the solar system, the majority of which being stochastic processes of their collisions and resonance interactions with planets [13]. Unfortunately, any statistic data on the extent of orbits of the carbonaceous chondrites, as well as their direct measurements, are not available yet.

DETAILED STUDY OF TOPOGRAPHY AND MORPHOLOGY AND ESTIMATES OF THE TOTAL HORIZONTAL EXTENSION OF RIFT ZONES OF ATLA AND BETA-PHOEBE REGIONS, VENUS. E. N. Guseva, Vernadsky Institute, 119991, Moscow, Russia, guseva-evgeniya@ya.ru

Introduction. Rift zones of Venus are known since time of reception of results of radar survey by Pioneer-Venus Orbiter. It was found, that they resemble continental rifts of the Earth and form a global system of belts ~ 40.000 km [8, 9, 11]. Rifts of Venus were mapped and studied with the Magellan data, without division rift into age groups [10, 12], and then with their division on the younger and older groups [1, 5]. In [3 http://geo.web.ru/db/msg.html?mid=1169725*uri=fig_1.html; http://geo.web.ru/db/msg.html?mid=1169725*uri=fig_2.html], the global map of the Venusian rifts was presented. The map shows the distribution of the rifts in more detail than in [2]. In according with the earlier studies [e.g. 6], it was found that young rift zones occur mostly in equatorial area of the planet and correlate with positive anomalies of geoid. Topographically, young rift zones are mostly deep valleys [3] with elevated flanks.

Goals of the study. The study of young rift zones Atla and Beta-Phoebe (Fig. 1, 2) includes three tasks:
1) construction of a series of profiles for each of the three branches of the Atla and Beta-Phoebe rift zones; 2) identification of rift faults on the profiles and their mapping on the altimetry map; 3) measurements of vertical relief within rift zones, and estimates of the horizontal extension across the rifts.

Approach. The data for this study were collected from the Magellan topographical map resampled to the resolution ~5 km/px. The topographical characteristics of the major faults were taken from profiles and the horizontal extension across the rifts was estimated using the technique of Kiefer and Swafford [7].

Fig. 1. Radar altimetry map of Atla Regio with mapped rift zones.

Observations and results. 1) The rift zone of Atla Regio has the following characteristics: Three branches of rifts in this area start within the topographic rise and extend as swarms of curvilinear faults to the NW, SW and SE (Fig. 1). Topographically, rifts are usually depressions (Fig. 3) (rift valleys, profiles 19, 30-33), but sometimes the valley is absent and the faults are located on the topographic highs (profile 23). The valley flanks are usually elevated in relation to the background terrain. The rift flanks are sometimes at the same level but usually occur at different elevations: eastern flank is higher than western (profiles 19, 30-32) and vice versa (profiles 30, 32-33).

Fig. 2. Radar altimetry map of the Beta-Phoebe region with mapped rift zones.

Fig. 3. The profiles of rift-3 of Atla Regio.

The NW branch, rift-1 has a width of 325 to 470 km, and depth from the elevated flanks to the bottom of 2-5.5
km; The SW branch, rift-2 has a width of 275 to 450 km, and depth of 1.5 to 7 km; The SE branch, rift-3 (Fig. 3) has a width of 230 to 470 km and depth of 2.8 to 3.6 km.

For Beta-Phoebe region the following characteristics have been found: Like in Atla Regio, three branches of the rifts in this area start within the topographic rise and extend as swarms of faults to the N, W and S (Fig. 2). Topographically, they are usually depressions (Fig. 4) (valleys, profiles 4, 13-16) but sometimes the valley is absent and the faults are located on the topographic high (profiles 17-18). The valley flanks are usually elevated in relation to the background terrain. Their elevations are locally symmetrical (profiles 17-18) but typically asymmetrical: eastern flank is higher than western (profiles 13-14, 16) and vise versa (profiles 4, 15).

The N branch, rift-1 has the rift valley width 220 - 470 km, depth 2.8 - 6 km; The W branch, rift-2 has the rift valley width 230 - 395 km, depth of 1.7 - 3.2 km; The S branch, rift-3 Beta-Phoebe regio (Fig. 4) has a width rift 360 - 470 km, depth 1.2 - 5.8 km.

2) The mapping and analysis have shown, that rift valleys are outlined by large faults and series of smaller faults complicate the rift floor and flanks.

3) The total horizontal extensions for each rift estimated by a technique of Kiefer and Swafford [7] are as following. *Atla Regio*: 2.7 - 7.5 km for rift-1; 1.1 - 9 km for rift-2; 3.1 - 6.7 km for rift-3. *Beta-Phoebe regio*: 2.7 - 7.1 km for rift-1; 2.5 - 5.2 km for rift-2; 1.8 - 6.8 km for rift-3.

**Summary.** 1) On the basis of detailed study of topography of rifts *Atla* and *Beta-Phoebe* regions it was possible to conclude that the rifts are mainly depressions with the raised flanks.

2) In both regions for rift zones an asymmetry of the flank rises is a characteristic. More frequently, the western flank is higher than the eastern one both in Atla and Beta-Phoebe regions.

4) The rifts of Atla Regio and the Beta-Phoebe regions have approximately the same width: 276-463 km and 279-445 km, respectively.

5) Our estimates of total horizontal extension in the rifts are 2.3 to 7.7 km in Atla Regio and 2 to 7.1 km in Beta-Phoebe regions. If normalized on the rift zone width, the extensions are close to 1%. These estimates are practically identical to the 3 to 6.7 km estimates for Beta Regio made by Kiefer and Swafford [7], but are significantly lower than the 11.4 km mean estimate received by Connors and Suppe [4] using different technique.

**Future study.** We plan to extend our study of Atla and Beta-Phoebe regions making photogeologic analysis of the Magellan SAR images, and applying the technique of Connors and Suppe [4] to compare those new results with the earlier received data for rift zones of Atla and Beta-Phoebe regions. We also plan to compare the characteristics of the Atla-Beta-Phoebe rifts with the appropriate characteristics of the terrestrial East-African rift to look for similarities and distinctions between the Venusians and terrestrial rift zones.

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GEOLOGICAL EVOLUTION OF THE TERRESTRIAL PLANETS.  James W. Head, Department of Geological Sciences, Brown University, Providence, RI 02912 USA (james_head@brown.edu).

Introduction: In the 50 years since the launch of Sputnik, the first artificial Earth satellite, over a hundred missions have been sent to explore the terrestrial planetary bodies (Moon, Mercury, Mars, Venus), and the scientific results from these missions have radically changed our perspective on our own Home Planet, Earth. Here the major findings in this half century of exploration are reviewed, and evolving insights and themes are described.

Earth: In parallel with the advent of the space age, Earth scientists became increasingly aware that the Earth's lithosphere was not stable and fixed, moving predominantly vertically, but rather was segmented and mobile, forming at mid-ocean ridges, moving laterally at rates of cm/a, and flexing and subducting at convergent plate boundaries. This process, called seafloor spreading, continental drift and plate tectonics, synthesized many disparate observations into a compelling theory of the recent evolution of the Earth. Together with a new understanding of mantle plumes and hot spots, and how they operated together, Earth scientists had an impressive paradigm for Earth history. However, because of the dynamic nature of plate tectonics and erosional processes, 2/3 of the Earth's surface is less than 5% of the age of the planet, and the vast majority of Earth history, including its formative years, was not readily available for study. When did plate tectonics start? How and when did continents form? What was the early atmosphere like? When, and where, did life originate? These and other questions required an understanding of the major processes operating in the first half of Solar System history. Planets and moons were soon to join Earth as objects of geological interest and analysis.

Moon: The exploration of the Moon began to provide insights into the formative years of planetary history. Returned samples showed the very ancient age of the lunar highlands crust, and careful analysis of elemental abundances revealed that the highlands crust had formed from significant melting of at least the outer part of the Moon shortly after its accretion, during segregation and flotation of plagioclase in a "magma ocean". A fundamental link between the period of accretion of a planet and the formation of its crust and lithosphere had been made, bringing together accretionary theory and geological observations. Furthermore, analysis of isotopes and refined accretionary models showed that the Moon itself very likely formed from the impact of a Mars-sized object into the very early Earth. Clearly, the Earth had undergone a very traumatic event in its early childhood. Geophysical measurements and observations determined the thickness of the crust and lithosphere, and showed that the thermal evolution of the Moon had involved a rapidly thickening lithosphere that formed a single global plate (a "one-plate planet" in contrast to the Earth). This early stable global lithosphere then served to record the geological processes operating in the first half of Solar System history. Analysis of the geology of the Moon revealed that impact cratering was a fundamental process throughout lunar history; the geologic record showed an early period of crustal formation and stabilization, and disruption by a high, but declining, impact flux. Evidence emerged for a period of "Late Heavy Bombardment" in which a peak flux of projectiles may have formed much of the currently observed ancient crater population, in a period of otherwise monotonically declining impact flux. Geologists began to appreciate the singular and instantaneous effects of the formation of huge impact basins, many of which could easily resurface a lunar hemisphere. Impact cratering on Earth slowly began to be appreciated, not as an oddity, but as a fundamental process in early planetary history, and as a defining process in the biotic evolution of the Earth. Volcanic activity preferentially filled the nearside impact basins in the first third of lunar history, loading and flexing the thickening lithosphere, the efficient conductive heat loss permitted by the small size and the high surface-area-to-volume ratio (the "radiator" effect) of the Moon supported the final mare volcanic loads that are recorded by the mascons in the lunar gravity field. This early net state of contraction in the lunar lithosphere, further inhibited lunar volcanism. Lunar tectonism, then rather having a dominantly lateral sense as in Earth plate tectonics, was vertical, related to loading and planetary cooling. The last half of Solar System history on the Moon was relatively quiescent, with geological features related to exogenic impacts (e.g., Copernicus, Tycho) and the formation of an impact-produced regolith. Exogenically delivered volatiles may have accumulated in permanently shadowed polar regions.

Mercury: Mercury, slightly larger than the Moon, but containing a core of 50% of lunar size, was shown to be a one-plate planet, with impact craters and smooth/intercrater plains dominating the surface. The origin of Mercury's huge core, perhaps due to nebular temperature/pressure gradients during the period of planetary formation, or to early impact stripping of a once-thicker crust and upper mantle, remains a mystery to be explored by future missions. Recent Earth-based measurements are consistent with a molten core, raising the question of the dynamics of core-mantle heat transfer mechanisms, the nature and scale of mantle convection, and what evidence can be found in the geological record. Controversy surrounds the interpretation of the smooth and intercrater plains. Do they represent surface magmatic extrusion (volcanism), or alternatively, some process related to impact ejecta emplacement and flow? Huge scars attest to the global-scale contraction of Mercury early in its history. Could early phases of significant global contraction inhibit extrusive volcanism? Like the Moon, Mercury appears to have been relatively quiescent in the last half of Solar System history. Ice-like radar reflective materials were detected in polar crater interiors, suggesting that even on a body adjacent to the Sun, polar volatiles may accumulate in protected areas. Together with the Moon, these deposits may contain a record of ancient comets. Upcoming missions will explore the half of Mercury unobserved by spacecraft, and address these questions.

Mars: Mars, approximately one-half the diameter of the Earth, shows no signs of plate tectonics in the observable record, and it too can be classified as a one-plate planet with a significant record of early impact history and volcanism preserved in its crust. But Mars differs from the Moon and Mercury in eight fundamental ways: 1) a huge crustal dichotomy is observed, with a northern lowland characterized by thin crust, and a southern upland characterized by more heavily cratered uplands and thicker crust. This dichotomy, marked by a scarp-like boundary along much of its length, formed very early in its history, perhaps due to impact processes, or to early internal processes. 2) Two regions, Tharsis and Elysium, represent huge topographic rises characterized by concentrations of volcanic and tectonic activity, including volcanic edifices.
hundreds of km across and over ten km high; at Tharsis, a massive fault-bounded canyon radiates from the summit of the rise with a length comparable to the width of the continental U.S. Unlike the Moon and Mercury, these rises suggest there may have been a significant upwelling component on Mars to the vertical tectonics of one-plate planets. 3) While the Moon and Mercury appear to have been volcanically quiescent in approximately the last half of Solar System history, evidence has been found on Mars for volcanic activity extending up to the geologically recent past. 4) In contrast to the Moon and Mercury, Mars is a "water" planet. Evidence suggests that the climate on early Mars may have been "warm and wet". Noachian-aged stream-like valley networks and lakes indicate extensive surface runoff and possible pluvial activity. Ground-water appears to have been sequestered in the subsurface below an evolving and gradually thickening global cryosphere. In the Hesperian, huge outpourings of lava were emplaced, and these were followed by cracking of the cryosphere and release of massive quantities of seques- tered groundwater, to form outflow channels and deliver sufficient quantities of water to the northern lowlands to form large seas and possibly oceans, leaving a residue of sediment. Such flooding events appear to have continued at a lower rate even to the geologically recent past. Throughout the Amazonian, the climate appears to have been cold and dry, much like it is today. 5) Potentially extreme oscillations in spin-axis obliquity appear to charac- terize Mars. The presence of Earth's Moon stabilizes Earth and prevents it from undergoing extensive obliquity excursions. The lack of a large moon at Mars, together with other spin-axis and orbital perturbations, appear to have prevented Mars to undergo obliquity variations of in excess of 75 degrees, and to suggest that it may have spent much of its history at obliquities much different than that of the present. Testimony to the effect of potential obliquity variations is evidence for recent ice ages, mid-latitude plateau icefields and valley glaciation, and huge tropical mountain glacial deposits. The martian climate record holds promise in understanding climate response to ex- treme variations. 6) Magnetic anomalies many times stronger than those on Earth have been detected in the crust of Mars; their origin, and the cause of their magni- tude, are unknown, but some have hypothesized that they represent very ancient plate tectonic processes. 7) True polar wander may have occurred on Mars; the early forma- tion of Tharsis at higher latitudes may have caused its reorientation to the equatorial regions, where it resides to- day. 8) Of the terrestrial planets other than the Earth, Mars is the most likely habitat for the formation and evolution of life; an early "warm and wet" period could have provided conditions necessary for the origin of life and it could be sustained in the subsurface.

**Venus:** And what of Venus, the most Earth-like of the terrestrial planets in terms of size, density and position in the Solar System? Does it have plate tectonics, is it a one-plate planet, or is it something in between? Exploration has revealed a planetary surface that, like the Earth, has no remaining morphological record of the first two-thirds of Solar System history. There is no ancient heavily-cratered terrain. The surface is predominantly volcanic in origin, with small amounts of intensely tectonically resurfaced terrain called tessera. Could these be the equivalent of the Earth's young seafloor and ancient continents? The impact cratering record is sparse, indicating an average age of the surface of less than a billion years; furthermore, the crater population cannot be readily distinguished from a com- pletely spatially random population and the vast majority of the craters have not been modified. These observations have led to the interpretation that Venus must have undergone a global-scale resurfacing in its recent history in or- der to destroy all previous craters. This resurfacing must have been rapid, as most subsequent craters have not been modified by the volcanism and tectonism that form the resurfacing. Stratigraphic relationships place the tessera deformation early, followed by volcanism. What could have caused such a configuration? Among the hypotheses are: 1) a transition from a mobile lid to a stagnant lid lithospheric regime, 2) episodic plate tectonics, and 3) cata- strophic overturn of a vertically accreted depleted-mantle layer and a peak-like volcanic aftermath. Could similar processes lie in Earth's past or future?

**Synthesis and Implications for Earth:** Evidence for geologically recent global resurfacing of Venus, an Earth-like planet, has provided new insight into the role of den- sity inversions in planetary evolution, their influence on the early history of the planets, and the mechanisms for the initiation of plate tectonics on Earth. For example, it is now thought that following planet-scale differentiation in the earliest history of the Moon, some of the resulting layers were gravitationally unstable, resulting in density inversions that delivered dense, radioactive-element-containing materials to form the core, and to subsequently generate the period of mare basalt emplacement. Similar mechanisms may have been operative in the early history of Mars. Could such mechanisms have been responsible for lithospheric foundering and the initiation of plate tec- tonics on Earth?

It is now appreciated that impact cratering is a funda- mental planetary geological process. It influenced Earth history: 1) when the proto-Earth was hit by a Mars-sized projectile, stripping off a significant amount of crust and upper mantle to form the Moon; 2) when huge thousand-km impact basins formed, causing excavation of crust and upper mantle, its redistribution and ejecta emplace- ment, uplift of geotherms, pressure-release melting, emplace- ment of vast sheets of impact melt, and atmospheric strip- ping, modification and renewal; and 3) throughout its his- tory when impacts caused regional to global-scale effects that resulted in serious, sometimes terminal, influence on the existing biosphere.

All of these insights from fifty years of space explora- tion provide a new perspective on our own Home Planet Earth. No longer do we view the Earth in isolation. In- stead, Earth is now a member of a family of terrestrial planets that have shared similar events and phases in their histories. We look to the geological record of one-plate planets to understand the role of impact cratering with time, and how the processes works. We look to Venus to understand how tectonism and volcanism might appear during the Earth's Archean period, billions of years ago. We observe the thermal evolution of different terrestrial planets and wonder what the distant future holds for Earth. Will plate tectonics cease on our planet, and if so, what will it look like then? Will the Earth's lithosphere undergo catastrophic overturn in the future, and if so, what will be the aftermath and the effect on life? Many of the answers to such questions lie in the results of space missions to be undertaken in the next half-century of Solar System explo- ration, as we return humans to the Moon and on to Mars, we explore the outer solar system in detail, and we extend our perspective to solar systems around other stars.
EVIDENCE FOR NON-POLAR ICE DEPOSITS IN THE PAST HISTORY OF MARS. James W. Head¹ and David R. Marchant², ¹Dept. of Geological Sciences, Brown Univ., Providence, RI 02912 USA (james_head@brown.edu), ²Dept. of Earth Sciences, Boston Univ., Boston MA 02215 USA (marchant@bu.edu)

Introduction: The polar caps provide a record of the recent climate history of Mars [1]. Studies of the spin-axis/orbital parameter history of Mars provide a robust solution for the most recent ~20 Ma of martian history, but cannot be mapped further back into the past due to the chaotic nature of the solutions [2]. Thus, deconvolving the complex climate history of Mars requires analysis of the basic geological information, and interpretation of the depositional record of glaciation and glacial conditions at non-polar latitudes. These interpretations are assisted by an understanding of glacial and periglacial conditions in areas that are polar analogs to Mars (such as the Antarctic Dry Valleys) [3], and an understanding of the behavior of polar ice under different insolation conditions, using global climate models (GCMs) [4-5]. Finally, the availability of very high resolution images and topography (e.g., MOLA, MOC, CTX, HRSC, HiRISE) provide the ability to characterize and interpret these deposits. Here we report on recent analyses to assess the presence, age, and significance of non-polar ice deposits as evidence of the history of climate on Mars.

The Current Environment: Polar regions represent cold traps for planetary volatiles and analysis of these areas permits an assessment of the amounts and types of volatiles, their stability and mobility, and the long-term geological record of climate change. Present polar deposits on Mars consist of a thin residual ice unit (Apl) overlying a thick sequence of layered deposits (Apl), and are of Late Amazonian age [6]. The individual layers in the current deposits are thought to be related to variations in spin-axis/orbital parameters [2]. These variations cause changes in insolation and climate, and corresponding variations in dust and volatile stability, mobility, transport and deposition [e.g., 7-8]. Recent analysis of the history of orbital parameters has shown that the current martian climate is likely to be anomalous, and that Mars may have spent much of its history at considerably higher obliquity than its present value [2]. Thus, deconvolution of the most recent ~20 Ma of martian history, deglaciation and deposition were driven by spin-axis/orbital parameter-induced climate change [e.g., 9]. MOLA-derived roughness shows preferential smoothing at sub-kilometer scales above ~30° latitude in both hemispheres, attributed to a young, superposed surface mantle deposit [11]. MOC data analysis [12] revealed the presence of many features that also showed a latitude dependence. Mustard et al. [13] showed the presence of a distinct pitted mantle texture between 30°-60° latitude in both hemispheres, interpreted to be the dissected remnant of a former ice-rich dust deposit. Poleward of 60° in both hemispheres, the terrain was characterized by bumpy polygon-like features interpreted to be different types of contraction-crack polygons, thought to mark the presence of shallow ice-rich deposits undergoing thermal cycling [e.g., 3,14]. Also documented within the deposit was the local presence of multiple layers [e.g.,15-17]. Features interpreted to be very recent water-carved gullies were observed to be latitude-dependent in their occurrence, concentrated at 30°-50° [e.g., 18-22]. Viscous-flow features, interpreted to be the result of the accumulation, mobilization and flow of ice-rich material [16], in local microenvironments [e.g.,23], occur in the same latitude band as the gullies. The global distribution of interpreted water abundance from Odyssey-GRS/NS data [24,25] shows a remarkable correlation with the latitude-dependent deposits and features interpreted from MOLA and MOC data, confirming earlier predictions about the stability of near-surface ice in martian near-surface deposits [e.g., 10].

Latitude is the single variable with which all of these diverse observations correlate, and climate is the only process known to be latitude-dependent. The very strong correlation between the nature of the terrain smoothness, the continuity of the mantle, the high interpreted water content, and the theoretical stability of ice in the near-surface soil, all compellingly point to climate-driven water ice and dust mobility, and emplacement during recent periods of higher obliquity [2]. Degradation and dissection of the deposit in mid-latitudes further point to recent climate change [e.g.,13], perhaps reflecting return of mid-latitude ice to polar regions during the recent phase of lower obliquity [e.g., 7,9].

(2) Northern High Latitude Cold-Based Glacial Crater Fill: Some northern high-latitude craters have been observed to contain concentric ridges arrayed in lobate patterns that start at the crater rim, descend down the walls and across the crater floor, and separate around central peaks. These have been interpreted to be drop moraines, deposited during the advance and retreat of a lobate cold-based glacier, originating on the crater rim [26].

(3) Mid-High Latitude Concentric Crater Fill: Concentric crater fill (CCF) was initially observed in Viking data [27]; new data show details of morphology and structure that support the role of CCF in the deposition of ice [28]. Recent image data suggest that CCF craters may have been ice-filled and that CCF formed as part of regional glaciation [29].

(4) Mid-Latitude Lineated Valley Fill (LVF) and Plateau Glaciation: Earlier studies emphasized the role of vapor-diffusion-assisted emplacement of ice in slope-related talus piles, causing talus lubrication and plastic flow of the debris [27]. New data have confirmed some earlier interpretations [30] that significant ice was involved and that debris-covered glacial flow formed regional valley glacial landforms [31,32].

(5) Mid-Latitude Lobate Debris Aprons (LDA): Earlier thought to represent ice-assisted creep [27], the intimate association of LDA with LVF [33], and LDA internal structure and morphology now point to a debris-covered glacier mode of origin for many LDAs [34].

(6) Evidence for Mid-Latitude Ice Highstands: New data show evidence for highstands of ice (e.g., perched lobes in...
high-standing box canyons, trimlines, moraines) suggesting that almost a kilometer of ice has been lost from LVF [35].

(7) Low Mid-Latitude Phantom Lobate Debris Aprons: New high-resolution data show evidence for the former presence of ice-rich deposits surrounding massifs at latitudes even lower than the LDA, interpreted as representing the presence of former ice lobes [36] at even lower latitudes (~30°) than the LDAs.

(8) Tropical Mountain Glaciers: New data suggest that the fan-shaped deposits on the NW flanks of the Tharsis Montes and Olympus Mons represent huge tropical mountain glaciers [37–42] formed during periods of high obliquity [43] during the Late Amazonian.

(9) Near Equatorial Outflow Channel Rim Deposits: The graben at the origin of the major outflow channel Mangala Valles at latitude 18°S contains glacial-like features on its rim, suggesting that the climate earlier in the Amazonian was cold enough in the near-equatorial regions to cause glaciation, rather than runoff [44]. The lack of evidence of melting of these glacial features suggests that the outflow of water did not radically change the climate.

(10) South Circumpolar Ice Cap: The Hesperian Dorsa Argentea Formation: The set of Hesperian-aged south circumpolar deposits represented by the Dorsa Argentea Formation (DAF) [6,45–46] has been interpreted to be a volatile-rich polar deposit representing more than twice the area of the present Amazonian-aged layered terrain and residual polar ice, which it currently underlies. This huge polar ice-related deposit makes up about 2% of the surface of Mars and has undergone significant evolution since its emplacement. The deposit characteristics (e.g., smooth, pitted and etched deposits, pedestal craters, sinuous ridges interpreted by some as eskers, fluvial channels around the margins, and marginal plains thought by some to be remnants of ponds and lakes, etc.) have been interpreted to indicate that the DAF contained significant quantities of water ice, and that it represented an ancient circumpolar ice sheet [46]. These data also suggest that the circumpolar ice sheet subsequently underwent meltback and liquid water sub-ice-sheet drainage, ponding in adjacent valleys, and ultimately draining, because of their latitude and their altitude.

Summary: Together, these data provide insight into the climate history of Mars; they suggest that the climate of Mars has been similar to that of today for much of the Amazonian, with climate variations being driven largely by changes in spin-axis/ orbital parameters [2] and that obliquity was above 45° for part of the Late Amazonian. The Hesperian-aged DAF suggests that conditions were different in this important transitional period, with the possibility of a thicker atmosphere producing the huge south-circumpolar DAF. These observations provide an important context for the assessment of the Noachian climate history of Mars.

References:
Volcano-Ice Interactions at Arsia Mons, Mars.

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Introduction: Late Amazonian fan-shaped deposits on the northwest flanks of the Tharsis Montes volcanoes (Fig. 1) are interpreted to have formed as a result of cold-based tropical mountain glaciation [1-3]. Depositions and landforms interpreted to have formed during subglacial eruptions have been documented beneath and within these deposits. At Pavonis Mons sufficient meltwater may have been generated during the intrusion of steep-sided sill-like subglacial flows [2] to produce subglacial lakes [4]. The presence of subglacial lakes could have led to local wet-based conditions, local tongue-like wet based glacial surges, and to the release of meltwater from the margin of the glacier in fluvial drainage channels or jokulhlaups.

Subglacial intrusion, eruption, melting, and meltwater drainage: At Arsia Mons [1] (Fig. 2), the tropical mountain glacier deposits contain features and structures interpreted to represent subglacial and englacial eruptions [5]. These include low ridges interpreted to be dikes, lobate deposits interpreted to be steep-sided flows, linear mounds and low ridges interpreted to be moberg-like ridges and cones, and elongated depressions and trough-like features interpreted to be the result of subglacial and englacial phreatomagmatic eruptions [5]. Here we describe a series of features that together are interpreted to represent a linear subglacial eruption that caused the production of sufficient meltwater to form eskers, a local wet-based glacial surge that formed anomalously lobate moraines, and a distal series of channels emerging from the edge of the glacial deposit and flowing downslope into the surrounding terrain that is interpreted to represent subglacial drainage following the eruption.

A series of NW-trending preglacial lava flows extend down the flanks of Arsia and the fan-shaped glacial deposits are superposed. Facies in the NW part of the Arsia deposit include concentric ridged deposits interpreted to be drop moraines and hummocky deposits interpreted to be sublimation tills (Fig. 2). The western structure (Fig. 2, red arrows; Fig. 3) contains a lobe-shaped plateau and crater near the edge of the hummocky facies, and forms subaerial cones along its northern extension, between the inner and outer ridged facets, and again outside the glacial deposits to the north. The lobe-shaped plateau (∼9 km long and 6 km wide) extends downslope and the adjacent crater is ∼4 km wide and ∼100 m deep (Fig. 5). The plateau is ~130-150 m high and extends from the base of a ridge that is ~350 m high. We interpret the ridge, located along the strike of the linear trend, to be a sub-glacial moberg-like ridge, and the elongate plateau to be a subglacial, sill-like lava flow extending from the vent. Superposed on the lobe plateau, and extending downslope and out onto the subjacent lava flows, is a sinuous ridge that is generally continuous for ~14 km (Fig. 5); we interpret this to be an esker draining subglacial eruption-induced meltwater. Adjacent to and downslope from the ridge/lobate plateau the configuration of the drop moraines bows outward for a distance of ~5-10 km (Figs. 3, 5). The most prominent drop moraine occurs at the distal (downslope) edge of this inner set of drop moraines and several fluvial channels emerge from its base and extend at least 5-7 km into the surrounding terrain (Fig. 4).

We interpret this configuration to be the result of subglacial volcanism and volcanically induced meltwater generation and drainage. Sufficient meltwater appears to have been generated to cause a local transition from cold-based to wet-based conditions, producing local wet-based surging of a 30-40 km wide portion of the otherwise cold-based glacier. In addition, drainage of the meltwater from beneath the glacier out into the surrounding terrain formed local fluvial channels at the glacier margin (Fig. 4).

Subglacial explosive eruptions: The Aganippe Fossae system consists of two major troughs that extend in a NNW direction across the northwestern flank of Arsia Mons and are very closely associated with TMG deposits (Fig. 2; green arrows). The fossae typically range from ~2-5 km in width and are discontinuous in nature, having been modified by portions of the fan-shaped deposit. At the southern margin of the fan-shaped deposit, the wide southwestern trough abruptly changes its morphology into a narrow graben. Although the fossae have clearly been modified by subsequent events, the widest and most distinctive structures generally occur in the interior of the fan-shaped deposits (Fig. 2); at the northern end, large elongated pit craters occur (Fig. 6).

In some places the troughs appear to be buried by the knobby facies, thought to represent the collapse and vertical downwasting of the TMG, while in others the margins of the troughs appear sharp, but are likely to have been modified by post-formation mass-wasting and glacial modification of the trough walls. Stratigraphic relationships suggest that the troughs formed during the emplacement of the Arsia TMG. The occurrence in the western distal part of the Arsia TMG of dike-related subglacial volcanism strengthens the likelihood that subglacial eruptions might also have occurred in the thicker proximal portion of the ice sheet. As overlying ice becomes thicker (estimates range up to 2-3 km), dikes will intrude further into the ice sheet causing melting of marginal ice and the production of meltwater both in the crustal cryosphere and the glacier itself. We have modeled the intrusion of dikes into this configuration and find that 1) sufficient meltwater is produced to create hydromagmatic interaction and explosive phreatomagmatic eruptions, and 2) that further explosiveness occurs when the glacial ice is breached and exsolved magmatic gas undergoes catastrophic decompression to ambient conditions. For example, a typical 40 m wide dike emplaced into this configuration can create a phreatomagmatic eruption that will form an elongate depression ~2.7 km wide and ~2000 m deep. These phreatomagmatic eruptions would have produced ejecta composed of a combination of chilled fragmental magma, fragmented country rock, and aqueously altered tephra. Estimates of the initial volumes of the troughs suggest that as much as 600 km$^3$ of country rock ejecta might have been produced, forming a layer averaging ~4 meters thickness over the entire TMG deposit, potentially influencing glacial dynamics and stability. Subsequent glacial and mass-wasting activity has altered the fossae to their present configuration.

Figure 1. Fan-shaped deposits (yellow) interpreted to be tropical mountain glaciers [1] on the NW flanks of the Tharsis volcanoes.

Figure 2. Arsia tropical mountain glacier deposit [1] showing major facies; red arrows show location of deposits interpreted to be from a subglacial eruption (Fig. 3-5); green arrows show location of graben and pits interpreted to be from a phreatomagmatic eruption (Fig. 6).

Figure 3. Northwest part of the Arsia TMG, showing plateau and crater (interpreted to be subglacial eruption deposits), and convex outward drop moraines (interpreted to have resulted from local wet-based conditions). Contour interval is 20 m.

Figure 4. Channels emerging from the margin of the lobe shown in Fig. 3, interpreted to be the result of drainage of meltwater from the subglacial eruption (Fig. 3, 5).

Figure 5. Enlargement of Fig. 3 showing the location of the esker-like ridge (lower right, just below the crater) and the channels shown in Fig. 4 (upper left at the edge of the moraine.

Figure 6. Elongated pit crater located at the northern end of the Aganippe Fossae graben just at the edge of the main sublimation till (upper green arrow in Fig. 2).
NEW EVIDENCE FOR KILOMETER-THICK AMAZONIAN ICE DEPOSITS IN AN IMPACT CRATER IN PHLEGRA MONTES, MARS.

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Introduction: Lineated valley fill (LVF) in fretted valleys at the dichotomy boundary on Mars has been interpreted by some as glacial in origin [1–4]. Unknown are 1) the original thickness of the glacier ice, and 2) the amount of ice-surface lowering, through sublimation, retreat and ice loss, to its presently observed level. We recently addressed these questions through an analysis of an integrated LVF glacial land system in the Protonilus Mensae-Coloe Fossae area of the dichotomy boundary [3]. There, the elevation difference between the upper limit (determined from a lobe of LVF entering a box canyon at high elevations and being stranded there) and the current surface of the LVF at the study site, is \~920 meters. We interpreted this difference to reflect the minimum amount of ice-surface lowering of the valley glacier system during retreat. Consistent with a general lowering of the surface are multiple moraines and/or trimlines, and changes in LVF flow patterns indicating local reversals in flow patterns as the ice retreated and decreased in thickness. These data suggest that the major Late Amazonian LVF glaciation involved significantly larger amounts of glacier ice than previously thought. Here we report on additional evidence for thick mid-latitude ice deposits in the Phlegra Montes area, at the same latitude band in the northern hemisphere of Mars.

Phlegra Montes: The Phlegra Montes extend northward from the Elysium rise into the northern lowlands for many hundreds of kilometers (Fig. 1). These mountains, together with numerous areas along the dichotomy boundary and other regions in the 30°–50° N latitude band, display geomorphic features (lineated valley fill, LVF; and lobate debris aprons, LDA) interpreted to be indicative of 1) vapor-diffusion-generated, pore-ice-assisted creep [5,6] or cold-based glaciation [1–4].

Located on the western flanks of Phlegra Montes is a \~32 km diameter impact crater with a smaller, \~8 km diameter crater on its northeast rim (Fig. 1, arrow; Fig. 2). Although containing a sharp rim crest, the crater has undergone significant modification and its floor is filled with concentric ridged terrain arrayed around the interior, a texture typical of craters characterized by "concentric crater fill" [5]. MOLA altimetric profiles (Figs. 2, 4) reveal that the crater floor is very asymmetric, with the floor tilted poleward. The interior can be divided into five distinct segments from S-N, each of which corresponds to a different geomorphic texture (compare Fig. 2 and 4, bottom). From the southern rim of the crater, these include: 1) the steep upper part of the southern interior wall, about 3 km wide, that descends over 800 m from the sharp rim crest (~3060 m) to the base of the wall; 2) a shallow, convex-downward depression, about 4 km wide, that corresponds to a zone of radial ridges and depressions; 3) a convex-upward portion of the floor, about 7.5 km wide, characterized by concentric ridges and troughs, and pitted terrain in its downslope portion (the lowest portion of the floor, at \~4304 m, separates this segment from the next); 4) a slightly convex-upward portion of the floor about 8 km wide and sloping to the south that is characterized by concentrically textured and ridged material; 5) the northern wall of the crater, about 5 km wide, rising about 1320 m to the crater rim crest at -2880 m. These characteristics are very similar to the interiors of modified craters in the mid-high latitudes on Mars (concentric crater fill [5]), showing slope asymmetries [7] and interior modification related to ice-assisted downslope movement [8]. Uncertain, however, has been the relative roles of 1) ice-assisted creep of crater wall talus (diffusion, condensation, and freezing of atmospheric water vapor during periods of climate change) versus 2) the direct precipitation and accumulation of snow and ice to produce debris-covered glaciers. In scenario 1), the deposits are relatively local, and were dominated by pore ice and some accretionary ice. In scenario 2), significant accumulation results in glacial ice and flow, followed by climate change, retreat and recession to the presently observed feature, the final remnant of a debris-covered glacier.

The 8 km crater perched on the main crater rim (Fig. 2, 3) provides data to help distinguish between these two hypotheses. The interior is characterized by textures typical of LDA/LVF-type fill (Fig. 3), but instead of forming on the crater interior walls and flowing down across the breached low point in the rim, down the wall, and onto the large crater floor, it appears to have entered the small crater from the opposite direction, the direction of the crater interior. In addition, two smaller, but similar lobes, extend from the northern rim crest of the small crater (at \~11 and 1 o'clock) for several km, pointing northward. Taken together, these observations suggest that the large lobe was emplaced from the crater interior, and that the two small lobes represent overflow from a once-filled small crater.

Interpretation: On the basis of these relationships, we interpret the lobe in the small crater to have been formed from debris-covered ice that accumulated in the main crater interior and then flowed downslope northward into the smaller crater on the rim, filling it sufficiently to cause two small lobes to breach its northern rim. If this interpretation is correct, then additional areas of ice overflow should occur, and indeed, such structures are seen along the relatively low eastern rim. Topography data provide information on the initial configuration required for this scenario (Fig. 4). MOLA profile 13024 suggests that a total thickness of \~1000 m is required to fill the large crater, flow into, and breach the northern rim of the small crater. This thickness is comparable to that interpreted to have been present (and subsequently lost) in Protonilus Mensae-Coloe Fossae area \~920 m). What was the fate of this glacial ice? Detailed textures in the crater lobe suggest lowering and some backflow in toward the main crater. In the southern region of the crater floor at the edge of the main crater wall, a zone of parallel ridges and troughs, reminiscent of crevasses formed from glacial ice descending over underlying topography, is observed. At least 480 m of ice loss from sublimation and retreat are implied within the small crater, and an additional 530 m from the main crater interior. These data add to evidence [3] that the formation of Amazonian mid-latitude LDA/LVF deposits involved ice sheets with thicknesses measured in hundreds of meters to several kilometers.

Figure 1. MOLA topography over MOLA hillshade. Hecates Tholus is in the southwest of the frame, while Phlegra Montes is to the east.

Figure 2. MOLA topography over MRO CTX images (P01_001553_2232 and P01_001619_2232).

Figure 3. MRO CTX subframe of image P01_001553_2232.

Figure 4. MOLA track profiles from North to South across the larger host crater. Profile locations are shown in Figure 2.
Exploring the surface of Venus with VIRTIS on Venus Express. J. Helbert¹, N. Müller¹, P. Kostama², G. Hashimoto³, L. Marinangeli⁴, G. Piccioni⁵ and P. Drossart⁶ and the VIRTIS on Venus Express team, ¹DLR, Rutherfordstrasse 2, 12489 Berlin, Germany, joern.helbert@dlr.de, ²University of Oulu, Finland, ³Graduate School of Science and Technology, Kobe University, Nada-ku, Kobe 657-8501, Japan, ⁴IRSPS, Universita’ d’Annunzio, Viale Pindaro, 42, Pescara, 65127, Italy, ⁵INAF - IASF Roma, Via del Fosso del Cavaliere 100, 00133 Roma, Italy, ⁶LESIA - Observatoire de Paris, 61 avenue de l'observatoire, Paris, 75014 France

Introduction: Venus Express is since April 11, 2006 in orbit around Venus. VIRTIS (Visible and Infrared Thermal Imaging Spectrometer) has started only hours after orbit insertion to obtain data and has since then collected an immense and unique dataset. Of special interest for us is the wavelength range from 1-1.5 microns which includes the “atmospheric windows”. We will report here about ongoing activity to use this data to characterize the Venusian surface.

“Seeing the surface”: The M-IR channel of VIRTIS allows observing the surface of Venus in three small atmospheric windows at 1.02, 1.10 and 1.18μm. While the atmospheric windows show little CO₂ absorption on its way through the atmosphere, the thermal radiation is modified by scattering and absorption by clouds. Variations in the optical thickness of the clouds modulate the spatial distribution of upwelling radiation. Multiple reflections between surface and clouds generally wash out image contrast from surface emissivity. We have developed a data processing pipeline based on the approach by [1] which allows processing the huge amount of data returned by VIRTIS in a timely manner. The data processing corrects for various atmospheric effects including limb darkening, scattered sunlight, etc and reduces the cloud induced contrast variations. The algorithm has been extensively tested and is continuously improved based on the available data.

The results of this approach are ambiguous since neither surface emissivity nor surface temperature are well known. This means only one of these quantities can be retrieved at a time based on reasonable assumptions for the other quantity. Surface temperature is mostly a function of altitude, no diurnal or seasonal variations are expected. Furthermore aerosols or a gradient of absorbing gaseous constituents near the surface might affect the interpretation. However neglecting any effects of the near surface atmosphere and assuming parameters of the radiative transfer model arbitrarily within reasonable ranges it is possible to estimate either surface emissivity or surface temperature.

Mapping the surface: Based on averaging several hundred multi-spectral images obtained during the mission we have been able to map the surface temperatures for almost the whole southern hemisphere of Venus. Blurring in the clouds [1] limits the spatial resolution to more than 100km globally, but locally spatial resolution as high as 80km are achievable.
This flow dimensions are well in the lower range of lava flows observed by synthetic aperture radar (SAR) images obtained by Pioneer Venus and Magellan spacecrafts [4]. Small lava lakes covering an area of less than 1km² are detectable if the surface temperature exceeds 1200K. Assuming that the eruption temperature of the magma is approximately the liquidus temperature for basalt temperature up to 1500K can be expected. Estimates on the lava cooling rate by [5] indicate that an eruption would be undetectable after one Earth day. This implies that the chance of detecting volcanism increases if the time between surface observations of the same location is small. Therefore analyzing the differences between successive temperature maps is the most sensitive tool. So far we have only few example where we monitored the same area on consecutive days. Based on global and temporal coverage our chances to detect active volcanism if it still exists have been less than 10%. This number will increase with mission duration and possible adjustments to the observation plans. In the case of the detection of a thermal anomaly further confirmation can be derived from an analysis of the near surface atmosphere composition. Volcanic activity should produce a localized increase of volcanic gas emission (CO₂, CO, SO₂, HF, CH₄).

Emissivity variations: The most exciting and at the same time the most complex task is to retrieve compositional information for the surface of Venus from the hyperspectral data obtained by VIRTIS.

The currently used simplified atmospheric model included in the data processing does not allow a direct retrieval of the absolute emissivity of the surface. Given the limited knowledge and the variability of atmospheric properties it is unclear whether even a more sophisticated atmospheric model would allow to derive absolute emissivities with a high level of confidence. The approach used currently is to derive relative emissivity variations. Assuming that the atmosphere below the clouds is horizontally homogeneous, scattering and absorption are determined by the thickness and pressure of the atmosphere given by topography. Therefore this flux is a function of topography and surface emissivity. Assuming proportionality of flux to emissivity (an approximation which neglects multiple reflections (see [6])) and absence of a global trend of surface emissivity with elevation, local variation of emissivity is calculated as quotient of the local upwelling flux to a global polynomial fit of degree 2 to the flux with respect to topography.

This approach seems very stable and produces interesting results. Spatial variation of surface emissivity shows correlation with some geological features known from Magellan radar images. Specifically in the Lada region large lava streams, Cavillaca - and Juturna Fluctus, show increased emissivity with respect to neighboring regions of the same altitude. Other large lava streams in the region show a similar but less obvious relative emissivity. Quetzalpetlatl Corona has evolved in several phases [7,8]. The last phase produced the lava flows to the south of the corona – mainly Juturna Fluctus. These flows might therefore represent late stage volcanism on Venus. Tentative interpretations for the observed emissivity anomaly are either an indication of differences in the chemical weathering of the younger surface units compared to the surrounding older material or an evolution of the magma composition in the late stage of activity. Currently both hypotheses are support by the data, however the former would suggest a very recent emplacement of the lava flows.

**Laboratory support:** To support the data analysis of VIRTIS and especially the interpretation in the context of potential surface composition plans are under way to upgrade the Planetary Emissivity Laboratory (PEL) at DLR in Berlin. This upgrade would allow to measure directly the emissivity of analog materials at temperatures typical for the Venusian surface. Especially the measurement of basaltic rocks weathered in acidic environments would potentially allow to identify the cause for the observed emissivity anomalies.

**Summary:** While there have been observations using the atmospheric windows by Galileo/NIMS [2] and Cassini/VIMS [3] VIRTIS on VenusExpress allows for the first time to systematically map the surface radiance from orbit over a long period of time. The full potential of this unique data set only starts to emerge. This includes the first active volcano watch from orbit and the first indication for variability in the surface composition from emissivity variations. The data that VIRTIS can provide on the surface of Venus is highly complementary to the existing datasets from Magellan and the Russian lander probes and a combination of these datasets might significantly advance our understanding of the evolution of Venus.

**References:**  
SURFACE FEATURES IN SNEGUROCHKA PLANITIA (V1) AND THEIR IMPLICATIONS FOR MANTLE EVOLUTION ON VENUS.  D. M. Hurwitz and J. W. Head, Dept. of Geological Sciences, Brown University, Providence, RI 02912, debra_hurwitz@brown.edu

Introduction: The tectonic system on Venus is distinct from the plate tectonics system found on Earth. Venus lacks the low viscosity zone (LVZ) that facilitates Earth’s plate tectonics, leaving Venus with a single planet-encompassing plate that interacts directly with the upper mantle [1]. The direct coupling of the lithosphere and the upper mantle facilitates vertical tectonics (i.e. transient mantle plumes) rather than horizontal tectonics (i.e. subduction) and yields surface features unique to Venus. Previous work (e.g., [1-13]) has described possible formation mechanisms of surface features found on Venus within a vertical tectonic system. These descriptions can be used to define a possible mantle evolution for the Snegurochka Planitia (V1) region at the northern pole of Venus.

Description of Snegurochka Planitia (V1): Observations of Snegurochka Planitia are made using Magellan radar images, shown in Figure 1 in both a visual image and a sketch map. The region consists of several units characteristic of Venus, including deformational ridge belts (dorsae), regional plains (planitiae) with both compressional features (wrinkle ridges) and extensional features (chasmata), mountain belts (tesserae), coronae, volcanoes with associated lava flows, and fields of small shield volcanoes.

The dominant two dorsae in the region appear to intersect at the pole, with Dennitsa Dorsa lying at 200E and Sel-anya Dorsa lying at 90E. A third belt, Semuni Dorsa, lies near latitude +80°. These ridge belts rise approximately 1km above the surrounding plains and are characterized by their near-parallel ridges and radar brightness. Surrounding these deformational belts lie the regional plains of Louhi Planitia and the region’s namesake, Snegurochka Planitia. These plains are deformed by both compressional and extensional features and lie at lower elevations in radar darkness.

The tesserae in the region lie mainly to the south, with Itzpapalotl Tessera lying between longitudes 330E and 30E. This tessera belt extends into the neighboring region and merges with Fortuna Tessera and Lakshi Planum, highlands that include Maxwell Montes, the highest topography on Venus, rising 8 km above the surrounding plains [14]. A second tessera belt within the V1 region lies at the northern end of Sel-anya Dorsa. Tessera is characterized by radar bright, heavily deformed regions elevated above the surrounding plains.

These three features, dorsae, planitiae, and tesserae, represent the oldest features in the region. They have been the most heavily deformed, with extensional features crossing unit boundaries and with other features superimposed on them. These relatively younger features include coronae, large volcanoes, and small shield volcanoes.

Two coronae, Anahit and Pomona, formed on the western flanks of Itzpapalotl Tessera between 270E and 315E. A third corona, Meslenitsa Corona, formed at the southern end of Dennitsa Dorsa near +75°, 200E. These features are characterized by a raised circular plateau surrounded by a raised rim and bounded on at least part of the outer edge by a trench. Radiating extensional lineaments cut across the surrounding plains, suggesting that parts of the coronae formed after the plains. Renpet Mons, a large volcano with a diameter of 300km, lies between these two sets of coronae at +76° 235E [15]. This volcano fed extensive lava flows that overlie the surrounding Snegurochka Planitia.

While Renpet Mons represents the only large volcano in the region, many smaller volcanoes feed smaller-scale flows across the region. A large cluster of small shield volcanoes lies nearly parallel to Dennitsa Dorsa in Louhi Planitia near 180E. This shield field lies at a low elevation, and associated flows superpose the surrounding plains. A second cluster of small shield volcanoes lies between Meslenitsa Corona and Renpet Mons, a third between Anahit Corona and Renpet Mons, and a fourth, smaller cluster lies within Pomona Corona. The shields appear to superpose the plains on which they lie except for the shields within Pomona Corona which appear to have been embayed in the lower-lying regions of the corona’s center.

Discussion: The observations presented above indicate a relative stratigraphy for Snegurochka Planitia, from oldest to youngest, of tessera, dorsa, planitia, corona, large-scale volcanism, and small-scale volcanism. Previous work [1-13] has investigated formation mechanisms for these features, and using these explanations, a brief tectonic history can be approximated for Snegurochka Planitia.

Tessera and other high-elevation features form in response to crustal thickening as a result of either mantle up-welling or mantle down-welling. Diffuse mantle up-welling may induce partial melting that results in crustal thickening [e.g., 2]. While mantle down-welling might induce overlying crustal flow which also results in crustal thickening [e.g., 3]. In either case, the elevated crust is thicker and/or less dense than the surrounding plains crust.

Dorsae are also elevated features and form from compressive stresses associated with convective mantle down-welling [4]. These deformation belts are consistently surrounded by lower-elevation planitiae. Planitiae may mark sites of mantle up-welling or hotspots where the crust has thinned in response to dynamic uplift [6]. This juxtaposition of dorsae, formed by mantle down-welling, and planitiae, formed by mantle up-welling, implies that convective down-welling must also induce partial melting that in turn
gives rise to a localized, shallow, positively buoyant magma source that yields the volcanic plains [3].

Coronae form from uplift of buoyant diapirs or plumes that originate from either the core-mantle boundary or from a compositional layer in the upper mantle followed by gravitational relaxation [1, 2, 5, 6]. The relation between this type of mantle plume and the types that form the principal highlands is uncertain, but the plumes that form the coronae may be smaller or originate from a different location in the mantle [1]. Therefore the proximity of corona to tessera (thought to form from mantle up-welling) and to dorsae (thought to form from mantle down-welling) can be explained by corona-forming plumes originating at different mantle sources than the older features.

The final features observed are the volcanoes, both large and small. Renpet Mons formed from a rising magma plume that reached shallow depths and fed surface flows. The observed small shield volcanoes formed from independent magma sources [13]. Shield fields in Snegurochka Planitia appear to be concentrated within the plains units in regions of thin crust that allow magma from wide-spread, shallow sources to reach the surface.

Conclusions: The observed features and their associated formation mechanisms suggest that Snegurochka Planitia (V1) was deformed mainly due to region-wide mantle down-welling, with mantle up-welling occurring to the south, especially between 200E and 360E. The up-welling to the south induced the formation of Fortuna and Itzpapalotl Tesserae, and mantle plumes formed the nearby Anahit and Pomona Coronae and Renpet Mons. Mantle down-welling formed the deformation belts in the center of the region. Between points of mantle up-welling and down-welling, where mantle transport is relatively flat and has a greater horizontal effect on the crust, regional plains formed and, later, small shield volcanoes formed. This is a preliminary explanation of what shaped the surface of Snegurochka Planitia, and current mapping efforts underway will address complications on smaller scales in more detail in order to better define the mantle evolution in this northern region of Venus.

Introduction and the problem statement:

Ground penetrating radar (GPR) is the promising technique to explore planetary interiors, successfully applied in the Mars research program [1,2]. One of the obstacles limiting the GPR performance is the ionosphere. The radar signal phase distortion in the dispersive plasma is studied relatively well, while a limited attention has yet been paid to the influence of the fluctuations of the plasma density.

In the present work, HF GPR sounding from the orbiting spacecraft through the ionosphere has been simulated numerically. Phase distortion of the ultra wide band (UWB) chirp signal, introduced by the regular layered ionosphere and the effect of scattering of the radio waves by the small scale ionospheric plasma irregularities are both incorporated in the numerical model. The effect of planetary surface roughness on the chirp pulse distortion is also studied in the present work.

In addition, aperture synthesis is a relatively complicated operation, which presents significant difficulties in computer simulations. For this reason, many papers dealing with spaceborne GPRs treat the aperture synthesis with more or less simplifying assumptions and restrictions. Thus, strict modeling of the aperture synthesis are still of value. In the present study, correct simulation of the aperture synthesis is attempted.

Numerical simulations:

Physical model of the ionospheric irregularities is selected on the basis of experimental data [3]. The role of effects of wave diffraction caused by the small-scale structures in the ionospheric plasma has been studied. Influence of the anisotropy of the correlation function of the plasma density is discussed.

Degradation of the signal caused by the side clutter coming from rough surface of the planet is also estimated within the Kirchoff approximation. Numerical simulations are performed for different types of rough surfaces. The most reliable model, providing satisfactory explanations for much of experimental results, is the normally distributed roughness height deviation with the exponential height correlation function.

In the figure, simulated output signals of the synthetic aperture radar (SAR) are shown for several different lengths of the synthetic aperture (labeled by numbers near the each curve). Phase fluctuations of the wave due to ionospheric disturbances are moderate, so that deviation of the phase does not exceed one whole period. One can see that application of the aperture synthesis allows to significantly improve the performance of the instrument.

Conclusions and remarks:

It has been shown that the density fluctuations of the ionospheric plasma introduce significant distortions in the chirp signal. Satisfactory chirp signal compression may be obtained at frequencies high enough in comparison with the critical frequency of the ionosphere.

On the other hand, correction algorithms for compensating of the phase distortions in the regular ionosphere are relatively stable with respect to the small scale irregularities, at least for not so large density fluctuations, typically occurring in the terrestrial atmosphere. Thus, the conclusion can be made is that the regular phase distortions and the scattering on the small scale structures contribute independently in the degradation of the radar chirp signal.

Aperture synthesis can to a certain extent eliminate the destructive effects caused by the ionospheric turbulence, but when strong plasma density fluctuations occur, the signal spread due to the scattering on the plasma irregularities completely prevents observation of the subsurface echoes. The function of the synthetic aperture radar in the inhomogeneous ionosphere is thoroughly studied here, and the optimal parameters of signal processing are pointed out. In addition, the stepped-frequency radar (SFR) output has been simulated numerically. The stability of SFR with respect to the surface roughness and the ionospheric disturbances is estimated, and the results of comparative study of both types of radars are presented.

Acknowledgements:

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References:

GLOBAL GEOLOGICAL MAP OF VENUS: PRELIMINARY RESULTS. M. A. Ivanov, Vernadsky Institute, RAS, Moscow, Russia (Mikhail_Ivanov@brown.edu).

Introduction: The Magellan SAR images provide sufficient data to compile a geological map of nearly the entire surface of Venus. Such a global and self-consistent map would serve as an important document to address many key questions in the geologic history of Venus. 1) What units/structures characterize the surface [1-3]? 2) What volcanic/tectonic regimes do they represent [4-7]? 3) Did these regimes occur globally or locally [8-11]? 4) What are the relative time relationships among the units [8]? 5) Are these relationships consistent regionally or globally [8-10]? 6) What is the absolute timing of formation of the units [12-14]? 7) What is the history of the long-wavelength topography and geoid? 7) What model(s) of heat loss and lithospheric evolution [15-21] does this history represent? The ongoing USGS program of Venus mapping has already resulted in a series of published maps at the scale 1:5M [e.g. 22-30]. These maps have a patch-like distribution, however, and are being compiled by authors with different mapping philosophy that are not always in agreement with each other in terms of mapped units and their relationships. Here the preliminary results of global geological mapping of Venus at the scale 1:10M is presented, representing the current status of a global mapping project. The map represents a contiguous area extending from 82.5°N to 22.9°S and comprises ~70% of the planet. The new map permits one to address some of the questions raised above.

Mapping procedure: For the initial mapping analyses, images of high resolution (C1-MIDR and F-MIDR) were used to define units [1,9,11]. The map was then compiled on C2-MIDR sheets, the resolution of MIDR) were used to define units [1,9,11]. The map was completed using published USGS maps [e.g., 22-30] and the catalogue of impact craters [31]. The results suggest that the mapping on the C2-base provided a high-quality map product.

Units and structures: A set of material units and tectonic structures describes the geological configurations throughout the map area (Fig. 1). The complete stratigraphic column consists of the following units (from older to younger): Tessera (t) displays multiple sets of tectonic structures. Densely lineated plains (pdl) are dissected by numerous subparallel narrow and short lineaments. Ridged plains (pr) commonly form elongated belts of ridges. Mountain belts (mt) resemble ridge belts and occur around Lakshmi Planum. Shield plains (psh) have numerous small volcanic edifices on the surface. Regional plains were divided into the lower (pr1) and the upper (pr2) units. The lower unit has uniform and relatively low radar albedo; the upper unit is brighter and often forms flow-like occurrences. Shield clusters (sc) are morphologically similar to psh but occur as small patches that postdate regional plains. Smooth plains (ps) have uniform and low radar albedo and occur near impact craters and at distinct volcanic centers. Lobate plains (pl) form fields of lava flows that are typically undeformed by tectonic structures and are associated with major volcanic centers.

Specific structural assemblages accompany the material units: Tessera-forming structures (ridges and grooves), ridge belts, groove belts (structural unit gb), wrinkle ridges, and rift zones (structural unit rt). The tessera-forming structures and ridge belts predate vast plains units such as psh and rp1. Groove belts postdate tessera and ridge belts. Shield plains and regional plains mostly embay groove belts. In places, groove belts appear to form contemporaneously with the vast plains units. Wrinkle ridges deform all material units predating smooth and lobate plains. Rift zones appear to be contemporaneous with sc, pl, and ps and cut older units.

Crater statistics: Two factors, the atmosphere screening [32-34] and the observational bias [35], appear to affect the statistics of the smaller (<16 km) craters. For the larger craters these factors are negligible and the larger craters were used to estimate the crater density on mapped units. The shape and size of occurrences of units may also affect the crater statistics on Venus where the total number of craters is small. To minimize influence of this factor the crater density on large and contiguous units that have quasi-equidimensional occurrences was estimated. These units are: t, psh, rp1, rp2, pl, and rt; together, they cover ~84% of the surface. The mean densities (craters per 10^6km^2) of craters on these units are as follow: tessera: 1.046 (±0.158, one σ); shield plains: 1.004 (±0.126); regional plains, lower unit: 0.982 (±0.103); regional plains, upper unit: 0.614 (±0.140); lobate plains: 0.547 (±0.144); rifts: 0.609 (±0.173). The mean density of craters in the map area (all units) is 0.938 (±0.054). If the mean crater density corresponds to the mean surface age, T [19], then the ages of the above units as fractions of T are: tessera: 1.11T, shield plains: 1.07T, regional plains, lower unit: 1.05T, regional plains, upper unit: 0.67T, lobate plains: 0.58T, rifts: 0.67T.

These results are consistent with the mapped stratigraphic relationships and indicate that there are two groups of units: The older units (t, psh, rp1) are densely clustered around 1.1T and the younger units (rp2, pl, rt) were formed around 0.6T. The exposed area of the older units is ~59% of the map area (the true area
must be larger) and the younger units cover ~25% of the surface. Depending upon the estimates of the absolute value of T (from 750 Ma [36] through 500 Ma [37] to 300 Ma [38]), it is possible to estimate the duration of specific periods in the observable geologic history of Venus. The older units appear to form during a relatively short time, from 45 m.y (T = 750 Ma) to 18 m.y (T = 300 Ma). The minimum integrated resurfacing rate (both volcanic and tectonic) at this time was from ~4.2 to ~10.4 km²/y. The younger units spanned a longer time and the integrated resurfacing rate during their formation was from ~1.2 to ~3.0 km²/y. The large time gap (from 285 to 114 m.y.) between the two groups of units and the significant drop in the resurfacing rates suggest that the older and the younger units correspond to two different geodynamic regimes that were probably related to different patterns of mantle convection and lithospheric properties.


Fig. 1. Geological map of Venus. Scale is ~1:10M; Lambert equal-area projection.
THE HISTORY OF TOPOGRAPHY ON VENUS, M.A. Ivanov1,2, J.W. Head2, and A.T. Basilevsky1,2; 1 - Vernadsky Institute, RAS, Moscow, Russia, 2 - Brown University, Providence, RI, USA

Introduction: Topography of a planet can date from different parts of planetary history. The current bimodal topography on Earth reflects the combination of a thick low-density continental crust, and a thin denser oceanic crust forming and evolving as a part of a laterally mobile and cooling lithospheric thermal boundary layer. Venus is the most Earth-like of the terrestrial planets but displays no current plate tectonics and aqueous erosion. It is presently a one-plate planet, is losing its heat largely by conduction, and yet has an Earth-like average surface age (<~1Ga), and is characterized by a unimodal hypsogram. When did the major components of the current global topography of Venus form? We investigated this question through a synthesis of geological mapping over 70% of Venus, an assessment of the stratigraphic relationships and ages of the main units, and a comparison of their distribution and age to the present topography of Venus. Here we report on the sequence and timing of the global topography of Venus.

Major geologic units: Photogeologic analysis of radar images of Venus has revealed the major surface features, their general origin, and their sequence of formation. Initially identified stratigraphic sequences [e.g. 1,2] have been followed by systematic quadrangle mapping and global stratigraphic assessments [e.g. 3-11]. Ivanov [12] recently completed a synthesis of geological mapping from the 82.5°N to 22.5°S, ~70% of the planet.

Five units appear to be the most important on Venus. Tessera terrain (t; ~9.1% of the mapped area) is exposed as large plateaus and small outliers standing as much as several km above the surrounding plains [13]. The surface of tessera is heavily tectonized. Surrounding plains always embay tessera. Shield plains (psh; ~17.6%) form units a few hundred kilometers across characterized by numerous small (5-15 km) volcanic shields. Stratigraphic analyses show that shield plains largely predate regional plains [e.g. 14-17]. Regional plains (rp; ~35.7%) typically have a radar-dark smooth surface, which is interpreted to be solidified lava flows, and cover large contiguous areas [e.g. 18]. The plains are deformed by a network of narrow wrinkle ridges, a result of deformation of the surface by moderate horizontal contraction [e.g. 19]. The large areas, sinuous channels, flow fronts, and very gentle slopes, all suggest high-rate, low-viscosity eruptions of basaltic lava. Lobate plains (pl; ~7.5%) are observed in two partly overlapping environments: adjacent to rift zones and covering the slopes of volcanic edifices, often at the nexus of rift zones. The flows are often bright, n10-n100 km long, and a few to n10 of km wide. They overlie regional plains and embay other preexisting units. Rift zones (rt; ~6.3%) are usually topographic troughs a few km deep; walls and floors of the troughs are heavily deformed by fractures. Rift zones are contemporaneous with lobate plains.

The stratigraphic relationships of these units have been assessed by many workers, and there is general agreement that tessera is old, regional plains embay the tessera, and rifts and lobate flows are younger than regional plains [20,21].

Topographic configuration of the units: The distribution of the topography of the major map units shows that the oldest, tessera, is the highest (Fig. 1). The younger shield plains are concentrated at intermediate hypsometric levels (Fig. 1) and the plains sloping away from the tessera. Regional plains are the lowest unit hypsometrically (Fig. 1), and they primarily occur in the current lowlands. Stratigraphic relationships show that the majority of shield plains underlie the regional plains and thus they are probably also present in the low areas below the regional plains. The surface slope of the shield plains away from the tessera, and evidence that they underlie the regional plains in the present topographic basins, strongly suggest that the lowlands topography formed before regional plains, perhaps in broad concert with the formation of the tessera highlands. Rift zones and associated lobate flows form predominantly at intermediate to high elevations (Fig. 1). Rift zones are interpreted to be due to relatively recent upwelling and associated rifting-related crustal and lithospheric thinning. Lobate lava flows, emerging from rift margins and volcanic rises, trend down-slope into the adjacent midlands and lowlands, covering regional plains and shield plains.

Ages of the units: Magellan data revealed only ~970 impact craters, leading to the interpretation that the mean age of the surface of Venus, T, was ~750 Ma (300 Ma to 1 Ga) [22]. The mean age is usually expressed as T and the age of surface units as fractions or multiples of T (Fig. 1). A synthesis of the mean ages assigned to the units under consideration is tessera: 1.11 (±0.17) T, shield plains 1.07 (±0.13) T, regional plains1.05 (±0.11) T, lobate plains: 0.58 (±0.15) T, rifts: 0.67 (±0.18) T. These age estimates (Fig. 2) show that Venus was not resurfaced at a uniform rate. The deformation that characterizes the tessera terrain was rapidly followed by globally distributed small centers of volcanism (shield plains) and then by flood-basalt-like regional plains. All of this activity resurfaced ~62.4% of the mapped area (Figs. 2) in ~15% of the total preserved surface history of Venus. In the 85% of
history remaining to the present, rift and volcanic rises, and associated volcanism were the dominant processes, resurfacing only ~13.8% of the planet.

The History of Topography on Venus: These data now permit us to assess the age of the topography of Venus: what portion remains from its earlier history and how much is attributable to more recent activity? Comparison of units areal, altimetric, and age distributions (Fig. 1) strongly suggests that the vast majority of the current topography dates from the first ~15% of the observable (since tessera formation) history of Venus. Tessera occupies topographic highlands, shield plains occur preferentially on midland surfaces sloping toward the lowlands, and regional plains predominately occupy lowlands and the lower parts of the midlands. The fact that volcanic regional plains occupy lowlands is evidence that these lowlands were present at the time of their emplacement.

In the ~85% of morphologically recorded history that postdates regional plains emplacement, the major contributors to the current topography were rift zones, and the volcanic rises (e.g., Beta, Atla, Themis). Lobate lava flows emanating from these places clearly flow downhill into pre-existing surrounding midlands and lowlands, resurfacing regional plains and older units. The fact that this combined activity resurfaced only ~14% of the mapped area strongly suggests that: 1) the level of interior activity of Venus as measured by integrated volcanic flux decreased significantly following the emplacement of regional plains, and 2) the style of resurfacing changed dramatically from intense crustal deformation (tessera) and global high-flux volcanism (shield and regional plains), to local centers of volcanism interconnected by rift zones with associated local volcanism. We concur with other workers [e.g. 23] that this distinctive change is consistent with the early phase (tessera to regional plains) representing a major period of enhanced heat loss, while the later phase represents a prolonged more stable period of lithospheric thickening. We interpret the origin of the long-wavelength topography of Venus to be a remnant of the geodynamical forces related to tessera formation and the global volcanic aftermath represented by shield and regional plains emplacement. The preservation to modern times of the ancient topography of Venus is believed to be due to rapid lithospheric thickening following the emplacement of regional plains.

References:

Fig. 1. Altimetric distribution of topography (dot shows mean) of geologic units and structures (Fig. 1) as a function of mean age.

Fig. 2. Resurfacing rate as a function of mean age, in fractions of area covered by regional plains (rp).
INVESTIGATION OF ORDINARY CHONDRITES BY THE THERMOLUMINESCENCE METHOD. Ivliev A.I., Kuyunko N.S. Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow 119991 Russia; e-mail: cosmo@geokhi.ru. Tel.: (495)137-86-14

**Introduction.** The thermoluminescence (TL) method is one of the most commonly used methods of investigation of the stone meteorites [1]. There are two types of TL: natural TL$_{\text{NAT}}$ was accumulated by meteorite in the cosmic space, and induced TL$_{\text{IND}}$ was induced from external source of a radioactive radiation in the laboratory. The measurements of the TL$_{\text{NAT}}$ are mainly used for an estimation of the meteorite orbits [2], for analysis of a shocked metamorphism [3], for an estimation of the terrestrial ages of meteorites [4] and for determination of pair meteorite-finds [5]. The TL$_{\text{IND}}$ demonstrate the changes of a crystalline structure of the feldspar produced by thermal or shocked metamorphism. The measurements of the TL$_{\text{IND}}$ are used for analysis of a metamorphism of unequilibrated ordinary and carbonaceous chondrites [1, 6, 7] and also for a research of a shock - thermal history of meteorites [3, 8].

In the present work, the investigations of TL for recent measurements of the equilibrated ordinary chondrites: Barwell (L5), Chantonnay (L6), Dolgovoli (L6), Kilabo (LL6), Kunya-Urgench (H5), and Tugalin Bullen (H6) were made.

**Experimental method.** Investigated samples of meteorites weighing from 0.7 up to 1.0 g were crushed in the jasper mortar. The magnetic fraction of the powdered sample was moving off by a hand-held magnet. The nonmagnetic fraction of each meteorite was used to prepare of three samples each of which was equal 2 mg. After measurements of TL$_{\text{NAT}}$, samples were irradiated by X-rays or by γ-rays and then TL$_{\text{IND}}$ was registered.

**Shock loaded metamorphism.** The determination of a shock load pressures of the ordinary chondrites were determined by results of measurements TL induced by X-rays. The most sensitive TL parameter for determination of shock load degree of the ordinary equilibrium chondrites is area under of a glow curve in a temperature range 40-350 °C (Sp) [3, 11]. Measurements of the TL parameters in the samples of the 14 meteorites with known shock load pressures have shown the increase of values Sp at the increase of shock pressure up to 10 GPa (stages S1-S2), and subsequent their sharp decrease up to two orders of magnitude at further increase of shock pressure from ~10 up to 90 GPa (stages S3-S6). Using the results of our measurements and values of pressures of different shock classes of meteorites [12, 13], we have received the approximate formulas for an estimation of a shock load pressure of chondrites at collisions in space. For the shock classes S1-S2, it was obtained: P = 1.93 x ln (Sp) - 5.57, and for S3-S6: P = -12.28 x ln (Sp) + 91.74, where P is a shock load pressure in GPa, and Sp is area under a glow curve in a temperature range 40-350 °C.

<table>
<thead>
<tr>
<th>Meteorites</th>
<th>Type</th>
<th>Shock class</th>
<th>S$_{\text{p}}$</th>
<th>Shock class</th>
<th>P, GPa</th>
</tr>
</thead>
<tbody>
<tr>
<td>Barwell</td>
<td>L5</td>
<td>S3 [14]</td>
<td>590±35</td>
<td>S3</td>
<td>13.4±0.8</td>
</tr>
<tr>
<td>Chantonnay</td>
<td>L6</td>
<td>f [13]</td>
<td>7.0±1.0</td>
<td>S6</td>
<td>68±9</td>
</tr>
<tr>
<td>Dolgovoli</td>
<td>L6</td>
<td>-</td>
<td>868±60</td>
<td>S2</td>
<td>7.5±0.5</td>
</tr>
<tr>
<td>Kilabo</td>
<td>LL6</td>
<td>S3 [15]</td>
<td>262±10</td>
<td>S3,4</td>
<td>23±1</td>
</tr>
<tr>
<td>Kunya-Urgench</td>
<td>H5</td>
<td>-</td>
<td>928±100</td>
<td>S2</td>
<td>7.6±0.8</td>
</tr>
<tr>
<td>Tugalin-Bullen</td>
<td>H6</td>
<td>S1 [16]</td>
<td>575±25</td>
<td>S2</td>
<td>6.7±0.6</td>
</tr>
</tbody>
</table>

**Perihelion of the meteorite orbits.** The value of TL$_{\text{NAT}}$ stored by meteorites in the cosmic space can depend on an orbit of...
meteorites. It was assumed [17] that than less the perihelion of a meteorite orbit the above heating temperature of a meteorite and accordingly lower the value of accumulated $T_{LNAT}$. Hence, it is possible to suspect that the value of an irradiation dose of a meteorite in the cosmic space corresponds to value of a perihelion. For an estimation of perihelion value it was suggested to normalize intensity of a TL$_{NAT}$ of each sample to its sensitivity by measuring the value of TL$_{IND}$ per unit of a dose obtained from a radioactive source. This value indicated as an equivalent dose (ED) is calculated under the formula:

$$ED = D \times (TL_{NAT}/TL_{IND})$$

where $D$ is radiation dose of a sample in the laboratory. However, investigations suggest that it is more reasonable to calculate ED for two temperature intervals on the glow curves; $ED_{LT}$ at $T \sim 100-240^\circ C$ and $ED_{HT}$ at $T \sim 240-340^\circ C$. This allows us to reduce the error of ED estimate to $\leq 15\%$ and estimate more accurately the perihelion value.

### Table 2. Equivalent doses for the studied meteorites

<table>
<thead>
<tr>
<th>Meteorite</th>
<th>Type</th>
<th>Fell, Year</th>
<th>$ED_{LT}$, Gy</th>
<th>$ED_{HT}$, Gy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Barwell</td>
<td>L5</td>
<td>1965</td>
<td>382±10</td>
<td>1530±180</td>
</tr>
<tr>
<td>Dolgovoli</td>
<td>L6</td>
<td>1864</td>
<td>112±10</td>
<td>2260±290</td>
</tr>
<tr>
<td>Kilabo</td>
<td>LL6</td>
<td>2002</td>
<td>22±3</td>
<td>514±50</td>
</tr>
<tr>
<td>Kunya-Urgench</td>
<td>H5</td>
<td>1198</td>
<td>112±10</td>
<td>100±8</td>
</tr>
<tr>
<td>Tugalin Bulen</td>
<td>H6</td>
<td>1967</td>
<td>27±2</td>
<td>354±31</td>
</tr>
</tbody>
</table>

Procedure of an ED determination is more detail described in our papers [2, 11]. The values of $ED_{LT}$ and $ED_{HT}$ for the investigation meteorites are listed in Table 2. The measured value of TL$_{NAT}$ intensity in the meteorite of Chantonnay was lower than a limit of recording of a TL. The values of $ED_{LT}$ and $ED_{HT}$ listed in the Table 2. The dates exhibited in the Table 2 allow to suspect that the perihelion of the studied meteorites were within the limits of 0.8 - 1 A.U. The obtained perihelion sizes are typical for the orbits of the most of ordinary chondrites.

### References.

PEDESTAL CRATER DISTRIBUTION AND THE ROLE OF A LATITUDE DEPENDENT ICE-RICH REGOLITH.

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**Introduction:** Pedestal (Pd) craters, where both the crater and ejecta blanket are perched above the immediate surrounding terrain, are unique to Mars [1] (Fig. 1,2). Early attempts to identify a formation mechanism for these phenomena relied primarily on deflation: eolian erosion was proposed to remove the surrounding materials, leaving the crater and ejecta elevated. An armoring agent, perhaps related to increased impact melt production in volatile-rich impacts [2] or an atmospheric blast which strips away surface fines coupled with a thermal pulse that melts near-surface volatiles [3], prevents the ejecta blanket from receding entirely [4,5]. However, eolian deflation cannot easily explain the approximately circular planform of Pd craters, leaving the formation mechanism controversial [1,6].

In this study we investigate a new model of formation based on sublimation of an ice-rich, fine-grained surrounding terrain (Fig. 3). We use THEMIS VIS and IR data to determine that the distribution of all Pd craters with diameters >0.7 km between ~60°S and ~60°N. This distribution is used to assess whether Pd craters form in regions expected to have had near-surface ice during periods of high obliquity (>45°). In addition, we analyze physical attributes of the craters, including lobateness (\(\Gamma\)) and ejecta mobility (EM) ratio.

**The Role of Ice-Rich Material:** Viking data analysis revealed that Pd craters are concentrated in fine-grained materials. Recent proposals for a formation mechanism have taken advantage of the realization that many of the regions containing Pd craters may also be ice-rich [1]. The areas containing near-surface volatiles, as identified by the gamma ray and neutron spectrometer (GRNS), strongly correspond to regions proposed to have been covered in ice during past obliquity cycles [7-9]. Under current climate conditions, near-surface ice is unstable in equatorial regions. Ground ice at lower latitudes sublimates and water vapor diffuses through the regolith, dissipating into the atmosphere. It is eventually deposited at higher latitudes, primarily in the polar ice caps [7,10]. Models of the persistence of ground ice as a function of latitude and depth suggest that porous interstitial ice can persist at shallow depths due to recondensation of water vapor from greater depths. This can occur despite continuous diffusion into the atmosphere, creating a steady-state ice table [11]. Ice can then be maintained for geologically long timescales below the steady-state ice table. Pd craters excavating below this depth could entrap these volatiles in their ejecta.

If Pd crater formation requires ice-rich material, then we expect a correlation between Pd crater distribution and the history of Mars’ climate. Climate models show that during periods of high obliquity, increased insolation to the polar regions during the summer removes volatiles from the polar caps and deposits them at lower latitudes via snow or ice nucleated on dust [10,12,13]. Atmospheric humidity increases, and the latitude at which surface ice is stable moves toward the equator. During periods of low obliquity this latitudinal limit is usually around 60°, but at high obliquities the ice stability zone moves to ~30°. Most models also predict increased wind strength during the high obliquity eras, raising the atmospheric dust content. This dust is incorporated into the ice, potentially yielding several-meter-thick, fine-grained, ice-rich deposits between 30° and 60° [7,12,14]. When the obliquity decreases, the atmosphere dries and the ice-rich layer between 30° and 60° desiccates. The sublimated ice eventually returns to the poles, leaving behind an ice-poor regolith, although the generally short durations of low obliquity periods are not likely to completely remove the ice from the mid-latitudes [7,10,11,14].

**Pedestal Crater Distribution:** This study shows that Pd craters exist predominantly at latitudes thought to have had stable near-surface ice during periods of high obliquity (Fig. 4). In the northern hemisphere, Pd craters exist almost exclusively above 33°N, with the exception of an isolated field in the Medusae Fossae Formation between 7°N and 12°N. The majority exists between 45°N and 60°N. This latitudinal band of craters encompasses most longitudes with the highest and ejecta elevated. An armoring agent, possibly related to increased impact melt production in volatile-rich impacts [2] or an atmospheric blast which strips away surface fines coupled with a thermal pulse that melts near-surface volatiles [3], prevents the ejecta blanket from receding entirely [4,5]. However, eolian deflation cannot easily explain the approximately circular planform of Pd craters, leaving the formation mechanism controversial [1,6].

We see a strong correlation between the distribution of Pd craters and the hydrogen-rich regions measured by GRNS. Because GRNS data were acquired during a low obliquity period, any regions having high water-equivalent hydrogen concentrations were likely to have been even more saturated during high obliquity.

**Pedestal Crater Attributes:** Lobateness (\(\Gamma\)), a measure of ejecta sinuosity, is calculated from the ejecta perimeter (P) and area (A) [15]:

\[ \Gamma = P/(4\pi A)^{1/2} \]

Pd craters in this study have \(\Gamma\) values ranging from 1.01 to 2.49, with a mode of 1.05 and a mean of 1.1. \(\Gamma = 1\) indicates a perfectly circular ejecta blanket, while higher values correspond to greater sinuosity.

Ejecta mobility (EM) measures the ejecta extent as a function of crater size [16]:

\[ EM = (\text{ejecta extent})/(\text{crater radius}) \]

EM ratios of Pd craters in this study range from 1.18 to 13.2, with a mode of 2 and a mean of 3.09. The regions that show high \(\Gamma\) values also show high EM ratios.

The EM ratio is generally believed to reflect ejecta material fluidity at the time of ejecta emplacement [1]. Higher EM ratios suggest greater fluidity of material, corresponding to a greater concentration of subsurface volatiles. Generally, regions identified as having high EM ratios and lobateness values coincided with areas of high water content within the soil as identified by GRNS.

**Discussion:** The results of this study suggest that Pd craters form from small impacts into ice-rich material deposited during high obliquity periods. During low obliquity periods, ice sublimates from the fine-grained material, leaving the crater and ejecta blanket perched above the surroundings. This effect is most pronounced for smaller craters which form almost entirely within the fine-grained volatile-rich layer. This model is supported by the following observations: (1) correlation between Pd crater distribution and regions where ice-rich mantling material would be deposited during high obliquity periods, (2) small size (typically ~2-km-diameter) of Pd craters, (3) high EM and \(\Gamma\) values, suggesting the ejecta was volatile-rich during emplacement, and (4) correlation of Pd craters with the distribution of volatile-rich near-surface materials revealed by GRNS.

**References:**

Figure 1. A section of THEMIS VIS image V11743004 (78 m/pix) of a field of Pd craters. The image is centered at 56°N, 84°E. All Pd craters visible have diameters <2 km. Note the circular planform of the craters.

Figure 2. MOLA data superimposed on a THEMIS VIS image, showing a Pd crater and its ejecta blanket elevated above the surrounding terrain. The profile of the crater from the MOLA track has a VE of ~36X.

Figure 3. A four-step conceptual model for Pd crater formation: 1) Ice-rich regolith. 2) Impact distributes ejecta and impact glass, indurating the surface around the crater. 3) Armored ejecta blanket overlies ice-rich material. 4) Sublimation of ice from surrounding terrain leaves perched Pd crater.

Figure 4. A hillshade map of Mars derived from MOLA data, showing the distribution of Pd craters identified between 60°N and 60°S. With the exception of the MFF, Pd craters in the northern hemisphere do not form equatorward of 33°N. In the southern hemisphere, Pd craters exist almost exclusively poleward of 40°S. The general absence of Pd craters in equatorial regions is likely due to a paucity of ice-rich material at those latitudes, even during periods of high obliquity.
THE ASCRAEUS MONS FAN-SHAPED DEPOSIT: EVIDENCE FOR SUBGLACIAL VOLCANISM.
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Introduction: Fan-shaped deposits (FSD) extending to the NW of each of the Tharsis Montes have been interpreted to be a series of glacial deposits [1-4]. Geomorphological observations, supported by atmospheric and glacial modeling, suggest that the accumulation and sublimation of snow and ice produced the cold-based glaciers responsible for the deposits. The glaciers formed in the Mid- to Late-Amazonian [2,5,6] during periods of high obliquity [7], when increased polar insolation caused sublimation of the icecaps [4]; volatiles were transported to lower latitudes and deposited via precipitation of ice and snow, in this case on the NW flanks of the Tharsis Montes [4]. Volcanism activity associated with the Tharsis Montes occurred throughout this period [8-10], so it is anticipated that eruptions might interact with these glacial deposits.

The Ascraeus FSD consists predominantly of two facies: ridged and knobby; it lacks the smooth facies seen at Arsia and Pavonis. In addition to these two facies, there are a number of features which are indicative of lava-ice interactions. In this analysis, we use THEMIS and MOLA data in an effort to describe observed geomorphologies which suggest subglacial volcanic activity.

Flat-topped Ridge Region: “Lobate flows” [11] covering an area of approximately 2300 km² form a distinct feature extending northwards – normal to regional and local slopes – from a position proximal to the southwestern flank of the Ascraeus Mons shield base (Fig. 1). This lobate feature consists of at least three broad, elevated plateau features with steep margins that rise up to 500 m above the surrounding surfaces. Plateau surfaces are marked by segments of wide, leveed channels, and the two largest lobes have medial channels. At the distal end of the medial channels, smaller lobate features emerge and follow local topography. These characteristics make the lobate flows distinct from the adjacent lava flows to the south and west, which lack steep margins and wide levees. The upslope (southern) end of the feature is truncated by lava flows originating from troughs parallel to the shield base of Ascraeus [11], while the downslope (northern) end terminates abruptly at the scarp. There is, however, a small fraction of this northern end of the flow feature that appears to head in a westward direction, connecting to the southern end of the ridged facies (Fig. 1).

Flow-like features morphologically similar to these have been identified in the Pavonius FSD [2,9,11]. Based on Viking data, they were originally described as elongate sinuous ridges and hypothesized to be eskers [9]. Their broad, elevated morphology, however, is not consistent with typical terrestrial esker morphology [e.g. 12,13]. Furthermore, eskers are associated with wet-based glaciers and would therefore not be expected at the site of a cold-based glacier. Examination of Arsia and Pavonis has revealed no evidence, such as sinuous channels, lake deposits, and/or braided streams, for extensive wet-based glacial activity [1,2]. More consistent with the lobate nature of the features, their elevated topography, and their apparent relationship with fissures on the southwest flank of Ascraeus [11] is an alternative hypothesis that these features are distinct types of lava flows [9]. Zimbelman and Edgett (1992) interpret these features as pyroclastic flows that originated from troughs in the lower flanks of the volcanoes. They suggest that the effusive activity that produced these postdated the emplacement of the FSD and was triggered by the removal of a lithostatic load accompanying a large landslide event that would have formed the FSD. However, such a proposition is inconsistent with the observation of the knobby facies, interpreted to be sublimation till, superimposed on the lobate flow features at Ascraeus Mons and over comparable flow features at Pavonis Mons [2].

Plateau Feature: A relatively flat-topped, elevated plateau exists ~7 km west of the shield base in the central portion of the deposit (Fig. 2). The mesa-like feature, which is 34 km long and 19 km wide, is stepped in nature with two terraces, both lobate and elongated to the north, generally normal to the regional northward slope. Analysis of the topography reveals the presence of a small mound, a few tens of meters high, on the southeast end of the plateau. A narrow ridge extends from the base of the plateau at the southern end of the mound to the south for ~20 km and disappears under the flat-topped flows. Unlike the lobate flows farther south, the upper surface of this flow-like feature lacks any evidence of leveed channels. Surfaces do appear to be degraded, particularly on the western side where the terraced surfaces are observed, both of which are covered by knobs constituting the southern portion of knobby facies (Fig. 2).

The plateau is significantly more tabular in planform relative to the elongate, lobate morphology of the southern flow features. The tabular surface is reminiscent of tuyas or table-top mountains found in Iceland and north central British Columbia, which form via subglacial volcanic activity. We believe that the sinuous ridge leading to the plateau may be a dike intruding into the ice, and the mound at the SE end of the plateau is the eruptive center (Fig. 2). The terraces are thus interpreted to represent subglacial eruptions that formed marginally-chilled flows which headed north, downhill. A subglacial eruptive mechanism for the formation of the lobate flow features was one of the explanations proposed for comparable features observed within the Pavonis FSD [2]. Similar proposals were made for plateau-like and ridged deposits in the Arsia Mons fan-shaped deposits [14], where evidence for associated meltwater distribution is seen (e.g. local eskers, glacial surging and distal fluvial channels).

Discussion: On the basis of the stratigraphic relationships, the geomorphology, and the topography, we interpret both of these features to be due to lava-ice interactions. We prefer the interpretation of these lobate, steep-sided flow features as subglacial lava flows (non-pyroclastic), and suggest that they formed concurrent with the presence of a glacier west of Ascraeus Mons. These subglacial flows superpose older subaerial flows which were deflected around the glacier, forming an outer arcuate scarp. During a subsequent glacial advance which extended farther south, additional flows, this time subglacial, formed the steep-sided, flat-topped features. The observation of the knobby facies superposed on this lobate flow feature is consistent with our interpretation, supporting the rapid sublimation of the glacier after the flows were emplaced. Based on these relationships and the observation that the knobby facies also covers a portion of the second more centrally located plateau feature, it is likely that both flow features formed prior to the knobby facies in the southern half of the deposit.

Conclusions: The FSD extending to the NW of Ascraeus Mons, which we believe to be the result of cold-based glaciation, shows two morphologies indicative of subglacial volcanism: (1) Flat-topped ridges with steep margins and leveed...
channels, and (2) a terraced plateau feature with a distinct mound and an arcuate ridge leading into it. We interpret the flat-topped ridges to be subglacial lava flows, which are comparable to subglacial flows observed in the Pavonis FSD [2]. The plateau is a table-mountain, with an eruptive center at its SE end and an associated dike extending to the south. The knobby facies is superimposed on both of these features, suggesting sublimation of the overlying glacier after the features formed.

**Introduction.** Fossil track study for the silicate minerals of the most abundant meteorites ordinary chondrites is one the essential source of quantitative information about the early Solar system processes and environment condition. In early our works [1, 2] there were obtained results, indicated some particularities of the cosmic-ray (CR) VH-nuclei tracks in ordinary and carbonaceous chondrites no enriched by the solar type gases. The present report contains new track data obtained for the some of recently falls of chondrites.

**Samples and experimental procedure.** The two recent falls (2001-2002 yr) of meteorites of Bukhara CV3 [3] and Kilabo LL6 [4] were presented in our investigation by 33 and 67 silicate crystal grains, correspondingly. The individual olivine and pyroxene grains acceptable for the fossil track investigation were extracted from microcrystallitic matrix of chondrite samples by weight of 0.3-0.5 g. The silicate grains of 50-200 μm size fraction were used for study. Crystals were mounted in epoxy resin, polished and underwent to chemical track revealing etching by the well-known procedure [5, 6]. The next track parameters were measured by optical microscope techniques: 1) track-density (ρ) values in individual silicate grains; 2) statistical ρ_{min}, ρ_{med} and ρ_{max} values; 3) revealing and measuring of grains with the surface-depth ρ-gradient.

**Results and discussion.** The characteristic track parameters of the grains under investigation, positioned with each other inside of mm-distance in the meteorite body, were determined: a) A range up to three orders of magnitude of the track density values, which were measured in the individual silicate grains, extracted from samples; b) A range of ρ_{min} values observed for the no smaller of one fourth part of grains under investigation, that corresponds to the VH-nuclei of the galactic CR flux for this meteorite and grain localization in the preatmospheric meteorite body; c) The ρ values discrepancy for the silicate grains, separated from the same samples; d) The average ρ values that in comparison with the data, observed in the solar gas-rich chondrites, achondrites and especially lunar regolith samples are essentially lower.

The statistical track density distribution for the olivine and pyroxene grains are characterized by the histograms, presented in Figure. It is seen that the investigated samples have very specific modes of grain number vs track density distribution. It can be seen that the Bukhara CV3 crystal samples under investigation for which there is no observed the tracks concerned with pre-accretion exposure stage and practically total radiation effect due to the galactic CR VH-nuclei. But in chondrite Kilabo LL6 the pre-accretion VH-nuclei exposure traces are sharply defined: in the total track density interval of (5-10^4 ≤ ρ ≤ 2-10^5) cm^-2 clearly may be observed the two crystal groups, the average gaussian (avg) values for which are: (ρ_{avg})_{min} = (2.1 ± 0.4)×10^5 cm^-2 and (ρ_{avg})_{max} = (5.0 ± 0.5)×10^6 cm^-2. Besides, it is important to note for this chondrite that corresponding statistical parameter N_{max}/N_{min} ≈ 0.6 (N_{max} and N_{min} are numbers of the crystals in the high and low of the track density groups, accordingly) for obtained crystals distribution indicate very high pre-accretion irradiation probability.

**Discussion and conclusion.** Obtained data suggested a certainly different pre-accretion conditions of the primary chondritic matter irradiation in the two examined chondrites. In the contrast to the simple, one-stage radiation history for the Bukhara CV3 chondrite one for the Kilabo LL6 chondrite is more complex. Track parameters determined in our work for this meteorite could be best understood in the term of the pre-accretion VH-nuclei cosmic-ray irradiation that occurred in the primitive solar nebula in the gas-dust cloud environmental conditions. The flux value and estimated effective energy spectrum of these VH-nuclei indicate the possibility of locally accelerated multiple charged Fe-group ions which were contained in thermalyzed solar wind flux [7, 8]. These ions are accelerated up
to energy $E > 100$ MeV/nucl in the early active Sun.

Also, across it the principally distinguishable point of view can be considered taken in account the results of the rare-gas isotopes and cosmogenic nuclides analyses [9]. Two base stages of the cosmic ray irradiation are considered in this approach: the beginning in the parent meteorite body at the more than meter depth from the surface during of 30-40 my, and then in the meteoroid body at the comparatively small depth during 5 my.

According to the tracks of VH-nuclei, the shielding depth of the investigated sample of the Kilabo chondrite is $6 \pm 3$ cm.

References:


Fig. Track density distributions in the Bukhara CV3 and Kilabo LL6 chondrites.
TOPOGRAPHY OF SELECTED LUNAR AREAS FROM SMART-1 AMIE DATA. V. Kaydash1, M Kreslavsky1, Yu. Shkuratov1, S. Gerasimenko1, P. Pinet1, S. Chevrel1, J.-L. Josset2, S. Beauvivre2, B. Foing5
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Introduction: We report on a photogrammetric study of selected lunar regions using images obtained by the Advanced Moon Micro-Imager Experiment (AMIE) camera onboard SMART-1 spacecraft in 2006. We used algorithms that we have applied earlier for mapping topography of the Apollo-17 landing site with combined Hubble Space Telescope and Clementine images [1]. Here we study Gruithuisen volcanic domes and crater Lavoisier, and estimated the thickness of Aristarchus ejecta blankets.

Processing of AMIE multiangular spot-pointing data: In the spot-pointing regime, SMART-1 imaged several lunar areas in the wideband spectral filter under the same Sun illumination geometry but varying the viewing angle [2]. We converted raw image data counts to the values proportional to the surface brightness with the preliminary pipeline calibration [3]. Then we reprojected individual frames to the same Mercator projection using up-to-date versions of SMART-1 SPICE kernels [4]. Spatially coregistered image pairs reveal mutual parallactic shifts of surface features due to the difference in views of local topography. We found these shifts using subpixel coregistration of images and then transformed the shifts into relative elevation knowing the viewing geometry. Absolute heights relatively to the selenoid needed to be defined from additional data. The precision of the derived elevations is controlled by horizontal spatial resolution of the source images. For sites studied here this precision was typically of ~100-150 m.

Gruithuisen domes. The area we studied (centered at 39.5 W, 35.8 N) covers Gruithuisen domes supposed to be of volcanic origin of Imbrian period [e.g., 5]: the δ dome, southern flank of γ dome, crater Gruithuisen B and surrounding area in western part of Mare Imbrium (Fig. 1a). Several successive images from the orbit 2236 with the Sun incidence angle 45º, viewing angle 4-18º and the resolution ~100 m/pix are used for the photogrammetric analysis. Fig. 1a presents our topographic map for this area with color-coded relative elevations in km. An albedo image is used as a background. Positions of Clementine Lidar (CL) footprints available for this area [6] are shown with stars. The linear regression for elevations from AMIE vs. CL is shown in Fig. 1b with error bars ±40 meters, as was reported for CL data [7], and ±100 meters for our data. The value of the regression slope ~0.66 reveals a discrepancy of two data sets. A possible reason for such inconsistency is sparse coverage of the δ dome and its vicinity by CL bounce points. The site (~2.11 km) located at the northern flank of the dome largely affects the slope of regression in Fig. 1b. The topography map shows a deep crater (Gruithuisen B) with inclined rim. The total elevation of upper rim relatively to the crater floor is ~2 km; this is greater than the 1.3 km value published in the lunar chart LAC 23 [8]. Flat top of the δ dome shows the maximum elevation of ~1.9 km relatively to adjacent mare surface, this estimation is also higher than the LAC 1.66 km value [8]. Northern flank of the Gruithuisen ζ dome is seen just at the bottom of the topography map, it shows maximal elevations of 1.25 km above the mare level.

Fig. 1a. Topography map for Gruithuisen Hills area. Relative elevations are color-coded in km. Stars mark all CL footprints in the area.

Figure 1b. Correlation between derived AMIE elevations and CL ranging for the area in Fig. 1a. Linear regression (bold line) parameters are presented in the upper inset.
**Crater Lavoisier.** A large (70 km) pre-Nectarian cracked-floor crater Lavoisier (80.8 W, 38.2 N) was imaged in orbit 2251. We used images with the incidence angle 45º, viewing angle 0º and 25º, and the resolution ~120 m/pix. Our topography map (Fig. 2a) shows the largely eroded Lavoisier rim with maximal height ~1.2 km above the lowest point of the crater floor. The northern part of the floor is raised up to 0.3 km relatively to its southern part. The local rise (~200-300 m) near the center of the crater Lavoisier carries two 5-6 km size craters. The southern one is double-shaped, indicating layered structure of the Lavoisier floor. The floor of the northern small crater appears as a deep place located at ~0.5 km below the adjacent Lavoisier floor. Tectonic ridges as well as dark spots (lava flows) in the crater do not reveal any prominent topography. Comparison of CL profile (stars in Fig. 2a) with the derived topography shows the regression slope close to unity (Fig. 2b).

**Fig. 2a.** Topography map for crater Lavoisier. Relative elevations are color-coded in km. Stars are CL footprints.

**Fig. 2b.** Correlation between derived AMIE elevations and CL ranging for the area in Fig. 2a. Linear regression (bold line) parameters are presented in the upper inset.

**Eastern Aristarchus.** The eastern part of crater Aristarchus and proximal ejecta apron were imaged in orbit 2238 with a photometric study underway [9]. The incidence angle for the images used was 36º, the viewing angle changed in the 2-30º range, and the resolution was 100 m/pix. Fig. 3a presents our topography map for eastern part of the crater Aristarchus. We were not able to compare the derived topography with CL data because of lack of them for this area. The calculated topography shows a high rise of the Aristarchus rim and a small part of its floor (the floor-rim elevation difference is ~2 km). We note different elevations of ejecta deposits in the proximal zone (~1 radius of the crater Aristarchus). The thickness of ejecta blankets is ~300-400 m.

**Fig. 3.** Topography map for Eastern Aristarchus crater area. Relative elevations are color-coded in km.

**Conclusions:** We carried out a photogrammetric analysis of SMART-1 / AMIE camera images. We constructed topographic maps with ±100 meters vertical accuracy for three lunar areas. Our analysis yields a 3-D model of the volcanic dome Griethuisen δ, the structure of the crater Lavoisier floor and rim, and allow an estimation of ejecta blankets thickness in the proximity of the crater Aristarchus.

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**PEDESTAL CRATERS NEAR APOLLINARIS PATERA: FORMATION, DISTRIBUTION, AND IMPLICATIONS.**  L. Kerber and J.W. Head, 'Department of Geological Sciences, Brown University, Box 1846, 324 Brook St., Providence, RI 02912 (laura_kerber@brown.edu).

**Introduction:** Pedestal craters have sometimes been interpreted as a marker for subsurface volatile-rich layers [1]. While pedestal craters are common in the polar latitudes of Mars, they also appear in units nearer to the equator, such as the Medusae Fossae unit [2]. The pedestal craters that appear in the Medusae Fossae differ from those found at high latitudes in several ways, particularly that the raised ejecta blankets usually have an asymmetric, urchin-like morphology, as compared to polar pedestal craters which exhibit a far more circular shape. With the aid of high resolution data (a combination of CTX images, HRSC images, and THEMIS images), it is revealed that large numbers and high densities of pedestal craters with small (<1km) diameters and highly asymmetric shape can be found in the vicinity of the medium-sized volcano Apollinaris Patera (-8°S, 174°E), on the edge of the Medusae Fossae region, and that the distribution of these pedestal craters varies greatly according to where they are found on the volcano.

**Results:** The pedestal craters found around and on Apollinaris Patera show particularly well the crater erosion progression described by Schultz [3,4] (Figure 1). Additionally, while these pedestal craters differ in a consistent way from their polar counterparts, they also show differences from one another depending on the unit in which they occur and where they are situated on the volcano. Greeley and others [5] noted that polar pedestal craters were likely not a result of wind deflation because their “pedestals” would be expected to appear asymmetric and line up with regional wind patterns. Pedestal craters found on the northwest side of the volcano are asymmetric and line up well with fields of yardangs, which indicate the direction of prevailing winds. (Figure 1. B, C, and D). Thus instead of requiring the deflation of the surrounding terrain by sublimation, these pedestal craters remnants were likely the result of primarily eolian deflation. This difference in process may account for the morphological differences between these crisply-outlined asymmetric pedestal craters and the somewhat more subdued and symmetrical remnants in the polar latitudes, where the process of removing volatiles could have softened the outline of the ejecta blanket over time.

In the caldera of Apollinaris, which provides some shelter, pedestal craters can be seen emerging from a heavily cratered (close to saturated) parent unit. (Figure 1A). As it erodes, some pedestal craters survive to...
become free-standing edifices, while others are destroyed in the erosion process. This area demonstrates how heavily cratered surfaces can be preferentially erased, revealing a young-looking surface below. Even if remaining pedestal craters and knobs were counted in order to determine the age of the unit, the final number would still be a minimum age for the visible terrain, as well as being a minimum residence time for the mantling unit that formerly covered it.

On the eastern side of the volcano, a unit interpreted by Scott and others [6] as a deeply eroded Apollinaris lava flow was found to be replete with pedestal craters, many below the current HRSC resolution for the area but visible in CTX. The pedestals surrounding these craters are somewhat preferentially preserved on the southern side. This unit is the same in which the landslides on the eastern flank have taken place. It is possible that the unconsolidated ash or pyroclastic deposits that are thought to make up this side of the volcano [7] are not only prone to mass wasting events, but also provide an excellent platform for pedestal craters to form because they can be welded or cemented into an armored ejecta blanket by high impact atmosphere temperatures (as described by [8]) and the surrounding deposit would remain fine grained and unconsolidated and thus easily removed by the wind. Unlike other areas of the volcano with pedestal craters, the parent unit for these craters is not obvious.

Pedestal craters appearing on the Apollinaris Patera’s large southeast fan are being derived from a textured unit which has been described by Ghail and Hutchison [9] as depositional. It appears in the interiors of craters as well as in hollows on the slope of the fan. Because of the presence of the pedestal craters, this unit, while it may have been depositional in the past, is likely now eroding, remaining only in low places that are sheltered from the wind, as well as in the armored ejecta blankets of pedestal craters.

Significantly, pedestal craters appear in abundance in some units and are sparse or absent in others. This unequal distribution could have several causes, specifically that units containing pedestal craters were the only units that were: 1) covered with a water or ice-rich mantling unit 2) not subject to so much erosion that all remnants of the pedestals have been removed, and 3) subject to enough deflation that the pedestals were revealed. The unit must also avoid being covered up by a later mantling unit or dust cover deep enough to cover the pedestal craters.

Discussion: There are many different hypotheses concerning the origin of the Medusae Fossae Formation. It has been traditionally mapped as Amazonian [6,10] but Schultz [3,4] has suggested that it could be as old as Hesperian in age because of its misleading erosional regimes.

The remnant knobs and domes present in the Medusa Fossae, if they are craters, would make the surface much older than previously supposed. If the deposit were Hesperian as opposed to Amazonian, it is possible that they could be deposits from Apollinaris in the form of ashfall and/or pyroclastic deposits.

Summary and Conclusions: The pedestal craters that surround Apollinaris differ in many ways from the traditional pedestal craters of the polar regions. This difference could be caused because there is a different agent of deflation (eolian instead of sublimation) or because the mantling unit out of which the pedestals are being eroded has properties different from mantling units in the polar regions. They show an erosional progression from a normal crater to remnant knob. Differences in their morphology and distribution can be used to help piece together the history of the Apollinaris area.

![Figure 3. Distribution of pedestal craters on and around Apollinaris Patera. Left leaning images are CTX images. Right leaning are THEMIS images. While resolution becomes an issue between image types, especially at HRSC scale, differences between units are also evident within single images.](image-url)
Introduction: As NASA implements the nation’s Vision for Space Exploration [1] to return to the moon and travel to Mars, new considerations will be be given to the processes governing design and operations of manned spaceflight. New objectives bring new technical challenges; Safety will drive many of these decisions.

Historical Context: For the Apollo program, safety requirements were individual standards for individual subsystems or components. The Space Shuttle Program (SSP) combined these requirements into a uniform standard, “Fail Operational, Fail Safe” or FO/FS. If any one failure occurs then the Shuttle can continue to operate and complete all flight mission objectives. Following a second failure, FO/FS specifies that the vehicle is Safe, able to return home safely, though may not be able to complete all mission objectives.

Payload Safety Process instituted Two Fault Tolerant (2FT) Requirements to add an additional layer of protection between Crew/Shuttle and Payloads.

The International Space Station (ISS) based their safety analysis on the Payload standard of 2FT, motivated by the lack of ground servicing of the ISS following failures of safety-critical hardware. The Shuttle has the option to perform an emergency deorbit, in the event of a major failure. There is no similar contingency option for the ISS.

Safety Requirements Allocation. System engineering calls for breakdown of requirements by function and allocation of portions of requirements to separate subsystems. Unfortunately, safety requirements can not follow this model. Safety requirements generally are required to be applied to all of the lowest level specifications intact.

Achieving 2FT. There are numerous ways to achieve satisfactory compliance of requirement. One could use independent, triple-redundant must-work systems or independent three inhibit most-not-work systems throughout the vehicle. Unfortunately, such design standard is not possible due to restrictions of weight and volume. Additionally, it is not necessary as there are other ways to meet 2FT: Use of Unlike Redundancies. Separate systems working together can provide an overall higher level of safety. Use of Other Available Margins, taking advantage in margin in a different system to meet 2FT requirement.

Non-Compliance of 2FT. Alternate approaches, when necessitated by design are acceptable.

Competing Must-Work and Most-Not-Work Functions. Consider rendezvous. No single or double events can result in collision. Crew controlled flight via joystick is inherently vulnerable to crew error. Instead, training, design of approach corridor and control of approach speed allow crew to recognize errors and adjust.

Equivalent Safety. This is used when risk to Crew/Vehicle is very low with just Single Failure Tolerance Implemented. The emphasis is on other controlling factors and not just probability numbers. For example, a single failure tolerant system with large time to effect is considered acceptable ‘equivalent’ risk to a 2FT system because there is plenty of time to avert undesired effect after the failure using operational controls.

Design for Minimum Risk (DMR): From the ISS Safety Requirements [2]: Design for minimum risk are areas where hazards are controlled by specification requirements that specify safety related properties and characteristics of the design that have been baselined by the ISS program requirements rather than failure tolerance criteria. The failure tolerance criteria … shall only be applied to these designs as necessary to assure that credible failures that may affect the design do not invalidate the safety properties of the design. Examples are mechanisms, structures, glass, pressure vessels, pressurized lines and fittings, functional pyrotechnic devices, material compatibility, flammability, etc.

Safety Analysis Documents. Safety engineering analyses are documented in Hazard Reports, Failure Mode Effects Analysis (FMEA) and the Critical Items List (CIL) [3].


The Severity level is an assessment of the worst-case effects of a hazard for a given cause. By definition, Catastrophic severity is a hazard which could result in a mishap causing fatal injury to personnel and/or loss of one or more major elements of the flight vehicle or ground facility. Critical severity is a hazard which could result in serious injury to personnel and/or damage to flight or ground equipment which would cause mission abort or a significant program delay. Marginal severity is a hazard which could result in a mishap of minor nature inflicting first-aid injury to personnel and/or damage to flight or ground equipment which can be tolerated without abort or repaired without significant program delay.

The Likelihood of Occurrence assesses the probability that the worst-case hazard will take place, with the controls in place. Probable: expected to happen in the life of the program; Infrequent: could happen in the life of the program, controls have sig-
significant limitations or uncertainties; Remote: could happen in the life of the program, but is not expected, controls have minor limitations or uncertainties; Improbable: extremely remote possibility to happen in the life of the program; there are strong controls in place.

The risk matrix places each cause into a grid in a matrix, with the Likelihood of Occurrence on the vertical axis and the Severity on the horizontal axis. For the sample Risk Matrix with three causes (Fig. 1), two causes (A&C) are Remote/Catastrophic; one cause (B) is Improbable/Catastrophic.

Failure Mode Effects Analysis (FMEA) and the Critical Items List (CIL). FMEA/CILs identify specific potential hardware failures and describe the root cause of failures for components and subsystems. [5]

Each FMEA/CIL is assigned both a ‘functional criticality/hardware criticality’ as follows: ‘1/1’ Single failure which could result in loss of life or vehicle; ‘1R/2’ Redundant hardware item(s), all of which failed, could cause loss of life or vehicle. First failure has no effect on mission; second failure may result in loss of mission; ‘2/2’ Single failure which could result in loss of mission; ‘2R/3’ Redundant hardware item(s), all of which failed, could cause loss of mission. First failure has no effect; ‘3/3’ All others.

FMEAs are included on the CIL if they are of criticality ‘1/1’, ‘1R/2’, and, for some cases, ‘1R/3’.

Spacecraft Operations Considerations. During Shuttle Flight Operations, the Mission Management Team (MMT) assesses on-orbit anomalies. Engineering teams review relevant FMEA/CILs and Hazard Reports. The relative positions in the risk matrix are used to guide actions to protect against the highest risk. Each anomaly is assigned a position on the Anomaly Risk Matrix, shown in Figure 2. The Fault Tolerance Remaining compared to the Next Failure Consequence are tracked. Subsequent to the anomaly, it is frequent that the position in this matrix moves. That can be because the mission circumstances change and the next failure consequence becomes less significance or it can be moved because more flight data is obtained or further analysis has been performed.

ISS operations are handled slightly differently. Following an anomaly, the engineering investigation team assesses Magnitude of Potential Consequences compared to Likelihood Probability the corresponding condition or event will happen. Subsequently, the team presents options and identifies the associated risk and the corresponding reliability after implementation.

Conclusion NASA has continued to develop processes for design and operations to flight safety. The Vision for Space Exploration will bring new technical challenges and necessitate new approaches to design and operations.

Atmospheric wave granulation in the solar system: the star – planets – satellite. Kochemasov G.G. IGEM of the Russian Academy of Sciences, 35 Staromonetny, 119017 Moscow, Russia, kochem@igem.ru

Up to now there are 7 rather well studied atmospheres in the solar system: Sun’s photosphere, Venus, Earth, Mars, Jupiter, Saturn, Titan. They have differing radii, thickness, masses, densities, compositions, physical states, belong to celestial bodies of three types, but one property unites them. They are structurized by inertia-gravity waves (as well as their lithospheres) and obey one law of the wave planetology [1, 2, 3 & others]: higher orbital frequency smaller atmospheric granules and, vice versa, lower orbital frequency larger granules. “Orbits make structures” – this three word sentence is an essence of the wave planetology – the only science uniting all so different heavenly bodies on a basis of their orbital properties. Always present orbital eccentricities and frequencies and body rotations are main reasons for their wave structurization. All mentioned atmospheres demonstrate this rather clear. Arranged in a row of diminishing orbital frequencies they show increasing relative atmospheric granule sizes. The relation frequency – size is scaled to the photosphere: 1/month – πR/60 or Earth: 1/year – πR/4 (R – a body radius).

The saturnian atmosphere rotating or orbiting the center of the saturnian system with period of 10.2 (10.8) hours (frequency 1/10.2 – 1/10.8 h.) reveals in the IR radiation under clouds a vague scarcely resolvable fine granulation comparable with a grainy sandstone texture (PIA08934). A size of separate sand particles is about 50 to 100 km. This size is comparable with the theoretical one – 55 - 61 km (πR/3488–πR/3082) (Fig. 5).

The jovian atmosphere rotates (or orbits the center of the jovian system) with the period of 9.9 hours (frequency 1/ 9.9 h). The theoretical granule size is πR/3539 or 63 km. These grains or spots can be detected in the high resolution Galileo’s P-47938 BW images (415 & 886 nm filters) [4, fig. 1, 2, 7].

The venusian atmosphere rotates or orbits the center of Venus with the period of 4 days (frequency 1/4d.). Corresponding granule size is 65 km (πR/295). Measured granule size (PIA 00072) is about 50 to 80 km (or dark nodules like “beads on a string” ~ 100 km across according to PIA00072 – a Galileo image) (Fig. 4).

Titan orbits Saturn (and rotates) in 16 days. Corresponding granule size is 88 km (πR/91) what suits nearly perfectly to observations (IMG001101-br500) (Fig. 3).

The solar photosphere rotates (or orbit the center of the solar system) with a monthly period (frequency 1/1 m.). Corresponding granule size is about 30 to 40 thousand km (πR/60) what matches well with sizes of long ago known solar supergranules (Fig. 2).

The Earth’s atmosphere and lithosphere orbiting frequency around Sun is 1/365 days. This gives granule size πR/4 or about 5000 km across (Fig. 1) what is observed in lithosphere and sometimes in atmosphere where weather systems (anticyclone & cyclone) reach this dimension. Much higher atmosphere orb. fr. around the Earth’s center (rotation) gives granule size πR/1460 (~14 km) – similar to tornado cyclone or mesocyclone.

The martian theoretical granule size πR/2 (orbital frequency 1/687days) gives two bulges separated by two hollows in a big circle what is observed in the solid body and atmosphere. Dust devils could mark smaller atmospheric grains due to martian rotation (πR/1340, ~8 km across).

Along with described grain sequence granules of other sizes simultaneously exist in atmospheres. They represent waves due to other orbits as satellite Titan and atmospheres of planets move not only around centers of their planetary systems but at the same time around Sun. These low around planets frequencies with production of side frequencies and corresponding waves and granules [5 & earlier publications]. For examples, there are such granules at Saturn (πR/460, “leopard skin”, PIA08333, Fig. 7 [5], PIA09001, “cloud phantoms”), Venus (πR/49, PIA00073, Fig. 6 [6]), Titan (πR/12, the Hubble ST image at the pre-Cassini era [7]), Earth (πR/365 = 55 km, actually typical marine stratocumulus cells are 15-45 km The modulation strictly witness for wave processes involved in structurization of the Solar system’ bodies.


Fig. 1. Earth, PIA00729, South polar projection, mosaic of Galileo images, regularly spaced weather systems (πR/4 grains) are visible.
Fig. 2. Sun, supergranulation, πR/60.
Fig. 3. Titan, atmospheric granulation, IMG001101-br500, πR/91
Fig. 4. Venus, PIA00072, Galileo image, “beads on a string”, πR/295.
Fig. 5. Saturn, a portion of PIA08934, grainy “sandstone” texture, πR/3082
Fig. 6. Venus, PIA00073, near IR Galileo image, granulation πR/49.
Fig. 7. Saturn, PIA08333, South pole, IR image, “leopard skin” spots, πR/460.

Images credit: NASA/JPL
Plato' polyhedrons as shapes of small satellites in the outer solar system. G.G. Kochemasov
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The wave planetology [3-8 & others] main assertion is: "Orbits make structures". As all celestial bodies move in non-round (elliptical, parabolic) orbits with periodically changing accelerations they all are subjected to an action of inertia-gravity forces. These forces arouse in them warping waves that in rotating bodies (but they all rotate!) acquire a stationary character and 4 directions of propagation (ortho- and diagonal). Interferences of these waves produce three kinds of tectonic blocks: uplifting (+), subsiding (-) and neutral (0). Their size depends on warping wavelengths. The longest fundamental wave 1 produces ubiquitous tectonic dichotomy – an opposition of two segments: uplifted and subsided, expanded and contracted ($2\pi$R-structure). The first overtone wave 2 superposes on this segmentation smaller features - sectors ($\pi$R-structure). Next overtones give smaller features.

An essence of tectonic dichotomy is in tendency of 4 interfering waves 1 to make from a body a tetrahedron – the simplest Plato' figure. A dichotomous nature of this figure is revealed in opposition of a vertex and a face (cutting any of its 4 axes one always gets from one side a vertex, from another a face). In one direction three faces narrow towards a vertex (contraction), in opposite direction they expand towards a fourth face (expansion). Most often in small bodies (not only in satellites but also in asteroids and comets) one observes an oblong convexo-concave shape [9 & others] but sometimes at certain points of view a flatten concave side and a sharpened convex side are presented by such a way that a tetrahedron develops (Fig. 3). Interfering waves 2 produce an octahedron. At the first time it was observed in a shape of Amalthea (see Kolva’s drawing of this satellite after Galileo mission), and name "diamond" was pronounced but no explanation followed. Now some octahedron faces one can observe at a number of small bodies, just to mention Phobos, Phoeba, Yanus (Fig. 5). Interfering waves 4 produce a cube (Fig. 6).

Shorter wavelengths – more vertices in a polyhedron: tetrahedron 4, octahedron 6, cube 8 and so on. Various polyhedrons are present in a body simultaneously because the wave warping occurs in various wavelengths at the same time but particular view points present better view of one of them (for examples, Amalthea, Hyperion, Helene).

An experimental confirmation of geometrization of a sphere by standing waves is presented in Fig. 1. In Fig. 2 shapes of native copper crystals are shown for comparison with shapes of small cosmic bodies. Dichotomous structures of larger satellites keeping spherical shapes because of enhanced gravity but having differing chemistry of opposite hemispheres are shown in Fig. 4. In Fig. 7 are shown typical hexagons developed in cosmic bodies due to intersections of three directions representing three faces of a structural tetrahedron (see above). The images of cosmic bodies credit: NASA/JPL/Space Science Inst., University of Arizona.
Fig. 1. Experiment on warping an air bubble by standing waves. Right – an air bubble diameter ~0.1 mm in water. Left – its changed shape after applying sound waves with a frequency ~24.5 khz. The shape tends to be cubic (a square in section) because 4 waves are inscribed in a circumference. Other wavelengths produce other forms, for an example, tetrahedron [1].

Fig. 2. Native copper crystals (size 0.2 to 0.9 mm); a) octahedron, b) cuboctahedron, c) cube [2].

Fig. 3. Tending to a tetrahedron dichotomous shape of small satellites: a) Thebe (PIA02531), b) Hyperion (08904), c) Telesto (07546).

Fig. 4. Dichotomy of larger satellites: a) Europa (00502), b) Iapetus (06169).

Fig. 5. Tendency to octahedron shapes: a) Yanus (06613), b) Phoebe (06066), c) Helene (08269), d) Amalthea (01074), e) Phobos (04589), f) Prometheus (07549).

Fig. 6. Tendency to cubic (a, b) and more complex (c) shapes; a) Epimetheus (07531), b) Helene (07547), c) Pandora (07530).

Fig. 7. Structural hexagons: a) Saturn (North pole, portion of 09188), b) Mimas (Hershel, portion of 06259), c-d) Rhea (portions of 07566).

References:
MORPHOLOGICAL SURFACE FEATURES IN THE WEST PROMETHEI TERRA REGION, MARS.

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Introduction: The studied W Promethei Terra region (36-50°S, 90-106°E) is roughly ~700 km across. It occupies a unique area on the smoothened E Hellas basin rim, and consists of two parts: a gentler (~0.07°, eastward of ~97°E) and a steeper (~0.88°, W of ~97°E) regional slope.

To the NE, E, and S the studied region is surrounded by Noachian cratered terrains, and the central area is cut by the large canyons of Harmakhis, Reull and Teviot Valles. The western and central areas exhibit smooth Hesperian age plains [1-10] that display a set of features that do not occur elsewhere on the eastern side of the Hellas basin (channels, mesas, several types of ridges, etc).

Here we describe the surface features identified within the smooth plains. We analyzed them using the data from HRSC camera [11, 12] in conjunction with THEMIS [13] and MOC NA images [14]. Where applicable, HRSC and MOLA digital terrain models and single MOLA tracks [15] were used to obtain topographic information. This research continues our study of the fluvial history and evolution of the Hellas basin and its surroundings [16-22].

The smooth plains have been interpreted to be due to a wide range of possible origins, but no thorough investigation of the region has taken place yet. We find that the plains exhibit a variety of features, some genetically related to the basement material and others due to deposition/modification of younger materials. The walls of the terrain cutting canyons reveal the subsurface to be layered (layer mean thickness ~40-50 m, total layer stack at least 1-1.5 km). This is explained either by sedimentary strata or by volcanic lava sheets. The layer roughness, their similarity to volcanic province layering (e.g. Lunae Planum, Syrtis Major) and the presence of wrinkle ridges on the surface of the plains collectively favors volcanic interpretation. The age of the plains appears to vary; the ones to the N of Reull Vallis are of Hesperian age [23], while the area S of Reull corresponds to the transition from Hesperian to lower Amazonian [16,17,23].

Features of the smooth plains: Mesas: A set of ~30 mesa-like hills is scattered throughout the plains on the S side of Reull Vallis [7,20]. They occur predominantly just at the break in the regional slope or to the E of it. The mesas’ relationship with the plains material has been controversial [compare 7,8]. Their morphological similarity and their distribution suggests that they may represent isolated fragments of a previously more widespread and continuous deposit [20].

The mesas are usually several to 10s of km across and about 100-300 m high. They have flat tops, but at high resolution they exhibit smooth, hummocky or slightly eroded summit...
terraces on their tops, which may indicate a rough 50-75 m layers on the mesa bodies. However, no layers have been observed on the mesa walls.

Typically, the mesa edges are very sinuous to scalloped. Some mesa walls are characterized by debris aprons (extending up to a km away). The aprons are smooth and featureless at scales of 10s of m, but at high resolution they are revealed to be lined out or heavily etched. The aprons either occur preferentially or tend to be larger on S-facing walls. The scalloped edges with the debris aprons suggest that the mesas consist of low-strength material(s). The most developed capes and concave intrusions into the mesa structure are often occupied by channels originating from gullies and broad shallow depressions on the terraces inside the mesas. The smaller channels join together within the mesa to form the larger channels that extend outside the mesas and cut the surface of the plains. In places, similar channels are observed in between two neighboring mesas, being fed by gullies on the mesa walls. These relationships suggest that the channels are fed by water run-off from the mesa body. The mesa-related channels tend to disappear further downstream, and no clear connection between them and other fluvial channels in the plains can be observed.

**Sinuous channels:** The plains are cut by E-W trending narrow (100s m to a few km wide) and sinuous channels. They are usually inserted within V- or U-shaped valleys several km wide and <150 m deep. They tend to occur mostly within the W enhanced topographic gradient area. The channels have either none or only a few tributaries and do not form dendritic patterns of integrated fluvial features occurring on the cratered highlands around Hesperia Planum [24,25]. Several 10s individual channels are found in the region, none having obvious sources. In places where the channels become broader, the high resolution images often reveal a braided pattern.

**Wrinkle ridge** -type ridges occur throughout the smooth portion of Promethei Terra. They are morphologically similar to the ridges characterizing vast lava plains on Mars (e.g. Hesperia and Lunae Planum, e.g. [26]). See [22] for further discussion. In places the mesas superpose the wrinkle ridges, thus being younger.

**Long straight narrow ridges** (widths < km, heights 10s m, lengths 10s km) are seen on the surface of the plains. They occur in preferentially NE-SW-oriented groups. The regional topography does not appear to control the distribution of the ridges. Their morphologic characteristics, areal distribution, and close association with the lava plains are consistent with and suggest that the straight ridges may represent exhumed dikes [27], which have served as feeders for the lava plains.

**Summary:** The found features suggest that the region is of volcanic origin. The lava sheets have experienced post-emplacement compression and dike injections. For further discussion on the endogenic activity and the early evolution of the region, see the associated [22].


**Eolian deposits:** In a number of places, accumulations of materials (100s m wide, up to several km long) are seen at the base of N/NE facing walls of mesas, impact craters and other topographic rises. Some of them are partially layered (see e.g. MOC NA S0702871 and R1702372). Initially [20] they were interpreted as fallout blocks from the mesas, as the materials are most prominent along the mesa walls, but their association with different types of topographical features suggests that they are in fact regional depositional features, probably accumulations of wind-blown materials.
We describe the ongoing mapping process, categorize the bens (Fig. 2) on the volcano itself as well as in its vicinity. So far, we have observed and documented a large number of straight /curvilinear ridges, fractures, troughs and grabens (Fig. 2) on the volcano itself as well as in its vicinity. We describe the ongoing mapping process, categorize the found features and propose scenarios for their formation.

2. Study rationale: Radial / concentric giant dike swarms have been shown to extend for hundreds of kilometers from their sources on other large volcanic provinces on Mars, as well as on Earth and Venus [9]. Dike systems have previously been identified on the N side of the Hellas basin [10,11]; see A in Fig. 1. The origin and distribution of the dikes in the greater Hellas region is of importance when discussing the regional geology, and especially when attempting to synthesize a chronology of and causal relationships between events. As the dike patterns and sizes are related to the whereabouts and characteristics of the feeding magma bodies, the distribution of the dikes around HP gives a hint about the heat flux in the region. This in turn reflects on the possible formation scenarios of e.g. the outflow channels just S of the volcano and the formation of several floor-fractured craters in the NE Hellas region [12]. Additionally, mapping out the dike patterns around HP helps to determine whether the volcano is the only magmatic center in the region, or if other previously unknown sources exist.

3. Volcanic environment:
   3.1 Hadriaca Patera is a low-relief volcano (slopes angles 0.8°, height 1.2 km, diameter >300 km) [13]. It has a 700 m deep caldera with a diameter of 90 km. Its flanks extend ~120-170 km from the caldera center except in the SW, where they reach ~350 km towards Hellas.

   The HP flanks are characterized by ~100-200 m deep and ~5 km wide downslope-running troughs. They appear to have a complex evolution of both fluvial and lava origins [14]. Morphological analyses of HP indicate that fluvial processes were the dominant influence in the initiation and subsequent development of the many dissecting valleys. However, over 2/3 of the valleys display lava-associated features, indicating that lava and possibly volcanic density flows were important as the valley-forming processes. Fluvial erosion appears to have been widespread and dominated the region surrounding HP [15].

   Models derived from studies of terrestrial pyroclastic flows have been applied to HP deposits, showing that their characteristics are consistent with an origin by the emplacement of gravity-driven ash flows generated by hydromagmatic or magmatic explosive eruptions [16-20]. Due to the fact that “the Martian paterae and terrestrial ignimbrite shields both display significant, layered pyroclastic deposits with low-angle slopes, shallow summit depressions, no resurgence, and post-explosive effusive lava flows”, Byrnes [20] interpreted Hadriaca to be of ignimbritic composition. On the other hand, HP may be above tension zones resulting from the formation and evolution of Hellas basin, and, thus, may in fact be directly analogous to terrestrial mafic explosive volcanoes [21].

   3.2 Hesperia Planum is a saddle-shaped province of volcanic lava plains, characterized by its prominent wrinkle ridge population. It is connected to Hellas by a 300-km wide smooth valley, thought to be indicative of and carved by ancient huge volatile outbursts from Hesperia, caused by heat transport from volcanic activity [22]. Later, similar but smaller, outbursts have formed the channels Dao and Niger Vallis immediately south of HP [e.g. 1,5,6]. These originate from obvious, several km deep chaotic regions.

4. Methods, used data: HRSC images were chosen for this study due to their high resolution and large areal coverage. To pick out the dike-related features, we use the desktop HRSCview [23] to obtain their properties.

5. Preliminary findings: Four features categories are recognized (cf. Fig. 2). Each of the four feature types is found transforming into others. However, the most common transition is from a linear ridge to a linear fracture.

   5.1 Linear ridges are generally tens to hundreds of km wide, have reliefs of only few m, but may extend for tens of km. They are clearly distinct from the wrinkle ridges found widely throughout the study region. Linear ridges are proposed to be dikes exhumed by later erosion processes, and thus expected to occur especially in local lows.

   5.1.1) A tightly packed group of N-S orientated short (<2 km) ridges occur in a location S of Harmakhis Valles (at 40°S in Fig. 3). Their shape starts from a rounded southern end, and narrow up towards the tail-like N ends. This resembles typical drumlin fields found on Earth. Drumlins are glacier formations, in which the ice either deposits carried sediments next to an obstacle such as sturdy protrusions in the bedrock, or reshapes the rocks into streamlines or "rock drumlins". In the Harmakhis case, the “sturdy/hard rocks” may in fact be pre-existing dikes. Additionally, several non-drumlin-like long linear ridges are visible in the same locale, indicating the possible extent of the glacier, or post-glacial dike formation.
5.2) **Linear fractures** are almost identical to the ridges but with inverted vertical topography. In places such as in the chaotic regions seen at 36.7°S and 34°S in Fig. 3 they cause the surface to be cut and fractured intensely. The fractures are in thus often manifestations of subsurface dikes. They occur all over the region, but concentrate near the outflow channels.

![Figure 2](image.png)

**Figure 2.** Examples of the mapped features found on the HP caldera rim: linear ridges (1), linear fractures (2), grabens (3) and pit chains (4), all indicative of dikes. Note how they transform from one type to the next. North is up, image from HRSC-nd 528.

5.3) **Grabens** are depressed land blocks bordered by parallel faults. They tend to be hundreds of m to km across and occur near smaller fractures. Shallow grabens may be purely tectonic extensional features, but occur also over and are suggestive of deeper dike injections [9].

5.4) **Pits and pit chains** are usually elliptical depressions, comparable in width to grabens. They occur in places where the collapse has not been complete; often in marginal areas close to fracture- or graben-forming regions.

6. **Distribution and alignment:** The preliminary results of our study show that the identified features are concentrated on the HP volcano and in various isolated locales throughout the study region.

6.1) Most of the dikes on HP are concentric to the summit and the caldera. Here the transitions between ridges and troughs are very common, indicating a certain dike. Also, the abundance of ridges is higher on the volcano summit and flanks; elsewhere depressional features prevail.

6.2) The areas in and immediately around the Valles exhibit intense fracturing, grabens and pit chains. These areas are found to be two-fold: a) the tops of the steep canyon walls have fracture, most probably due to gravity and stresses, eventually causing mass wasting, while b) the canyon floors tend to go through more intense fracturing into chaotic terrains. The latter may be caused by collapses and material volume diminution due to volatile removal from the materials. However, in places the fracture patterns resemble those of the floor-fractured craters [11,12]; see also crater at 31.5°S, 94.3°E. These have been shown to be caused by dikes propagating through pre-existing fracture patterns in the subsurface [24], which are abundant on Mars [25].

**Summary:** We have identified several, possibly dike-related features. Others are radial / concentric to HP, while others appear not to be controlled directly by the volcano or related obvious magma chambers. However, almost all of the latter may be explained through cracking, mass movements or other tectonic phenomena, especially near the steep cliffs of the large channels found S of HP.

![Figure 3](image.png)

**Figure 3.** Distribution of the identified features south of Hadriaca (HP). Ridges (blue) tend to occur as concentric features at the summit and at the flanks, but form radial and irregular patterns further away. Fractures and grabens (red) occur mostly near the outflow channel walls and chaotic terrains. Impact craters (yellow) are marked for reference; background image is Viking MDIM2. D = Dao Vallis, N = Nigri V., H = Harmakhis V.


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WESTERN PROMETHEI TERRA SMOOTH PLAINS REGION, MARS: A VOLCANIC PROVINCE?

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Introduction: An area of smooth plains is located within the western Promethei Terra, around the major channels of Harmakhis and Reull Valles (Fig. 1). Massifs of ancient cratered terrains are rare within this region [1-9]. Different authors have mapped the area of Promethei Terra based mostly on the Viking data [5,7-11].

The smooth plains which occupy the low areas between the massifs: The plains occur on both sides of Reull and Harmakhis Valles and are more abundant in the area between them. The suite of materials that make up the surface of smooth plains is usually interpreted as sedimentary deposits with some contribution of volcanic materials modified by fluvial, eolian, and perhaps periglacial processes [5, 7-11]. With the high-resolution data very important features are observed on the surface of the plains and in their cross-sections provided by the deep canyons. The multi-layered nature of the plains is seen on the walls of the canyons that cut them. The average visible thickness of the layers at MOC resolution is ~70-80 m and the typical measured slope of the walls of the canyons is ~25-30°. This gives an estimate of the thickness of the layers, which is ~35-45 m. The thick stacks of sub-horizontal layers are seen in other regions of Mars where the interiors of lava plains are exposed. For example, the layered structure of Lunae Planum is obvious in places where Kasei Vallis cuts the plateau. The same structure characterizes volcanic plains of Syrtis Major. Both Lunae Planum and Syrtis Major are classic volcanic provinces the layered structure of which was formed by successive emplacement of sheet lava flows that followed the general topographic trend.

Besides the observed layering, in many places narrow wrinkle ridges (WR) deform the surface of the plains. WR mostly occur in the eastern portion of the area near Reull and Teviot Valles but some of them are seen near Harmakhis Vallis in the west. WR appear to avoid the area of the steeper topographic gradient. In Promethei Terra, the ridges are less prominent than in the other provinces and are poorly detectable at the Viking MDIM resolution. At the higher resolution provided, for example, by the HRSC images (~25 m/px) the ridges are obvious and it is seen that many of the ridges are covered by younger materials.

The layered structure of the smoother plains in Promethei Terra to the south of Reull Vallis and the presence of wrinkle ridges on the surface strongly suggest that the basement material of the plains represents a thick stack of lava flows.

Promethei Terra volcanic province: The area of smooth plains in the central-western portion of Promethei Terra is ~265,000 km². It is characterized by a variety of structures suggesting the fluvial, wind, and glacial modification [13] of the basement and appear to be formed during Hesperian time. The basement material must predate the surficial deposits and it clearly postdates the Noachian cratered terrains. Thus, the time of emplacement of the basement material may correspond to the Late Noachian-Early Hesperian epochs and, thus, is close to the time of massive volcanism on the rim of the Hellas basin.

The layers exposed on the walls of Reull and Harmakhis Valles are seen almost from the upper edges down to the floors of the canyons. The layered structure in the head depression of Harmakhis Vallis can be traced to the depth of about 1.6-1.7 km over the horizontal distance of about 9 km. Assuming that the layers are parallel to the surface, which slopes to the NW at about 0.9°, the true thickness of the stack of layers is estimated to be about 1.3 km. This number ap-
pears to be close to the upper value of the thickness because it is the larger depth to which the layered structure of the smoother plains in Promethei Terra is exposed. The approximate volume of the layered material in this region is estimated to be \(-0.3 \times 10^6 \text{ km}^3\), which is close to the volume of the volcanic fill in Hesperia Planum, \(-0.5 \times 10^6 \text{ km}^3\) [14].

This interpretation suggests that the western-central portion of Promethei Terra represents a sizable volcanic province among the others (e.g. Hesperia and Malea) on the rim of Hellas basin. The steady regional slope and the absence of topographic barriers along its western side mean that the lavas erupted in this region must flow into the Hellas basin and partly fill it. The obvious difference of the proposed volcanic region from the others known on the rim the Hellas basin is the absence of distinct volcanic sources such as Hadriaca and Tyrrhena Paterae in Hesperia Planum and Malea Patera in Malea Planum. This suggests that the sources of lava were more distributed and the volcanic activity there may have a sporadic character. There is also a possibility that the volcanic material in the basement of the smoother plains may have an external source. The plains in Promethei Terra are separated from Hesperia Planum by a contiguous and high-standing area of Noachian terrains and are hypsometrically higher than plains of possibly volcanic origin to the west and north of Harmakhis Vallis. Thus, the surface lava flows from Hesperia Planum and its southwestern extension could not be the source for the lava filling in the central-western Promethei Terra. However, dikes propagating from the volcanic centers in Hesperia Planum could have potentially served as the feeders of lava. But, the area of the smoother plains in Promethei Terra is topographically higher than Hadriaca Patera, which makes that patera to be an unlikely source of lavas. However, the topographic position of Tyrrhena Patera allows this volcano to be a potential source of volcanic material in Promethei Terra.

The thickness of the lava plains in Promethei Terra appears to be large (roughly a kilometer) and the surface of the smoother plains is clearly lower (by about 0.5 km) than the base of the surrounding cratered terrains. Thus, emplacement of such a volume of volcanic material creates the problem of space. The room for the emplacement of lavas may have been provided if a broad depression existed on the rim of Hellas in the region of Promethei Terra before formation of the lava plateau there. A depression of this size is likely to be the one with shallow dipping edges. There are observations, however, that are poorly consistent with the presence of such a depression. (1) Isolated massifs of the cratered terrain amidst the smoother plains are sometimes as high as topographic peaks within contiguous areas of the cratered terrain to the north and south of the plains. This may indicate that the original surface of the Noachian cratered terrains was roughly at the same mean level before emplacement of lavas. (2) Relatively narrow zones of enhanced topographic gradients bound the area of the smoother plains. These regional scarp-like features suggest that the area occupied by the plains has a U-shaped topographic profile, which is inconsis-

Conclusions: The smooth plains basement unit in the central-western portion of Promethei Terra appears to be volcanic in origin. The approximate volume of the layered material in this region is estimated to be \(-0.3 \times 10^6 \text{ km}^3\) and the time of emplacement of the material may correspond to the Early Noachian-Early Hesperian epochs. As favorable modes of placement for the volcanic material from remote sources do not exist for this unit, it is possible that it may be an independent volcanic province.

Observations performed in 1998—1999 [1] with the upgraded Arecibo radio telescope corroborated the existence of anomalous regions with unusual properties on Mercury. [1]. All these areas are crescent-shaped and were identified with impact craters located in the polar areas of Mercury. It was suggested that areas with unusual properties are clusters of volatile elements in the "cold traps" of the polar regions of the planet. Accordingly [3] water-ice deposits remain stable for a long time if the maximum temperature does not exceed 110 K. The temperature limit increases to 130-150 K in the presence of regolith [6]. The maximum temperatures in craters R, T, P2, S, and Q are exceed 170 K even in shadowed areas. We estimate the temperature regime in such craters.

Moses et al. (1999) [2] estimate that external sources of ice on the Mercury can supply ice deposits 0.05 – 0.6 m thick. For such crater as crater with features S we assigned thickness of ice deposit as 0.05 m. The thickness of regolith cover is probably between 0.1 and 0.5 m [3].

For an estimation of subsurface temperature using a two-layer model from [4]. Model consist from two layer: top layer by thickness 2 cm, with density 1300 kg/m³, and lower with density 1800 kg/m³. The dependence of thermal conductivity on the temperature is given by as k = k0*[1+\chi*(T/350)] [4], where k0 is solid conductivity, T is temperature, \chi is the ratio of radiative to solid conductivity at a temperature of 350 K. Values for k0 and \chi for both layers are taken from [4]. The temperature of surfaces determined by direct solar flux (for permanently shade area it is equal 0), the internal heat flux, emitted infrared flux and the secondary reflected light flux from illuminated surface of the crater. We ignore the effect of the solar wind, because it barely penetrates into permanently shadowed areas. Fig.1 shows variations of subsurface temperature for point, lies in permanently shaded region of crater, identified with feature S. This crater (79,1° N, 16,3° W) has diameter 21 km. We adopted morphometric data for Mercurial craters from [5]. 18 % of crater interior is permanently shaded. Point lies in the south part of crater at the distance of 5 km from center.

The maximum diurnal temperature in this point exceed 230 K. We estimate the temperature variations in this point for solid water ice deposit, dry regolith and water ice covered by a 10 and 50 cm-thick layer of regolith. Diurnal variations of subsurface temperature reaches 20 cm. The minimal temperature for water ice is 180 K and significantly above the limit for long-term stability for water ice. The temperatures for dry regolith and water ice covered by a 10 and 50 cm-thick layer of regolith increases to 150 K at the depth 4 cm and reaches its minimal significance 138 K at the depth 7 cm. This temperature is too high to allow any volatiles except sulfuric compounds to persist in these craters for a long time.


The dependence subsurface temperature from depth for feature S: 1 – dry regolith; 2 – water ice without regolith cover; 3 - water ice under 10 cm regolith cover; 4 - water ice under 50 cm regolith cover.
**ASSESSMENT OF "WET" MECHANISM OF SLOPE STREAKS FORMATION ON MARS.**

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**Introduction:** Dark and bright streaks without any apparent topographic expression are forming on steep slopes on Mars [1 - 5]. They occur in wide regions in the equatorial zone; these regions are characterized by high albedo, low thermal inertia and fine-dust coverage. Formation of dark slope streaks is an ongoing process; new streaks were revealed by repeating imaging. The slope streaks have been mostly interpreted as results of some kind of dry mass wasting of fine dust [e.g., 2, 5]. It has been also suggested that the streaks are results of some "wet" processes, namely brine-rich debris flows [1]; in [6] it has been suggested that some streaks result from springs formed by salty groundwater discharge. There are numerous examples of slope streaks on isolated peaks and mesas, which excludes groundwater origin, while morphologic similarity implies the same origin of all slope streaks. In [3] it has been shown that some features of slope streak distribution suggest some role of H₂O phase transition in formation of the streaks. Recently, striking morphological similarity of the slope streaks on Mars and in Antarctic Dry Valleys has been demonstrated [7,8]. Antarctic slope streaks are formed by percolation of water and dilute brines above the ice table and wicking to the surface. The source of water is seasonal transient melting of snow. Because no snowpacks have been detected in the slope streak regions on Mars, and subsurface flow of water and dilute brines is not consistent with observed temperature regime, the complete Antarctic analogy cannot be immediately applied to Mars. However, the striking morphological similarity impels us to consider the possibility of "wet" scenarios [7,8]. Below we outline a "wet" mechanism of slope streak formation on Mars, which is marginally possible under the existing physical conditions and given known observational constraints [9]. Then we consider observations that can help distinguish "wet" and "dry" mechanisms.

"Wet" mechanism (Fig. 1). Shallow subsurface contains frozen soil with some ice or highly hydrated mineral phases. Under present conditions ice or highly hydrated phases are not securely stable against water vapor diffusion; they were emplaced the soil under previous wetter climate conditions (e.g., recent obliquity maxima) and currently undergo slow desiccation. This H₂O-rich soil is overlaid by centimeters- to decimeters-thick layer highly enriched in chlorides, first of all, CaCl₂ and FeCl₃. On top of this layer, there is a few mm to few cm-thick layer of fine dust, responsible for the observed distinctive optical and thermal properties. This layer plays a role of a barrier for diffusion of water vapor and provides high vapor concentration in pores of the salty layer, which leads to formation of small droplets of highly concentrated brines, especially during warmer seasons. Sometimes at some places, when amount of liquid phase is large enough, the droplets coalesce; the liquid percolates downhill within the salty layer on top of impermeable ice-filled soil, gathers further portions of liquid phase spread in the pore space and produces a run-away process of subsurface flow. This high-concentration brine wicks up through he uppermost dusty layer and dries up quickly, as soon as reaches the surface (or freezes, if this occurs at night, and sublimes next morning). This process alters the uppermost dust layer structure (e.g., by cementing dust with hydrated salts like antarcticite, CaCl₂·6H₂O, or in any other way), which affects photometric properties and makes the subsurface flow observable as a dark slope streak. Further gradual changes of this new surface structure (for example, dehydration of antarcticite) lead to slow brightening of the streak, in some cases, formation of a bright streak, and final fading away.

Main advantage of the "wet" mechanism. The main shortcoming of "dry" scenario is its difficulty to explain branching, anastomosing pattern of the streaks together with the absence of any observable accumulation of material at the distal streak ends, especially, taking into account that slope streaks sometimes occur at rather gentle slopes [10]. Among numerous types of dry flows known in terrestrial geology, technology and laboratory, there are no dry flows, that would have these characteristics simultaneously (see further discussion in [10]). Percolation of liquid in shallow subsurface does reproduce these characteristics simultaneously, as proven by Antarctic observations [7,8].

Main difficulty of the "wet" mechanism. The mechanism outlined above assumes a thin shallow layer highly enriched in Cl (a factor of 5 and more in comparison to typical martian soils). At current level of knowledge, there is no any consistent explanation, how such a layer can form.

Possible tests for "wet" mechanism. Below we list tests of "wet" mechanism that can be carried out by analysis of existing data sets or data currently being obtained by working instruments onboard Mars-orbiting spacecrafts.

*Spectral signature of hydrated chlorides* can potentially be registered by CRISM instrument onboard Mars Reconnaissance Orbiter (MRO). If slope streaks contain any trace of hydration bands in the NIR spectra, this would make "wet" hypothesis much more probable, but will not prove it, because...
metastable hydrated salts can also be exposed by dry avalanches.

Gentle slope of slope streaks, especially in their uppermost parts (<20°) would pose very serious difficulties to the dry avalanche scenario, hence, would strongly favor "wet" mechanism. In [10] gentle general slope of a few streaks was demonstrated with MOLA profiles; stereo imaging by HRSC (Mars Express), CTX (MRO) and especially HiRISE (MRO) cameras gives a possibility to analyze slopes in detail.

Observation of slope streak formation during cold season can be critical. Despite generally low-latitude location of slope-streak regions, some of them have significant seasonal variations of the day-average surface temperature, especially on steep slopes of certain orientation. Analysis in [3] showed that observations of newly formed streaks available by that time do not contradict streak formation during warm season. If new observations demonstrate that streaks are formed when shallow (~1-3 dm, within seasonal thermal skin layer) subsurface temperature is below ~190 K, this would reliably reject "wet" mechanism outlined above. Suitable imaging observations have been carried out by MOC (Mars Global Surveyor, MGS) and have been carrying out by HRSC, THEMIS VIS (Mars Odyssey), CTX and HiRISE. Subsurface temperatures can be reliably derived from thermal emission observations (TES (MGS) and THEMIS IR) through careful and cautious thermal modeling.

Observation of a slope streak in the process of formation would almost prove "wet" mechanism, at least, almost reject dry avalanche scenario. There are many observations of slope streaks absent in earlier and present in later images, but there is no any observation of a streak that would be present in earlier image and longer in later image. Such observation with high probability would mean somewhat long duration of formation process, which is perfectly consistent with "wet" mechanism and almost unexplainable with "dry" avalanche scenario.

Absence of a shallow high-thermal-inertia layer would be inconsistent with "wet" scenario. The very surface layer (within diurnal thermal skin depth) in the slope streak regions has low thermal inertia. Long-term observations of night-time surface temperatures (TES and THEMIS) in the regions with significant seasonal temperature changes can give information about thermal inertia of shallow subsurface within seasonal thermal skin through careful and cautious thermal modeling [11]. Soil structure assumed by the "wet" mechanism outlined above predicts increased effective thermal inertia beneath the diurnal thermal skin due to thermal effect of seasonal phase transitions. Under current level of knowledge, it is not clear, whether TES and THEMIS data are accurate enough for such a test.

Geomorphologic observations supporting subsurface ice in the slope streak regions would favor the "wet" mechanism. Rich geomorphology specific to slope streak regions is seen in HiRISE images and waits for detailed analysis and interpretation.

OYSTER-SHELL CRATERS IN MAMERS VALLIS, NORTH ARABIA TERRA, MARS: DEFINITIONS AND IMPLICATIONS. A. Kress and J. W. Head
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Introduction:

The term “oyster-shell crater” was so coined because the morphology of these craters is similar to the scalloped shape of an oyster’s shell [1]. The morphology of oyster-shell craters coupled with their apparent distribution only on lineated valley fill (LVF) and lobate debris aprons (LDA) led to the hypothesis that the ice content of LVF and LDA was instrumental in the formation of oyster-shell craters [1-4]. LVF and LDA form in connection with fretted terrain and channels in northern Arabia Terra [5-7]. In this study we analyzed craters on the LVF and LDA filling Mamers Vallis (Fig. 1).

Fig. 1. Deuteronilus Mensae in northern Arabia Terra. Mamers Vallis to lower left.

Few oyster-shell craters exhibit evidence of deformation by flow, thus they must have been emplaced after LVF and LDA flow stopped but while there was still significant ice in the substrate. Oyster-shell craters may therefore potentially be diagnostic of certain climate conditions on Mars, namely, the transition from facilitation of ice flow to favoring ice sublimation and deflation as the dominant surface modification processes.

Uncertainties regarding oyster-shell craters include: their precise significance in terms of ice content of the substrate; their mode(s) of origin; the significance of the range of morphologies of oyster-shell craters; how these different morphologies are related; and how oyster-shell craters are affected by degradation.

Classification of craters in Mamers Vallis:

Mangold [1] identified five crater morphologies occurring on LDA and LVF: fresh, beginning to be degraded, strongly degraded, oyster-shell crater, and ghost crater (Fig. 2, A-E).

Fresh craters are bowl-shaped depressions with mostly smooth interiors and distinct, continuous rims (Fig. 2, A).

Craters beginning to be degraded are also bowl-shaped depressions with smooth interiors, but while their rims are distinct, they are often discontinuous, softened or cut through by LVF or LDA lineations. Crater floors may be flat and some may have deposits, recognizable by scarps or albedo changes (Fig. 2, B).

Strongly degraded craters are depressions that are still noticeably circular, but their rims tend to be indistinct and often broken by pitting or lineations (Fig. 2, C).

Mangold defines oyster-shell craters as degraded craters whose interior “presents the form of an ‘oyster shell’” [1]. According to this definition, an oyster-shell crater comprises a flat-topped mesa with a scalloped edge surrounded by a moat that may or may not exhibit LVF/LDA texture and that rises up to meet the surrounding LVF or LDA (Fig. 2, D).

Lastly, ghost craters are more heavily degraded oyster-shell craters. Their shapes are only vaguely circular, but they still maintain the mesa-and-moat structure particular to oyster-shell craters (Fig. 2, E).

We identify four additional crater morphologies, all similar to “oyster-shell craters” as Mangold defines them [1]. They are 1) one moat surrounding a circular, unscalloped mesa, 2) a circumferential moat and mound around a central, circular depression, 3) two concentric moats, separated by a circular mound, surrounding a circular mesa, and, 4) an isolated, near-circular mesa (Fig. 2, F-J).

1) The craters with a single moat and circular mesa are composed of a circular, flat-topped mesa with a smooth unbroken edge. There may be a slight depression or pitting in the top. The moat may be smooth, it may contain dunes, or it may exhibit LVF/LDA texture, and it may have a distinct rim separating it from the LVF or LDA (Fig. 2, F).

2) The craters with a circumferential mound and moat have central depressions that are near-circular, shallow, and generally smooth inside, although some may exhibit LVF/LDA texture. The surrounding
mounds are generally pitted, rounded, and relatively wide in comparison to crater diameter (Fig. 2, G). Some, however, narrow into ridges and appear more like the rims of degraded (but not oyster-shell) craters (Fig. 2, H).

3) Craters with two concentric moats are rare, but they are very morphologically similar to craters with a single moat. The interior mesa can be flat-topped or slightly rounded, the circumferential mound and moats may vary in width and may be rounded or pitted as well (Fig. 2, I).

4) The isolated, near-circular mesa, or inverted crater, is generally circular in shape, usually flat-topped, although occasionally the top may be rounded, its surface may be pitted, or it may have a circular depression on top (Fig. 2, J). These mesas are very morphologically similar to the interior mesas of some oyster-shell craters; they may represent the end result of the erosion of an oyster-shell crater’s moat as well as the surrounding material.

**Modes of formation of oyster-shell craters:**

It will be important to determine how oyster-shell craters form and whether their morphologies are primary or secondary structures. Did oyster-shell craters have the same morphology as fresh craters just after their emplacement? This would imply that post-impact degradation processes are responsible for forming oyster-shell craters. It has been observed that crater emplacement can cause “armoring” of the impacted material [8]. Pedestal craters, for example, have more resistant ejecta blankets and thus stand above the surrounding terrain [8]. It may be possible in the case of oyster-shell craters that the impact into an ice-rich substrate armors the crater floor even as the ejecta may weaken the surrounding area and facilitate its becoming a moat.

On the other hand, if oyster-shell craters had similar multiple-ringed morphology not long after impact, then it is likely that the impact and settling processes are responsible for formation of oyster-shell craters. Acoustic fluidization may be one mechanism by which oyster-shell craters form, but it is generally invoked to explain much larger craters [9].

**Summary and Conclusions:**

Of the nine different types of craters identified on LDA and LVF in Mamers Vallis, six of these are morphologically nearly identical to or morphologically related to oyster-shell craters as defined by Mangold [1]. Due to their morphological similarities, it is likely that the same processes were responsible for forming all variations of oyster-shell craters. In addition, oyster-shell craters have been seen only on LVF or LDA, material once more ice-rich. It is likely that ice content was instrumental in the differential armoring and weathering of the oyster-shell craters. Further study of oyster-shell crater ages, distribution, and possible formation mechanisms may yield a better understanding.

**References:**


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**Fig. 2. Crater types on LVF/LDA in Mamers Vallis.**

- **A)** Fresh, **B)** beginning to be degraded, **C)** strongly degraded, **D)** “oyster shell” interior, **E)** ghost crater, **F)** circular mesa and one moat, **G)** central depression, surrounding moat and mound that narrows to rim, **H)** central depression with surrounding moat and wide, rounded mound, **I)** central mesa and two moats, and **J)** isolated, near-circular mesa. All images from MOC; MOC image number is listed below each.
Morphological parameters of the inside depressions of elliptical coronae on Venus.

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We have studied the interior ellipse depressions of Venusian coronae (1) as well as the morphological parameters of their outside boundaries (2, 3). Our earlier measurements have shown that, in some cases, the azimuths of long axes of the outside ellipses of coronae do not coincide with the azimuths of long axes measured from the inside ellipses. Moreover, the ratio of short to long axes of these two ellipses measured may differ. We found that the controversy between the measured outside and inside parameters of corona ellipse contours may provide an additional tool to compare corona distributions on Venus.

Fig. 1 shows the distribution of long axes of corona azimuths measured from the inside contours. It is similar to the distribution of the same parameters of the outside boundary (Fig. 1 in ref. 3). Coronae group to form similar areal clusters as seen if the distribution of their outside contour parameters are used. Only a small number of coronae (22 out of 219 measured coronae or 10%) has an essential difference in the long axes of their ellipse azimuths measured from the outside and inside ellipses. This is an insignificant fraction of coronae and does not change the overall result of their orientation. As a rule, these unmatched coronae are also strongly affected by additional geological processes.

In the observed areal clusters, there is an evident difference in the ratio of short axes to long axes between inside and outside contours of coronae. In some cases, there is a change from a low ratio to higher ratio while in other cases it is vice versa and the high ratio value changes to a lower value. There are also coronae in which the value of the ratio of short axes to long axes does not essentially vary between the two contours. In the areal corona clusters with an obvious spatial regularity the change in the ratio between the main axes is not observed. In general, the areal corona clusters have the contours of their inside ellipses similar than found from borders of the outside ellipse contours in these areally distributed corona clusters. This indicates that the inner and outer ellipses in each of these coronae are lineated along the same tectonic zones (Fig. 1 in ref. 3).

Comparison of the distributions of the measured corona parameters of leads to the important understanding of their distribution on the surface of Venus. During the formation of these coronae, there was a geological situation stable enough to allow the orientation of azimuths of the long axes of both the inside and outside contours of these ellipses to form as observed. The cases where the concurrences between the directions of the two ellipse axes are not observed may display the effects of additional geological processes that have to be taken into the account. Most probably, the rejuvenated corona activity has led to the change in the direction of the azimuth of the long axis of the inside ellipse contour.

The ratio between the two main axes of corona ellipses defines an amount of corona deformation away from the circle. The observed difference in the ratio values between outside and inside contours in any corona are most likely caused by geology during the formation and development of these rounded structures. As we noted above, the ratio between short axes to long axes of the inside corona contours can be above or below the parameters obtained from the outside contours, i.e. the difference in the trend and in the degree of deformation can be observed. There is a plan to study such change-indicating trends in the future.

Fig. 1 Distribution of the corona parameters measured from their inside depression (azimuths of long axes and ratio short/long axes).

Legend

<table>
<thead>
<tr>
<th>Directions of long axes of coronae</th>
<th>Ratio of short/long axes of coronae</th>
</tr>
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<tbody>
<tr>
<td>W - E (-75° - 90°, 75° - 90°)</td>
<td>0.9 - 1</td>
</tr>
<tr>
<td>-75° - -45° (-60°)</td>
<td>0.8 - 9</td>
</tr>
<tr>
<td>-45° - -15° (-30°)</td>
<td>0.7 - 0.8</td>
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<tr>
<td>S - N (&lt;15° - 15°)</td>
<td>0.6 - 0.7</td>
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<tr>
<td>45° - 75° (60°)</td>
<td>0.5 - 0.6</td>
</tr>
<tr>
<td>15° - 45° (30°)</td>
<td>0.4 - 0.5</td>
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<td>&gt;15°</td>
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LARGE-SCALE CRUSTAL EXTENSION AND VOLCANISM: AN EXAMPLE FROM LADA TERRA, VENUS, P. Senthil Kumar1,2 and James W. Head2, 1National Geophysical Research Institute, Hyderabad 500007, India, senthilngri@yahoo.com; 2Department of Geological Sciences, Brown University, Providence, RI 02912, USA, james_head@brown.edu.

Introduction: Large-scale crustal extension and volcanism are common to Earth and Venus. The East African rift system is one of the best examples of such belts on Earth. We show an example of a complex network of several thousand kilometers long and a few hundred kilometers wide extensional belts from Venus, which are associated with volcanism. Geological mapping of the V-56 quadrangle (Fig. 1) reveals various tectonic and volcanic processes in Lada Terra that consist of tesserae, regional extensional belts, coronae, volcanic plains and impact craters. This study aims to map the spatial distribution of different material units, deformational features or lineament patterns and impact crater materials. In addition, we also establish the relative age relationships (e.g., overlapping or cross-cutting relationship) between them, in order to reconstruct the geologic history. Basically, this quadrangle addresses how coronae evolved in association with regional extensional belts, in addition to evolution of tesserae, regional plains and impact craters, which are also significant geological units of Lada Terra.

Geologic mapping: We used 250-m-per-pixel Magellan SAR image to prepare a geologic map at a scale of 1:5,000,000. Wherever necessary, full-resolution (75-m-per-pixel) images are used for fine details. This quadrangle is bordered by Kaiwan Fluctus (V-44) [1] and Agnesi (V-45) [2] quadrangles in the north; Mylitta Fluctus (V-61) [3,4], Fredegeonde (V-57) [5] and Hurston (V-62) [2] quadrangles in the west, east, and south, respectively. From the geologic mapping, we report on the distribution of the following material and structural units, and reconstruct the geologic history.

Material and structural units: The oldest known material units are tesserae. They are radar bright areas characterized by multiple orientations of lineaments; two sets are dominant: NNW-SSE and ESE-WNW oriented lineaments. Tightly spaced ridges and troughs generally characterize tessera. The third dominant lineaments are NNE-SSW and NNW-SSW oriented a long rift zones, namely, Chang Xi Chasmata and Sco-Ne Chasma; but these are apparently restricted to the Cocomama tessera. In the northeastern part of the quadrangle, terrains (TLT, Fig. 1) similar to tessera are found. They have NNE-SSW to NE-SW oriented ridges, which are cut by ESE-SNW to NW-SE oriented troughs. The spacing of these structures is greater than the structures of the tessera. The tessera units contain intra-tessera basins, which are filled with lava flows of different ages; most of them are derived from the units outside the tessera, and a few are from intra-tessera volcanism.

Regional plains units embay the tessera terrain; they have wrinkle ridges and a few young fracture systems. The oldest known plains (but younger than tessera) are lineated (LP, Fig. 1), and are closely associated with shield plains (unit 5, Fig. 1) and the tessera. The LP is characterized by tightly spaced, NW-SSW oriented fractures, which are also common in the tessera. Two types of shield plains are present: a few occur in the pre-regional plains areas, while others occur in the core of coronae and adjoining areas.

The older regional plains (unit 4, Fig. 1) are cut by two regional extensional belts [6]: (1) NNW-SSE trending, 6000-km long and 50-200 km wide, Alpha-Lada (AL) belt, and (2) NNE-SSW trending, 2000 km long and 300 km wide, Derceto-Quetzalpetlatl (DQ) belt. These two belts are composed of fractures, rift basins and strike-slip zones. The DQ belt is punctured by Sarpanitum, Ethinoha and Quetzalpetatl coronae, while the Otygen, Demvamvit and Okhim-Tengri coronae occur along the AL belt. A few coronae have a circular central dome and outer concentric depression; they are defined by fractures, rift basins and ridge belts. Asymmetric and multiple coronae also occur in the southern part of AL belt. Two other extensional belts branch from the AL belt. Corona (Dymenyuuo, Toyo-uke, Loo-wit and Kshumay Mons) puncture these extensional belts. At many places, corona structures cut across the regional extensional belts, while at other places, the extensional belts cross the corona structures. There is a clear overlapping time relationship, as the one affects the formation of the other. Corona volcanism and tectonics are also closely related to one another. Lava flows erupt along the coronae fractures, for example, in the Ethinoha corona. Lava flows emanating from coronae travel several hundred kilometers. Volcanism is also related to shield volcanoes in many places.

The DQ and AL belts separate the plains units of Lavinia Planitia, Aibarchin Planitia and Mugazo Planitia, where lava flows are abundant; principally there are four plain units, of which the oldest one appears to be common to all the planitia units. Most of the younger units are locally derived from the coronae. The regional plains occurring to the east of Otygen corona have undergone intense fracturing and emplacement of dykes after post corona-extensional belt deformation. These fractures occur in two
directions: ENE-WSW and NW-SE. It appears that they represent the latest deformation, and could probably be related the terrain uplift, as is evident in many terrestrial examples. Impact craters are the youngest geologic units, except for one that is affected by the extensional belt deformation and the other embayed by regional plains. Most impact craters show complex geometry and a few have a bowl-shape. Many complex craters show run-out flows characteristic of oblique impacts. Further detailed mapping is underway to reconstruct the complete geologic history of this quadrangle in near future.

Introduction: Structural analysis of impact craters provides insight into the shock deformation of target rocks associated with meteorite impacts. Particularly, extremely well-preserved bowl-shaped simple craters, such as Meteor Crater (southern Colorado plateau, USA) and Lonar Crater (Deccan basalt province, India) are excellent examples for impact crater formation in sedimentary rocks and basalts, respectively, which are common target rocks on the surface of Mars (Fig. 1). We have undertaken structural geological mapping of these craters and the surrounding regions to understand the nature of the impact structures. In addition, we have assessed the role of pre-existing weakness zones (e.g., fractures), which are likely to influence the final crater morphology. For example, the squarish shape of the Meteor Crater is thought to have been influenced by the network of pre-existing fracture systems [1]. In contrast, Lonar Crater is circular and the pre-impact fractures are less abundant in the target rocks. The results of the structural analysis of these craters are as follows (Fig. 2).

Fig. 1. An overhead view of (a) Meteor Crater and (b) Lonar Crater, thanks to Google Earth. Note that Meteor Crater has a squarish outline, while Lonar crater is circular. Rim-to-rim diameter is ~1.2 km and ~1.8 km for (a) and (b), respectively.

Meteor Crater: The ~50000 year old Meteor Crater is a bowl-shaped simple crater with ~ 1.2 km diameter and ~ 180 m depth; it was excavated on the > 1 km thick, flat-lying Paleozoic-Mesozoic sedimentary rocks of southern Colorado in north-central Arizona (see summary in [2]). Structural mapping of the Moenkopi and Kaibab Formation outside the crater suggests that, before the impact event, the target sedimentary rocks were characterized by horizontal bedding planes, which are cut by at least three prominent sets of fracture systems, most of which are sub-parallel to the crater long-walls. Similarly, structural mapping of the Moenkopi and Kaibab Formations exposed on the crater wall reveal that the rim is dissected by radial tear faults, whose kinematics appears to have controlled the geometry of the impact deformational features. The faults occurring in the crater diagonals are prominent ones, causing greater vertical displacement. The tear faults define the crater rim into twelve tectonic blocks, where the bed rock shows upward turning of sedimentary sequences, with a distinct rim uplift. In particular, the southern rim suffered minimum uplift. The crater rim is also cut by three distinct groups of fractures: radial, concentric and conical fractures, which have some preferred orientations. The radial fractures have more or less steep dips and strikes that are approximately perpendicular to the crater rim, although some of them are oriented obliquely. The radial fractures occur as single, conjugate pairs, and branching types. Most of them are straight to curvilinear. The concentric fractures strike more or less parallel to the crater rim and dip toward the crater floor. The amount of dip of these fractures is also highly variable. The conical fractures are similar to the concentric fractures in their strikes but dip away from the crater center, and have a highly variable amount of dip. When the crater rim is restored to pre-impact condition, the geometry of the radial and concentric fractures resembles the pre-impact fracture populations, pointing to their origin by the reactivation process. Probably, these fractures might have also controlled formation of the squarish outline of the crater. Interestingly, the conical fractures are dissimilar to the pre-impact fractures, and could have formed purely from the impact. Post-impact slumping of crater wall rocks is also observed at a few places.

Lonar Crater: Lonar crater is a simple, bowl-shaped impact crater in the ~ 65 Ma Deccan basalt provinces in India (see summary in [3]). The crater was formed fully in basalts, almost at the same time that Meteor Crater was formed. Unlike Meteor Crater, it has a circular outline with a diameter of ~1.8 km and a depth of ~ 120 m. Structural analysis of the crater wall has documented the impact deformational structures in the massive basalt, well-exposed on the upper crater wall, where the basalt shows upward turning of the flow sequence, resulting in a circular deformation pattern. Like Meteor Crater, three fracture systems (radial, concentric, and conical) are exposed on the inner crater wall. On the fracture planes, plumose structures are common. Uplift and tilting of the basalt sequence and formation of the fractures inside the crater are clearly related to the impact event and are
different from the pre-impact structures such as cooling-related columnar joints and fractures of possible tectonic origin, which are observed outside the crater; however, the pre-impact fractures are less abundant in the target basalts. Slumping is common throughout the inner wall, and listric faulting displaces the flows in the northeastern inner wall.

**Discussion:** Impact structures documented from Meteor Crater and Lonar Crater can be related to the response of target rocks to the shock wave propagation at the time of crater excavation (e.g. [4]). Crater wall slumping is related to the modification after excavation (e.g. [5]). However, the presence of pre-existing weakness zones and their geometry would play a decisive role as to how the expanding shock wave would exploit them to form the transient crater and the subsequent collapse. In the case of Meteor Crater, the orthogonal sets of pre-existing fractures have well contributed to this process, unlike the Lonar Crater, where they are less abundant. In both craters, the target rocks are layered, but the dynamic strength of basalts is much higher than the sandstones and dolomites (e.g. [6]). An understanding of the impactor composition and the angle of impact is also necessary. The impact structures of the craters are broadly similar to those at other simple terrestrial craters in granites (e.g. [7]) and even small-scale experimental craters formed in gabbro targets [8]. However, further experimental studies are needed to understand the cratering process in a fractured rock medium to simulate the Meteor Crater impact event. Although structural analysis provides insight to shock deformation, they would also be helpful in modeling the impact parameters, provided the spatial extent of impact-related fractures is known. Xia and Ahrens [9] modeled these parameters for Meteor Crater, but it requires further refinement considering the contribution of pre-existing weakness zones to the cratering process. We plan such modeling studies for Lonar Crater in the near future.


**Fig. 2.** Equal-area plots (Schmidt net) showing the geometry of impact structures exposed on the (a) Meteor Crater and (b) Lonar Crater [3]. Abbreviations used: n–number of strike/dip measurements, r–pole density range in percent, and i–contour interval in percent.
Introduction: One of the compelling pieces of evidence for the presence of liquid water on Mars is the geomorphic expression of gullies, which are formed on the inner walls of impact craters and on the other steep slopes. For example, Malin and Edgett [1,2] initially described a class of young features on Mars that they termed gullies, consisting of an alcove, a channel and a fan (Fig. 1). Restricted to middle and high latitude locations, these features were interpreted to have originated through processes related to the presence of liquid water through groundwater discharge and in some cases by the surface runoff; the potential presence of liquid water on the surface of Mars currently or in the very recent geological past, when liquid water is metastable [3], generated a host of alternative explanations for the gullies [see summary in 4]. Detailed analysis of the conditions under which H2O could flow as a liquid in the current Mars environment shows a range of conditions under which gully-forming activity is possible [3,5]. Recent observations of changes in gullies, interpreted to mean that a few gullies are currently active [6], have intensified this discussion. Terrestrial analogs to martian environments may provide insight into the processes operating on Mars. For example, the nature of perennial saline springs forming channels on Axel Heiberg Island in the Canadian High Arctic has been used to support the argument that martian gullies formed from subsurface groundwater springs [7], and field studies in the Antarctic Dry Valleys (ADV), a hyperarid polar desert analog for Mars [8-11], have provided ample evidences for top-down melting of annual and perennial snow and ice. However, the geologic factors, which control the distribution of gullies on the crater inner wall, are still poorly understood. Terrestrial simple impact craters emplaced on sedimentary rocks are potential analogues to Mars throwing light on the gully distribution. We chose to study the Meteor Crater, which shows a spectacular development of gullies on its inner wall (Fig. 1).

Meteor Crater: Meteor Crater is a ~1.2 km wide and ~180 m deep bowl shaped impact crater formed as a result of the impact of Canyon Diablo meteorite onto the southern Colorado plateau in north central Arizona [12,13]. The crater was excavated on the >1 km thick flat-lying sedimentary rock layers resting on the crystalline basement. Like many impact craters, it shows a spectacular development of centripetal drainage pattern in response to rainwater precipitation, snow melting and groundwater discharge. Like martian gullies, the drainage system is composed of alcove, channel and fan. These gully systems are radially arranged on the inner wall of the crater. We examine these gully features as well as the entire crater wall to understand how the geologic factors such as lithology and deformation features (fractures and faults) control the location and geometry of the gullies. We argue that the understanding from Meteor Crater can well be extended to gully formation in the craters of other planetary bodies (e.g., Mars), where liquid water may play an important role in the landscape evolution.

Gully distribution: The gully distribution in Meteor Crater has the following characteristics (Fig. 1): (1) some gullies originate from the rim crest; (2) most gullies originate from the middle wall, at the contact of Kaibab dolomite beds and Coconino sandstones with deep incision in the Coconino sandstones; (3) the gullies are located along the tear faults, (4) the gullies are located along the radial fractures, (5) channels are developed in the talus deposits, (6) alluvial fans occur along the periphery of the crater floor, (7) caves are formed at the base of the Kaibab limestone-Toroweap sandstone contact.

The controlling factors: The gully location on the crater wall is principally controlled by the position of radial fractures and tear faults. Recently, we have carried out an extensive structural geological mapping in and around the Meteor Crater that point to the existence of fracture networks in and around the crater [14]. Particularly, the crater wall has a dense network of impact-related fracture systems and faults. The regionally occurring fracture-fault networks (pre-impact weakness planes in the target rocks) can be efficient zones of groundwater recharge, where the surface water can percolate. The clastic sedimentary rocks (e.g., sandstones) of the Moenkopi and Coconino Formations, which are exposed on the crater wall, are the most efficient aquifer systems in this region. The fractures/faults exposed on the crater wall are the favorable locations where surface runoff can preferentially flow causing the degradation. Also, these are the favorable locations for groundwater flow, which can be discharged on the crater walls in the form of springs. As the present-day climate in and around Meteor Crater is arid characterized by scanty rainfall and smaller quantities of snow fall during winter, the gully geomorphic features are less likely to have formed by the present climate system. Also, the present-day groundwater level is below the crater floor, and therefore, cannot form the springs on the crater.
wall. Therefore, we need to understand the role of past climate, particularly, during the Pleistocene period, when the crater was excavated, at ~50000 years ago. Climate reconstructions (e.g., [15]) suggest that, at the time of impact, the regions around Meteor Crater were wetter and that continued for long time, probably up to ~11000 years ago (see for discussion [16]). Late Pleistocene outburst flooding has been reported elsewhere in this region [17]. Consequently, the groundwater level in Meteor Crater could have been at a shallower level (~30 to 45 m) than today [12,13]. The ~30 m thick lake sediments provide significant evidence in support of the past pluvial climate, groundwater conditions and the existence of springs on the crater wall. Caves exposed on the Toroweap-Kaibab contact may point to percolation of surface runoff and selective discharge through the fractures on the crater wall. In addition, the structural uplift and shock compression at the time of impact might also have elevated the groundwater level around the crater, allowing the transient groundwater discharge forming the springs. Shoemaker and Keiffer [18] also observed an ancient soil profile developed on talus in the crater wall that may point to a higher water table, rain fall and groundwater discharge. The gully features are preferentially the sites of groundwater discharge that flowed along them. Fractures and faults must have enhanced this process.


Fig. 1. (a) Southeastern inner wall of Meteor Crater showing spectacular development of gullies. Note that some gullies are originated from the crater rim, while others are from the middle wall, where lithology changes from dolomite to sandstone, (b) gully network on the southwestern wall showing the typical alcove-channel-fan morphology, (c) gully originates where the radial fractures are exposed on the bedrock, (d) a possible example for groundwater discharge: formation of caves at the contact of Kaibab dolomites and Toroweap sandstones, (e) an example for martian gully showing the alcove, channel and fan.
MARS: POSSIBLE EFFECT OF THE WATER AMOUNT INCREASING IN THE UPPER LAYER OF THE SOIL DURING WINTER SEASON AROUND THE LATITUDES ±50° BASED ON THE TES TI AND HEND DATA ANALYSIS. R.O. Kuzmin1,2, E.V. Zabalueva1, P.R. Christensen2, I.G. Mitrofanov3, M.L. Litvak3

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Introduction. As on the Earth the resistance of the Martian surface materials (rocks and soils) to heating and cooling processes indicated by the time-dependent variations in the temperature during full daily cycle (24h, 39.6m) and defined as thermal inertia parameter $I=(k\rho c)^{1/2}$, where $k$ is thermal conductivity, $\rho$ – density and $c$ – specific heat of a materials. With high value of thermal inertia the less temperature variations are exist at a surface boundary per cycle than those with lower $I$ values [1, 2, 3]. The parameters $\rho$ and $c$ of the Martian surface materials do not vary greatly and the variations in thermal inertia are mostly caused by variations in the thermal conductivity. As far as, in the current period the surface materials on Mars regionally are characterized by both unchangeable structure and a chemical composition, their thermal inertia values may to be changeable mostly due to influence of two factors: at high values of the atmospheric dust opacity (resulted to reduction of the surface temperature amplitude) and at rising of the thermal conductivity values due to water ice (or frost) appearing in the soil porosity volume during winter season [4]. As it has been shown recently, the essential increasing of the water amount (ice + bound water) in the Martian surface layer was observed during winter season [4, 5, 6]. In the work we had conducted the joint analysis of the TES TI and HEND (neutron’s albedo) data with goal to investigate the order of the winter-time increasing of the water amount within the Martian surface layers with thickness in 3-10 cm (based on TES data) and 20-30 cm (based on HEND data).

Analysis and results. For the study we conducted mapping of the TES I and HEND fast neutrons flux measurements within one of the sectors of Mars (±50°, 230°-300°W) for summer ($L_s=$150°-160°) and winter ($L_s=$300°-310°) seasons. Both analyzed seasons are characterized by low atmospheric dust opacity [7] that excluded the influence of the parameters on the thermal inertia determination. So, comparison between the summer-time and winter-time mapping results for two of the data served as base for study of the seasonal dynamics of the water amount in the surface layer.

TES TI analysis. The results of conducted mapping of the TES TI for the selected seasons are presented on the Fig.1a,b. Mapping results demonstrates that the values of the TI during winter season in the latitude range 30°-50° (in both hemispheres) are becoming notably higher to comparison with summer-time values. Apparently, such winter-time increasing of the TI value has been provoked by appearance of some water ice (or frost) amount within the surface layer. To estimate the possible ice increase in the soil during the winter we used the nomogram, created for ice content determination [4] based on relationship between the TI dry soil and the TI icy soil values (computed for different soil’s ice content from 0% to 10%). For this the mapped summer and winter TI values were zonally averaged (in 5° latitude belts) and then were plotted on the nomogram (Fig.1c,d). The location of the plotted points relatively of the drawn curves of the nomogram let us to define what possible ice increase corresponds to the TI values observed in the winter-time. As it is seen from Fig.1c,d, the zonally averaged winter-time TI values are corresponds to the ice amount of 5-10 vol. % for latitude ranges 35°-50°N and 40°-50°S respectively. At the lower latitudes (0°-30°N and 0°-35°S) the winter-time TI values are consistent with much less soil ice increase (mostly < 2 vol. % and up to dry soil). To compile the winter-time map of the ice distribution within surface layer in the studded sector of Mars we fulfilled next procedure. From all mapped TES TI data we extracted only those TES surface footprints (from summer and winter maps), who’s geographic location coincides each with other in longitude at accuracy < 0.05°. The ice amount for the coincided TES TI footprints has been estimated by method similar to the one in [8]. Two-component mixture (soil+ice) are characterized by next thermal parameters: $\rho = \rho_{dry}\rho_{ice}+\rho_{dry}\rho_{dry}+\rho_{dry}\rho_{dry}$ and $k = k_{dry}+k_{dry}/(\rho_{dry}\rho_{dry})$. These two expressions are substituted into formula of thermal inertia. After simple manipulation, one can receive the quadratic equation: $ax^2+bx+c=0$, where $a=\rho_{dry}\rho_{ice}k_{dry}$, $b=\rho_{dry}\rho_{dry}k_{dry}+\rho_{dry}\rho_{dry}k_{dry}$, and $c=k_{dry}k_{dry}$. At that, $I_{dry}$ and $I$ represent the thermal inertia values for summer and winter seasons respectively. All parameters for ice were calculated at $T=200K$ ($\rho_{ice}=925$ kg/m³, $c_{ice}=1553$ J/kgK, $k_{ice}=2909$ W/m²K, $c_{dry}=838$ J/kg K). So, having the TES TI data for the same place ($\phi, \lambda$) from
two seasons we have solved the equations relatively unknown parameter (\(\varepsilon\) - ice vol. %) and compiled the map of the winter-time ice distribution within the studded sector of Mars for both hemispheres (see Fig.1e). Some of the equatorial areas in the studded sector of Mars have higher TI value in the summer than in winter that resulted to negative value of the ice content. This effect may be resulted by more intense loss of the water from surface layer during winter due to strong deficit of the humidity in the atmospheric boundary layer during winter season, while the local dusty environment during summer time could be also responsible for such effect. The Fig 1f shows the zonally averaged (in 5° latitude belt) of ice content as function of the latitude. From the figure one can see distinctive trend to rising of the averaged ice increase in the surface layer (3-10 cm) from 0-1 % on the latitudes 20°-30° (N,S) to 8% on the latitudes 45°-50° (N,S).

**HEND data analysis.** To study the seasonal variation of the water amount within thicker surface layer (up to depth 20-30 cm) in the same sector of Mars, we had analyzed the HEND fast neutrons flux data (with energy range 2.5-10 Mev (FN2)) accumulated during two the Martian years from the “Mars-Odyssey”. For each year we conducted the mapping of the FN2 flux data for the summer and winter seasons ranges (\(L_s=150°-165°\) and \(L_s=300°-315°\)). At this, the HEND FN2 data have been normalized on the FN2 value from the driest area on the planet, Solis Planum, where water content in the soil is less than 2% [9]. To compare with the summer maps, the maps of the winter-time FN2 flux variations in both hemispheres of the planet shown notable decreasing of the neutrons flux on the middle and high latitudes of the planet. To estimate the water equivalent in the surface layer corresponding to the mapped normalized neutrons flux distribution we used a numerical simulation method developed to convert the measured neutron fluxes to the equivalent water contents [9, 10]. The results of such converting from FN2 data to the water equivalent (for the winter season) are shown on the Fig.2a,b. Zonally averaged (in 5° latitude belt) the water equivalent values related with the summer and the winter seasons are shown on Fig.2c,d. The water equivalent differences between summer and winter seasons (for each year of the observations) in comparison with zonally averaged ice increase, derived from TES TI data, are shown on Fig.3. As well seen from Fig.2a,b, the character of the water equivalent distribution within studded sector of the planet for both Martian years is notably similar except the relatively higher water content on the latitudes 30°-50°N during first year.

**Conclusion.** The received results of the joint analysis of the TES TI and HEND data had demonstrated developing of the strong seasonal effect of the water amount variations (in the form of ice or frost) in the surface layer with thickness from \(~3-10\) cm to 20-30 cm. The winter-time ice increase (derived from TES TI data) in the soil is strongly raising in direction from the latitudes 20°-30° to 50°(N,S), approaching the averaged amount of 6-8 vol.%. For water equivalent in the deeper layer (up to 20-30 cm depth), derived from the HEND data, we observes the same tendency only for the northern hemisphere, while in the southern hemisphere the water equivalent value variations in the soil are revealed itself much slightly. Observing meridional trend of the water equivalent difference between the winter and summer seasons testify about existing of winter-time positive water mass balance in the surface layer on all northern latitudes and on the southern latitudes 40°- 50°. In the latitude belt 0°-20°S obvious negative winter-time water mass balance is observing that is indicative of losing of the soil’s water equivalent during the winter season.

![Fig.2. The regional maps of the winter-time distribution of the water equivalent in the surface layer (20-30 cm) derived from HEND fast neutrons flux data, received during first (a) and second (b) year of observation. c,d - zonally averaged (in 5° belt) water equivalent in the summer (red) and the winter (blue) seasons, derived from HEND data for the first and the second Martian years respectively.](image)

![Fig.3. Zonally averaged values of the winter-time ice increase in the surface layer, derived from TES TI data (blue and dark blue) and zonally averaged difference between the winter and summer HEND water equivalent in the deeper surface layer from two years of observations (red and violet).](image)


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MARTIAN METEORITES - CLUES TO PETROGRAPHY AND PETROGENESIS OF THE PARENT BODY. Z.A. Lavrentjeva, Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow 119991 Russia; e-mail: aval@icp.ac.ru

To date 36 individual Martian meteorites are known. Shergotty, Zagami, Nahkla and Chassigny are only observed falls among them. The remaining 32 meteorites include 12 finds from Antarctica and 17 finds from hot deserts [1].

Among the SNC (Shergotty, Nahkla and Chassigny) Martian meteorites there is a group which has several analogies with terrestrial basalt – with xenolith rocks. These rocks are shergottites. They form three main subgroups: the basaltic – shergottites (i.e. Shergotty itself), the picritic – shergottites or olivine – phyric shergottites (i.e. Northwest Africa 1068) and the herzolitic or peridotitic shergottites (i.e. ALHA 77005) [2]. Warren and Bridges [3] have recently suggested a more genetically significant way of subclassifying shergottites based on geochemistry. There are again three groups, which they define as highly depleted shergottites (HDS), moderately depleted shergottites (MDS), and slightly depleted shergottites (SDS). These geochemical subclasses correlate only loosely with the mineralogical subclasses. Most HDS are olivine-phyric; but HDS QUE94201 is a basaltic melt-rock [4] most MDS are cumulate peridotites; Yet EET79001 contains two lithologies (basaltic and phyric-xenocrystic), both MD. Most of the SDS are basaltic without conspicuous phenocrysts; yet NWA1068 [5] is olivine-phyric.

Nahklites, cumulate clinopyroxenites, are other major group Martian meteorites and have distinctly different trace element and isotopic characteristics from shergottites [6]. The nahklites are the least affected by shock metamorphism among Martian meteorites. Their lack of pervasive alteration or shock melting suggest that their magmatic melt inclusions and nominally anhydrous minerals may provide clues to the volatiles abundances and H isotope signatures of their parental melts [7].

Chassigny is the only dunite among the Martian meteorites and namesake for chassignites [8]. It consists of ~ 90% magnesium – rich olivine (Fo68) as well as minor amounts of Ca-rich and Ca- poor pyroxene (~5%), alkalic feldspar (~2%), chromite (~1.4 %), and melt inclusion in olivine. Accessory phases include chlorapatite, sulphide, ilmenite, and rutile [9]. Chassigny records a complex history involving crystal accumulation, minor subsolidus recrystallization and latter shock metamorphism. In contrast to other SNC meteorites, Chassigny experienced only moderate shock metamorphism with peak pressure ~ 35 GPa [10].

The petrogenesis of the nakhllites is thought to be as a lava flow, or series of flows extruded onto the surface of Mars some 1300 million years ago [11]. The close similarities in mineralogy and mineral chemistry amongst the nakhllites imply that they all come from the same magma body, whilst petrographic variations suggest that they are derived from different depths within a cumulate pile. The extent of equilibration within individual ferromagnesian silicate grains has been recognized as a marker for crystallization depth [12, 13], and on that basis, since MIL03346 contains the least equilibrated of all silicates, it is assumed to derive from the outermost edge of the intrusion, perhaps even a chill margin [14].

Recent studies have shown that Martian magmas had wide range of oxygen fugacities (fO2) and that this variation is correlated with the variation of La/Yb ratio and isotopic characteristics of the Martian basalts, shergottite meteorites [15, 16]. The origin of this correlation must have important information about mantle sources and Martian evolution. In order to understand this correlation it is necessary to known accurate value of oxidation state of other Martian meteorite groups. Herd et al. [17], Herd [15], Wadhwa [16], and Geodrich et al. [18] demonstrated that the fO2 in Martian basalts represented by the shergottites varies by 2 to 4 log units and is correlated with geochemical parameters such as LREE/HREE and initial 87Sr/86Sr. These correlations have been interpreted as indicating the presence of reduced, incompatible element-depleted and oxidized, incompatible element-enriched reservoirs that were produced during early stages of martian differentiation (4.5 Ga) [15-21]. Martian basaltic magmatism represented by the shergottites is thought to represent mixing between these two reservoirs. Whether this mixing process is a product of assimilation or mixing of two mantle reservoirs during melting is still a point of debate.

The water – content of Martian magmas is a topic of debate among researchers. Some Martian
basalts are characterized with melt inclusions of biotite, apatite and amphibole; phases typically associated with hydration reactions on Earth [22-24]. However, the H-content of melt inclusions from these basalts is low and bulk-rock H_2O-contents range from a meager 0.013 to 0.035 wt % in Shergotty [25]. Products of low-T aqueous alteration processes on Mars are relatively abundant in the nahklites [26]. This includes mixtures of poorly crystalline clays and amorphous material, often dubbed indingsite. They can be found in the mesostasis or forming veinlets in olivine [27] as well as being part of mineral assemblages, e.g. inside carbonate veinlets in Lafayette [28,29], in association with halite [28], or riming anhydrite in Nahkla [30], or within melt inclusions in Nahkla olivines [31].

Martian meteorites converse the fingerprints of an intense shock event that probably occurred during their ejection from the parent surface. In addition to shock-induced mineral transformations and deformations, localized fusion of the stone is a characteristic feature of the shergottite subgroup (more than half of the Martian meteorites). Shock-melting is observed either as planar (shock veins) or as spherical areas (melt pockets), where rounded sulphurs and silicate glass testify of a past liquid stage. The mechanisms that produced the heterogeneous temperature distribution within shocked meteorites are still unclear. In the case of shock veins, frictional heating was proposed to have brought the necessary heat to initiate the melting process [32]. The generation of local hot-spots, the melt pockets, remains poorly understood. Melt pockets are an important constituent of strongly shocked Martian meteorites, containing trapped noble gases, N_2 and CO_2, which closely match that of the Martian atmosphere. Interestingly, the most direct argument for a Martian origin of these meteorites relates to melt pockets, which were found to contain large volumes of Martian atmosphere [33]. Knowledge of the mechanism by which melt pockets form is crucial to understanding how the Martian atmospheric gases are implanted. Mechanism(s) for melt pocket formation can be roughly divided into two scenarios, the first involving injection of extraneous molten material into cracks add fractures in the host rock, and the second, in situ melting by void collapse by chock[34].

References:

MORPHOMETRIC ANALYSIS OF THE LUNAR SURFACE ON THE BASE OF CLEMENTINE DATA. Evgeniy Lazarev and Janna Rodionova. Moscow State University of Geodesy and Cartography (MIIGAiK), 105064, 4, Gorokhovskiy pereulok, Moscow, Russia, zhecka@inbox.ru, Sternberg State Astronomical Institute 119899, 13, Universitetskiy prospect, Moscow, Russia zhecka@inbox.ru, jeanna@sai.msu.ru

Introduction. The lunar subpolar regions are the most interesting for investigations due to such formation as “cooled traps” where water ice is suggested. The investigation of relief at these areas with high accuracy is necessary.

Clementine data is the only available one for the whole lunar surface today. In 2000 A. Cook et.al. had compiled Digital Elevation Model (DEM) on the base of Clementine 1 km/pixel stereoimages with the relative height resolution of about 100 m [3].

Techniques of investigations. To create Lunar Subpolar relief map the authors obtained heights from the A. Cook et.al. raster image of South Lunar Subpolar region (latitudes from -60° to -90°) [3] being constructed in stereographic projection. After the raster was referenced and exported to GRID-format, the next step was the obtaining database with the help of ArcView v3.3 script grid2xyz.avx [5] so that each database line was corresponded to each image pixel. The attributes of each line were coordinates of pixel center and its brightness. The size of database obtained was more than 250 megabytes or more than 6 millions points. After editing database that is the removing of wrong lines, where brightness was equal 0 (black areas with data lacking) and 255 (the white raster frame) the total amount of points became equal 4,5 millions. Then the values of brightness were counted into values of height using the next equation:

\[ h = 75 \cdot I - 8700, \quad h \text{ – point height, } I \text{ – brightness.} \]

The morphometric investigations of the Lunar South Pole region surface have been fulfilled using our database. The detailed profiles of this area created by us describe the features of this region surface with the high resolution up to 100 meters (fig.1).

Authors [3] noticed, that indefinites of absolute subpolar heights fixing are still very essential. For example, in paper [6] it is proposed positive heights. For the rim of crater Shakleton being near the South Pole of the Moon radar data from the Earth surface [7] and authors [8] as well define positive heights of 1,6 km and 3±1 km accordingly while in papers [3] and [9] crater Shakleton has negative heights of -2,9 km and -3,9 km. That is why we decided to compare the absolute heights from Rosiek et.al. Lunar map [10] with the absolute heights of our map. The comparing was fulfilled using the height profiles along -90° and 90° meridians.

Paper [10] provides synopsis of a project to collect digital elevation models (DEM) from Clementine imagery. Topographic data were derived from overlapping nadir images collected by Clementine. This technique used stereo models formed by the imagery side lap of images from adjacent orbits. A relative elevation was derived at 1 km spacing and then the digital elevation model (DEM) was adjusted to fit the altimetry data or previously collected photogrammetric topographic data and contour lines for the lunar south pole were constructed [10].

The comparison of height profiles (fig.2) constructed using our data with the profiles constructed using the map at [10] showed the differences being of about 1 – 2 km having already noticed by authors [10] themselves.

The profiles of several lunar craters have been created on the base of our database too. The crater Schrödinger (fig.3) has smoothed rim, terrace and faults, the big external rim, ridge and a lot of hills at the rough bottom. The eastern rim has height of 1,3 km and the western one has height -2,5 km, that is crater Schrödinger has asymmetric rims. The internal ring of the ridges has raised rims. The relative height of this internal ring is equal 1 – 1,5 km.

Conclusions. Based on processing of the Digital Elevation Model (DEM) created using the Clementine data [3] we compile the Lunar South Pole area database including latitudes, longitudes and heights for more than 4,5 millions points. Using database we had obtained it became possible to fulfill the morphometric researches of the Lunar Subpolar relief. The detailed height profiles of different surface areas were constructed with the high resolution up to 100 m. The comparison of the profiles created using the data [3] and map [10] shows, that relief of South Subpolar region of the Moon is represented more detailed in paper [3], while absolute heights are differed from 1 to 2 km.

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Fig. 1. Height profiles along 0° and 180° meridian along

Fig. 2. Height profile along meridian -90° and 90°

Fig. 3. Height profile of Schrödinger crater along -74° parallel.

EXPERIMENTAL STUDY OF IRON-SULPHIDE AND SILICATE PHASE SEGREGATION IN A CENTRIFUGAL FIELD. E.B.Lebedev. Institute of Geochemistry and Analytical Chemistry, RAS, Moscow, Russia, leb@geokhi.ru

A simulation of the migration and accumulation of sulfide phases (FeS) under gravity and mechanical deformations, with the partial fusion of a model planetary substance (olivine-picrite mixture), was carried out in a high-temperature centrifuge. The separation and motion of sulfides in the intercrystalline space is shown to be in an intimate relationship with the degree of fusion of a silicate material.

An occurrence of large-scale Earth’s material melting processes is assumed at the early stages of Earth formation. This melting comprised the carbon and sulfide containing rocks, which composed the high horizons of the mantle. The metal and sulfide phases' occurrence is assumed for these rocks.

The determination of possible condition for the metal and sulfide phases segregation from partial melting zones, its accumulation and movement to the forming Earth’s and Moon’s cores are the important problems.

Sulfide and metal phases have a particularity in its surface energy, which differs them from the silicates when this phases accumulate in the inner Earth.

Experiments temperature was 1200-1450°C. The centrifuge ensured acceleration of 2000-4000 g. Centrifugation was carried out on mixture of iron-sulfide (FeS) and silicate matrix contained of olivine crystals and basic melt (picrite) with different proportions.

The initial mixture was: olivine crystals (85%) and basic melt (picrite) 10%; iron-sulfide melts 5%FeS (95%Fe, 5%S).

The permeability increases manifold, if the silicate-metal system is subjected to mechanical deformations (Rushmer, 1995). A number of papers deals with the experimental study of the silicate system are subjected to deformations (McKenzie, 1984; G. Hirth' and D. L. Kohlstedt, 1995; E. H. Rutter and D. H. K. Neumann, 1995; Karakin A.V., 1999; J. W. Hustoft and D. L. Kohlstedt, 2006).

Our experimental results showed, that mixture consisting of olivine crystals, silicate and iron-sulfide melts, after being separated in a centrifuge, is differentiated in density at definite physico-chemical conditions. Separation is observed more intensive at lg fO2 some low IW. It was evaluated also shock deformation with using of the ultrasonic method and mechanical deformation of silicate matrix on iron separation. Separation is observed more intensive at mechanical deformation of silicate matrix.

Based on the data obtained, the mechanisms of the chemical differentiation of planetary bodies and of the formation in them of iron-sulfide cores in the thermal and gravitational fields are discussed.

Work is supported by RFBR, grant N 07-05-00630 and PBR Pr. RAS N 18.

POLYGONS AS PART OF COLD-DESERT, NEAR-SURFACE FLUVIAL SYSTEMS. GULLY-POLYGON INTERACTION AND STRATIGRAPHY ON EARTH AND MARS: SAND-WEDGE POLYGONS AS PART OF COLD-DESERT, NEAR-SURFACE FLUVIAL SYSTEMS. J.S. Levy1, J.W. Head1, D.R. Marchant2, G.A. Morgan3, and J.L. Dickson1, 1Dept. of Geol. Sci., Brown Univ., Providence, RI, 02912, USA (joseph levy@brown.edu), 2Dept. of Earth Sci., Boston Univ., Boston, MA 02215 USA.

Introduction: Gullies on Mars are a class of young features initially interpreted to have formed by flow of liquid water through groundwater discharge from underground aquifers [1, 2], and are possibly still active [3]. Numerous alternative explanations have been proposed (see summary in [4]), and recent terrestrial analog work in the Antarctic Dry Valleys (ADV) has reported on the primacy of top-down melting of snow as a source for water flowing through the active layer and hyporheic zones of these terrestrial gullies [5-8]. Although gullies are found in periglacial terrains on Earth and Mars [9-11], the effects of polygonal patterning on the generation and modification of gullies has not been broadly discussed. We present field measurements and observations that suggest polygonally patterned ground (sand wedge polygons) contributes to the production, transport, and storage of liquid water in gullies located in Wright Valley, Antarctica. Further, stratigraphic and morphological relationships between gully and polygon features in the ADV provide a sequence of feature formation. Analogous features on Mars suggest that similar processes may be forming and modifying gullies in both environments.

Production of Meltwater. Water flowing through ADV gully systems is produced largely through the melting of perennial snowbanks in the gully alcove and the melting of wind-blown snow trapped in the gully channel [5-8]. Polygon troughs also act as accumulation sites for wind-blown snow (Fig. a); troughs intersecting gully channels provide a direct, downslope path for meltwater to the gully and marginal hyporheic zone (Figs. e, g) [7, 10]; polygons in the upper reaches of the gullies can provide snow accumulation comparable to snow accumulation in the gully channel. Analogous relationships are present on Mars (Fig. b); frost is present in polygon troughs within and surrounding gullies.

Water Transport. Polygonal patterned ground promotes surface and shallow subsurface flow of water in ADV gullies. Trough Annexation. Where gully channels intersect sand wedge polygon troughs oriented down-slope, gully channels change direction to follow the polygon trough (Fig. c). Gully channels follow troughs for tens of meters until the trough ceases to tend down-slope. Annexed portions of polygon troughs are widened and sinuous on ~1 m wavelengths; channel-annexed trough intersections with adjoining polygon troughs are commonly rounded rather than angular. Analogous channel-polygon relationships are present on Mars (Fig. d).

Water Storage. Water flowing through ADV gully systems accumulates in distal hyporheic zones [6, 7], which are patterned by modified sand wedge polygons. “Wet-topped polygons” (WTP) have dark polygon centers surrounded by light polygon troughs (Fig. f). The albedo difference between WTP centers and troughs is due largely to a difference in soil moisture (4-5 wt.% in the dark interior surfaces, <0.5 wt.% in the light trough surfaces). Water stored in distal hyporheic zones in Wright Valley is strongly partitioned into polygon interiors over polygon troughs. Despite the seasonal generation of meltwater, no ice-wedge polygons have been observed.

Stratigraphic Relationships. Overprinting. Gully fan deposits overprint polygonally patterned desert pavement surfaces in Wright Valley (Fig. i). Gully fans consist of bedded and cross-bedded pebbles and sands, which have been locally deflated. Contacts between fan material and patterned surfaces with well-developed pavements are gradational. Meter-scale outcrops of polygonally patterned colluvium with well-developed desert pavements are present within the fans, evidence of emplacement of locally high patterned ground areas by fan deposits. Subdued polygonal patterning is present within fan deposits, and is interpreted as continued expansion of thermal contraction crack (sand wedge) polygon troughs during and subsequent to fan deposition. Analogous features are present on Mars (Figure j). Truncated Polygon Troughs. Several polygon troughs are cross-cut by gully channels in Wright Valley (Fig. g). Polygon troughs intersect gully channels at near-orthogonal and oblique angles. Troughs present on one bank of the gully channel can commonly be matched to troughs present on the opposite bank. Taken together, these two lines of evidence suggest that the polygon troughs were cut by the gully channel, rather than forming subsequent to gully channel formation. Analogous relationships between polygonally patterned surfaces and gully channels/alocoves are observed on Mars (Fig. h).

Discussion and Implications. Observations of relationships between polygonally patterned ground and gullies in the Antarctic Dry Valleys suggest that the presence of sand-wedge polygons: 1) enhances the accumulation and transport of gully meltwater, 2) affects the partitioning of gully water in distal hyporheic zones, and 3) can locally affect the course of gully channels by trough annexation. Stratigraphic relationships between polygonally patterned terrain and gully
features provides insight into the sequence of processes modifying the ground surface in Wright Valley. Overprinting of polygonally patterned ground by fan deposits and cross-cutting of polygon troughs by gully channels further suggest that gully formation occurred subsequent to the initial formation of patterned ground in Wright Valley. We interpret the presence of subdued polygonal features within the fans as evidence that sand-wedge polygon evolution has continued subsequent to burial by fan deposits, suggesting that climate conditions have remained suitable (cold and arid) for sand-wedge polygon evolution during the entire process of gully formation.

The presence of polygonally patterned ground in Wright Valley subsequent to the formation of gullies requires the presence of shallow permafrost prior to gully formation. The continuing evolution of polygons in Wright Valley during gully formation implies the continued existence of the permafrost layer for the duration of gully processes, as the presence of a thick permafrost layer acts as an impermeable layer at depth over which snowmelt-derived water flows [5-8]. The presence of similar morphological features on Mars, with similar relationships between gullies and polygons suggests that analogous processes may have occurred in recent geological time on Mars.


Figures. a) Snowbank in polygon trough in ADV; 60-80 cm across short axis, and < 10 cm thick. b) Frost-filled polygon troughs in Terra Sirenum gully, Mars. Scale is 10 m. c) Polygon trough annexation in ADV. d) Vertical arrow, expanded polygons; horizontal arrow, sinuous depression, indicative of possible trough annexation in Terra Sirenum, Mars. e) Polygon trough cross-cut by ADV gully channel. Arrow indicates expansion of hyporheic zone. f) Polygon trough cross-cut by gully channel in ADV.

g) Wet-topped polygons in ADV; polygons are 10-20 m across. h) Cross-cut and non-cross-cut polygon troughs in Terra Sirenum, Mars. i) Gully fan (100 m wide) overprinting polygons in ADV. j) Gully fan overprinting polygons in Terra Sirenum, Mars. Bar is 50 m.
THE SEIS-EXOMARS EXPERIMENT: A PLANETARY SEISMOMETER FOR MARS. D. Mimoun¹, P. Lognonné¹, P. Schibler¹, W. T. Pike², D. Giardini³, Ulrich Christensen⁴, Arie van den Berg⁵, and the SEIS-ExoMars team, ¹IPGP (4 avenue de Neptune, 94107 Saint-Maur cedex, France, mimoun@ipgp.jussieu.fr), ²Imperial College (Exhibition Road, London SW7 2BT, England, w.t.pike@imperial.ac.uk), ³ETH (Institute of Geophysics CH-8093 Zurich, giardini@seismo.ifg.ethz.ch), ⁴Max-Planck-Institute for Solar System Research (Max-Planck-Strasse, 237191 Katlenburg-Lindau, Germany, christensen@mps.mpg.de), ⁵Institute of Earth Science (Utrecht University, Budapestlaan 4, 3584 CD Utrecht, NL, berg@geo.uu.nl)

Scientific objectives: The SEIS-ExoMars seismometer included in the GEP payload (Geophysics Package) of the ESA ExoMars mission will study the seismic activity of the planet and the frequency of meteorites impacts. These seismic events will be characterized by their approximate distance and azimuth, as well by their magnitude. The seismometer will also allow also to characterize shallow and deep interior of the planet, and especially the water environment as a function of depth in the deep subsurface, the crustal thickness of the landing site, the core size and possibly (if the seismic activity is between the middle and upper bound of present estimates) the mantle structure. The sensitivity and noise floor of the seismometers in the expected Martian environment are such that the detection of about 20 quakes with Ms magnitude from 4 to 5 and 10-20 impacts per year are expected for a mean model of seismic activity; our working hypothesis is based on the thermoelastic cooling of the lithosphere, which does not consider any tectonic activity possibly related to volcanoes

Instrument configuration: The seismometer will be powered and serviced by the GEP. It is based on an hybrid 4 axis instrument, composed of 2 Very broad Band (VBB) sensors and 2 Short Period (SP) sensors and has a mass of about 1700 gr, excluding all margins. This design reflects a significant mass reduction compared to design studied by previous ESA projects (i.e. MarsNet and InterMarsnet), while offering very little science return reduction as compared to a more classical 3 VBB +3 SP design.

Responsabilities: IPGP (F) has the overall responsibility of the experiment (development, delivery, testing and operation of the whole instrument). IPGP is in charge of management and system engineering aspects.

From hardware delivery point of view, IPGP is responsible for the sphere including VBB axis and environmental sensors, IC (UK) is responsible for the SP (Short Period) sensors, ETHZ (CH) is responsible for the electronics of the experiment (SEIS-AC subsystem), MPS (D) is responsible for installation and deployment devices (SEIS-DPL subsystem) and SRON (NL) is responsible for the delivery of mixed-signal ASICs to be integrated inside SEIS-AC subsystem.

Programmatic status: The Seismometer has been recommended for implementation on the European Space Agency’s Exomars mission during the 2005 Birmingham meeting. Following this meeting, an unsolicited proposal has been submitted to ESA by DLR and IPGP for the inclusion of a Geophysics Package (GEP) to the initial ExoMars Rover Payload. We are currently during the Payload Confirmation Review phase, with a first confirmation at spring 2007 and the finalization of the payload selection end of September 2007. On the technical point of view, the TRL level of the instrument is around 6/7 for the sensing part, and 2/3 for the deployment system, which is strongly linked to the descent module interface.

A breadboard has been delivered by industry (EADS-Sodern) in July 2004. Most critical parts have been tested, including shock tests (200g, 20 ms) for pivot, electronics components and displacement sensors. The electronics breadboard has also been delivered and tested. PDR is planned at spring 2008.

Performances: functional results are satisfying and noise optimization is under process. Preliminary noise results are encouraging. Martian noise is close to our STS2 terrestrial reference instrument.

References:


CHARACTERIZATION OF INTERMEDIATE UNITS AND LAYERED DEPOSITS WITHIN THE LVF/LDA DEPOSITS OF THE DICHOTOMY BOUNDARY OF MARS. Gareth A. Morgan1 and James W. Head1, 1Dept. Geol. Sci., Brown University, Providence, RI 02912 (gareth_morgan@brown.edu).

Introduction: The northern dichotomy boundary on Mars consists of an abrupt escarpment between the cratered highland and the northern plains. The boundary is characterized by fretted valleys which divide the highlands into a series of plateaus and mesas which are progressively smaller to the north and eventually merge with the northern plains [1]. Lineated Valley Fill (LVF) and Lobate Debris Aprons (LDA) deposits are a prominent feature of the dichotomy boundary and have been attributed to be the remains of debris covered glaciers [e.g. 2, 3] that were active during previous climatic conditions when snow was deposited along the dichotomy boundary [4]. Individual lobes of LVF have been observed emanating out of multiple alcoves within plateau walls and merging with main trunk deposits of LVF [2,3]. Flow lines along the surface of LVF deposited have been traced within high resolution images to form large scale integrated systems extending over areas of > 10,000 km² [3].

In a review of the observations made by the Mars Global Surveyor spacecraft of the dichotomy boundary, Carr (2001) characterized the slopes of the fretted terrain as consisting of 3 main components: 1) steep upper slope, bedrock may be visible, 2) an intermediate unit (IU) which appears smooth, and may have faint striae in the downslope direction, 3) debris apron (LDA). In addition to this layered deposits (LD) were also identified within the region. Since the publication of these observations, three additional spacecraft have been in operation in orbit around Mars. We have utilized these new data sets in conjunction to the MOC archives to reexamine the LVF/LDA deposits and provide hypotheses for the units characterized by [5].

Intermediate Units: IUs can be observed between slopes and both LDA and LVF deposits across the dichotomy region. The units are typically ~ 2 km wide and can extend continuously for up to several hundred km along the base of plateau slopes and completely surround the flanks of mesas (Fig. 1). As Carr (2001) observed, the IU appears to be draped up on the slopes and over the edge of the LVF/LDA surfaces in some areas, an observation which is consistent with the more recent data sets. HiRISE images have revealed that the striae consist of elongated ridges, several m across, and that patterned ground, possibly consisting of contraction crack polygons are present along the surface of some of the IU (Fig. 2). The IU are well defined in nighttime THEMIS IR images and appear bright relative to the adjacent LVF/LDA surfaces, indicating that the surface of the IU consists of a material with a relatively high thermal inertia. This in turn is a proxy for grain size of the surficial deposits, suggesting that the unit consists of a coarser material than that of the LVF. Hence we interpret the IU to represent a relatively recent debris cover (which has been sourced from the exposed slopes) which overlies the older LVF/LDA surfaces. In contrast we interpreted the LVF/LDA sur-
faces to be comprised of finer, reworked debris and may also be covered in a significant amount of dust that has become trapped within the pits along the surface and which would account for the lower thermal inertia. If this interpretation is correct then the IU may represent a martian equivalent of lateral moraine deposits protecting the underlying ice from sublimation under the current martian conditions and thus accounting for the lack of sublimation pits on the surface of the IU. The existence of polygons adds weight to this argument as the presence of *sublimation* type polygons on Earth (the most likely candidate for Mars due to their formation not requiring liquid water) is indicative to the occurrence of massive ice beneath a sediment cover [6].

As the individual lobes of LVF (which feed the main trunk LVF deposits) are enclosed within steep-walled alcoves, the surfaces of the lobes are also protected from the effects of sublimation. This has allowed many primary features associated with the lobes when they were active, including those similar to terrestrial glacial features (including marginal and Bergschrund crevasses [7]), to have been preserved and made distinguishable from processes relating to sublimation that have operated along the surface of the LVF/LDA deposits since they became dormant [e.g. 8].

Evidence for mass wasting is not obvious along the upper portion of slopes above the intermediate and LVF/LDA deposits. Instead the plateau flanks and surfaces appear smooth to granular within the high resolution images. This texture is abundant from the mid-latitudes to the poles within both hemispheres and has been attributed to an ice rich mantle unit that was deposited during the last Martian ‘ice age’ associated with the high obliquity (30-35°) event that occurred 2.1 to 0.4 Myr ago [9]. Such deposits are two orders of magnitude younger than the minimum surface ages of LVF/LDA deposits estimated by crater counts [e.g. 7,8], and so the mantle would have been blanketed over all the features associated with the LVF/LDA emplacement. However, in some areas, particularly equatorial facing slopes, the mantle is absent or significantly thinned (most likely the result of increased sublimation due to the direct insolation experienced along such slope orientations) and the underlying slope surfaces are exposed. In such instances the slopes exhibit alcoves, erosional scars and other features attributed to mass wasting and sub-aerial erosion (Fig. 3).
A new era of international lunar exploration has begun and will expand over the next four years with data acquired from at least four sophisticated remote sensing missions: KAGUYA (SELENE) [Japan], Chang’E [China], Chandrayaan-1 [India], and LRO [United States]. It is recognized that this combined activity at the Moon with modern sophisticated sensors will provide unprecedented new information about the Moon and will dramatically improve our understanding of Earth’s nearest neighbor. It is anticipated that the blooming of scientific exploration of the Moon by nations involved in space activities will seed and foster peaceful international coordination and cooperation that will benefit all.

Summarized here are eight Lunar International Science Coordination/Calibration Targets (L-ISCT) that are intended to a) allow cross-calibration of diverse multi-national instruments and b) provide a focus for training young scientists about a range of lunar science issues. The targets, discussed at several scientific forums, were selected for coordinated science and instrument calibration of orbital data. All instrument teams are encouraged to participate in a coordinated activity of early-release data that will improve calibration and validation of data across independent and diverse instruments.

As a whole, the small group of targets also provides a good introduction to lunar science. The process of understanding the character of these few areas is intended to educate and spark a desire to explore further with the more extensive data produced by the various missions. These few targets provide a common starting point for much discussion and comparison among the science community and for the public to become reintroduced to the mysteries and excitement of lunar exploration.

<table>
<thead>
<tr>
<th>L-ISCT</th>
<th>Longitude (E)</th>
<th>Latitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Apollo 16 Highlands</td>
<td>15.5</td>
<td>-9.0</td>
</tr>
<tr>
<td>2. Lichtenberg rim</td>
<td>293.0</td>
<td>31.5</td>
</tr>
<tr>
<td>3. Apollo 15 (Hadley Rille)</td>
<td>3.7</td>
<td>26.1</td>
</tr>
<tr>
<td>4. SPA (NW-N)</td>
<td>175.5</td>
<td>-30.5</td>
</tr>
<tr>
<td>NW-S</td>
<td>165.0</td>
<td>-41.0</td>
</tr>
<tr>
<td>5. Tycho</td>
<td>348.8</td>
<td>-43.3</td>
</tr>
<tr>
<td>6. Polar Shadows</td>
<td>118.0</td>
<td>-84.0</td>
</tr>
<tr>
<td>7. N. Schrödinger</td>
<td>135.0</td>
<td>-72.4</td>
</tr>
<tr>
<td>8. Mare Serenitatis (MS2)</td>
<td>21.4</td>
<td>18.7</td>
</tr>
</tbody>
</table>

The eight L-ISCT targets were selected to meet several criteria for coordinated or repeated measurements [discussed in Pieters et al., COSPAR in press]. A few targets are relatively homogeneous, while others exhibit diversity in morphology, composition, etc. All are approximately 200x200 km in dimension to allow instruments with different spatial footprints to be cross-compared. Since the field of view varies greatly between instruments, it is recommended that instruments with a small footprint target as close to the central portion as is feasible. Although specific calibration steps are unique to each instrument, this limited set of common calibration targets will allow cross validation of instruments using independent information. When L-ISCT data from different instruments are in agreement, confidence in the measurements is high and all teams benefit. If data are not in agreement, possible sources of error can be sought and resolved. For mutual benefit, it is recommended that LISCT data be publicly released soon after initial calibration. Teams are encouraged to establish ties with similar teams on different missions to allow early comparisons (and improvement) of data using the L-ISCT.
LISCT 3. Apollo 15 [Hadley Rille & Imbrium rim]

LISCT 4: SPA N area [Birkeland crater Th high]

LISCT 5: Tycho crater [prominent fresh crater]

LISCT 6: Polar shadows [scattered light test]

LISCT 7: North Shröndinger [small polar basin]

LISCT 8: Mare Serenitatis [two mare basalt types]
EXPLORATION OF THE MOON’S THERMAL EMISSION FROM THE DATA OF THE CLEMENTINE SPACECRAFT AND OF THE GOMS ARTIFICIAL EARTH SATELLITE. S.G. Pugacheva, V.V. Shevchenko. Sternberg State Astronomical Institute, Moscow University, 13 Universitetsky pr., 119992 Moscow, Russia, pugach@sai.msu.ru.

Introduction. New satellite measurements of the lunar surface radiation temperature were used to study of thermal radiation of the Moon in the infrared (10.5-12.5 micron) spectral range. The basic material for investigations are the scanned cosmic spectrozonal images of the lunar surface transmitted by the first Russian geostationary artificial meteorological satellite “GOMS” and digital images of the Moon acquired by the Clementine spacecraft. In this paper we describe an analytic model for the thermal field, which is realized as an angular function of the thermal infrared radiation emitted by the lunar surface and analyze thermal anomalies of the lunar surface.

The Russian Geostationary Artificial Meteorological Satellite “GOMS”. This satellite was placed in a circular orbit on October 31, 1996, in accordance with the program “Meteorological Service for the Population”. The orbit’s altitude is 35800 km, and the standing point is a longitude of 76°E. The artificial satellite had the onboard television complex (BTVC), whose optical system transmitted the real-time digital images of terrestrial clouds, snow cover, and ice cover. The IR channel of the BTVC recorded the radiation temperature of the oceanic surface and of the upper boundaries of clouds. The optical system of the spacecraft has a mirror objective 400 mm in diameter. The instantaneous fields of vision are 6.3 and 22.5 arcsec in the visible (0.4-0.7 micron) and in the IR range (10.5-12.5 micron) respectively. The infrared channel records thermal fluxes from objects with radiation temperature between 213 and 313 K. With certain geometry of the observation and illumination, the lunar disk was seen in the frame of the BTVC objective simultaneously with the Earth’s Image (fig. 1).

Figure 1. An infrared image of the Earth, obtained aboard the GOMS geostationary spacecraft. Upper left: an image of the Moon.

This circumstance predetermined the choice of the Moon as a natural object with steady-state characteristics of the reflected and own radiation for calibrating the instruments onboard the satellite. To calibrate the electrical signal of the on-board television complex, the computer simulates on its monitor a digital image of the Moon’s brightness in the visible ad IR range [1, 2]. The computer images are simulated from the photometric database of the ground-based observations [3]. Digital images of the temperature of the lunar surface are shown in fig. 2.

Figure 2. Cosmic image of the lunar surface in the IR range (10.5-12.5 μm), obtained on July 15, 1996, at the Moon’s phase angle of +35.4°. Photograph no. 07151400.r45. Contours represent surface-temperature variations within the isothermal-latitude range 250.0-394.9 K.

The Spatial Angular Function of Thermal Emission of the Moon. When constructing the angular function, we used as the input parameters the Moon’s obtained by GOMS, as well as the results of ground-based measurements of the thermal lunar-surface radiation [4]. At a fixed incidence angle, the angular thermal-radiation function is depicted in the rectangular coordinates \((x, y, z)\) as the surface described by the radiation temperature vector in the range of positive values of the angular parameters: the incidence angle \(i\), the reflection angle \(e\), and the azimuthal angle \(A\) between the plane of the incident and reflected rays. The analytic expression for the lunar-surface thermal radiation is a trigonometric function whose...
arguments are the values of the angular parameters $i$, $\varepsilon$, and $A$:

$$T_{A=0} = [110.233 - 95.070 \cos(i)] \times \cos(\varepsilon - i) + 186.364 \cos(i) + 199.283,$$

$$T_{A>0} = T_{(A=0)} \times \{1 - 3798.767 \times \cos(i) \times \varepsilon \times A\}.$$

The root-mean-square error in the determination of the radiation temperature is $\pm 0.94\%$ for mare regions and $\pm 2.24\%$ for highland regions.

Thermal infrared imaging of the Moon from Clementine. The main instrumentation on Clementine consists of four cameras, one of which was a long-wave infrared (LWIR) camera. The LWIR camera used a catadioptric lens with a $128 \times 128$ HgCdTe FPA. The FPA was operated at 65 K. Wavelength range was controlled by the cold filter to 8.0 to 9.5 $\mu$m. Brightness temperatures were calculated with the Planck function for emission from a blackbody assuming unit emissivity; brightness temperatures range from a high of about 380 K on the equator-facing interior southern rim to a low of about 270 K on the southward-facing interior northern rim.

A comparison between theoretical values of thermal radiation and numerical data of the measurements from Clementine. A comparison the common thermal models and results measurements of thermal emission show a systematic departure of the measured values from the average values. These deviations, depending on the surface albedo, characterize the photometric inhomogeneity of the lunar surface layer. The differences of temperature of the lunar surface layer indicate the extremely low heat conduction and high porosity of the material. Major factors of the photometric inhomogeneity are strong irregularities of the relief and the varied heat conduction of the lunar ground. We have compared the Lunar Prospector data of thorium and iron contents [5] and values IR radiation of the surface for landing sites. Figure 3 represents the diagram of relationship between fluctuation of thermal emission and local thorium and iron content in different lunar regions. The lines show a mean polynomial trend.

The correlation coefficients are 0.85 (Th) and 0.88 (FeO). The content of the elements Potassium (K) and Uranium (U) at lunar surface is showed on the figure 3.

The separate points represent areas of number of landing sites: Surveyor I, III, V, VI, VII, Lunokhod I and 2, Apollo 11 and 12, and an area in sinus Media (Lunar Orbiter II).

Conclusions. The distribution of the radioactive elements (U, K, Th) at the lunar surface is an important scientific task for investigations of lunar evolution. The radioactive elements have provided continuous heat over the lifetime of the Moon. The concentration of the radioactive elements provides a suitable condition for radioactive heating and basaltic flow to the basin. Probably Th and FeO enter into composition of ejecta lunar materials. KREEP-rich materials concern to mare basalt with a high content FeO. The local assimilation KREEP-rich materials ascribed to volcanic extrusions released or localized by impact and essentially influence on thermal balance of the Moon.

FACTORS OF REGOLITH FORMATION AND TRANSPORTATION ON SMALL BODIES. N.V. Pupysheva1,2, and A.T. Basilevsky1. 1 - Vernadsky Institute, Moscow, Russia; 2 – Department of Geography, Moscow State University, Moscow, Russia pypisheva@mail.ru.

Introduction. Factors controlling formation and transportation of regolith on asteroids, comets and small satellites include meteorite impacts, the body self-gravity and centrifugal forces \[3, 8, 9, 10, 12, 16, 18\]. For very small particles electrostatic levitation due to Solar photons can also play some role \[4, 5\]. For comet nuclei, sublimation of ices and related processes play the major role when comet is close to the Sun \[7, 11\]. At large distances, factors controlling formation and transportation of regolith for comets are probably the same as for asteroids and small satellites \[1, 2, 17\]. Meteorite impacts are responsible for fragmentation of body materials, for formation and movement of ejecta of impact craters, and for the impacted body seismic shake. Self-gravity controls the crater ejecta movement and movement of the surface material downslope. Centrifugal forces caused by rotation of the body around its axis and by the body orbital movement around the central body, if they are strong enough, may affect movement of the surface material particles and in some specific cases even destroy the body \[6, 13, 14, 15, 19\]. In this paper we concentrate on interplay of gravity and centrifugal forces.

Centrifugal forces due to rotation around body axis v.s. body self-gravity. This is applicable to asteroids and comets, which rotate around their axes rather quickly. We have taken from the literature and web-sources the body sizes, rotation periods and surface gravities. In a few cases when the body surface gravity was not known we roughly estimated the latter assuming that it is directly proportional to the ratio of this body radius to the radius of the body whose surface gravity is known and which belongs to the same class of the objects (comet, asteroid of certain class, and so on). Then for the bodies under consideration centrifugal forces for their equators have been calculated assuming that it is directly proportional to the ratio of this body radius to the radius of the body whose surface gravity is known and which belongs to the same class of the objects (comet, asteroid of certain class, and so on). For example, at the ends of elongated bodies the \(q/g\) ratio, increases to 0.6 for Ida and 0.43 for Eros. For nonequatorial body regions the centrifugal effect decreases coming to zero on the poles. For the rest bodies the centrifugal effects are obviously small even in the equatorial zones.

Centrifugal forces due to body orbital movement around its central body v.s. the body self-gravity. These were expected to be noticeable for planetary satellites, many of which are close enough to their planets to have synchronous rotation around their axes. So we have calculated the centrifugal forces due to the body movement around its central body and plotted them v.s. the small body self-gravity (Figure 2). Sources of the data for calculations were the same as in previous section of this paper.
and this is why this body is a satellite. Hower this balance is perfect for the whole satellite and for a “plane” being normal to the planet gravity vector and going through the satellite center of mass, but for other parts of satellite there is some disbalance between the centrifugal force and the central body gravity attraction. In the anti-planet point of satellite, distance from the center of rotation is larger than in the satellite center of mass, but angular speed of rotation is the same. This increases the centrigugal force by \( dq \).

Besides, because the anti-planet point is more distant from the planet center of mass, the planet gravity here is smaller by \( dg \). As a result, differential centrifugal and gravity effects generate force directed from the planet and from the satellite center. In the anti-planet point this force is directed against the satellite self-gravity. For sub-planet point the interplay between the centrifugal force and planet gravity attraction leads to appearance of force directed towards the planet that is again against satellite self-gravity. We calculated these effects, normalized sum of them by the satellite surface gravities (\( g \)) and then plotted v.s. the surface gravities (\( g \)) (Figure 3).

![Figure 3](image)

Figure 3. Normalized sums of differential orbital centrifugal and differential planet gravity attraction forces \( (dq+dg) \) for satellite anti- and sub-planet points v.s. satellite mean surface gravity (\( g \)).

As it is seen from Figure 3 for several satellites: Pandora, Phobos, Prometheus, Epimetheus, Amalthea and Janus the effects decreasing surface gravity are about 0.1g and thus potentially may influence movement of crater ejecta and surface materials. Outside anti- and sub-planet points \( dq+dg \) become to be smaller and oblique to vectors of satellite self-gravity.

For asteroids and comets distances from the center of rotation (Sun) are very large so comparing to them the body diameter is very small and values \( (dq+dg)/g \) are very small: \( \exp(-7 \text{ to } -8) \) and obviously do not influence the surface processes.

In future work we plan to do a comparative study of surface morphology of small bodies in the body areas where the interplay of centrifugal forces and self-gravity is relatively large and where it is small to negligible. One of the study objects will be asteroid Eros (Figure 4) which is rather large (33 x 13 x 13 km) and has elongated shape in combination with relatively large size (33 x 13 x 13 km) provide rather large \( q/g \) (0.43) at its distal ends while on its poles this parameter is close to zero.

![Figure 4](image)

Figure 4. Asteroid Eros, image taken by NEAR spacecraft. Courtesy of NASA.

Another study object will be Martian satellite Phobos (Figure 5) which is rather large (27 x 21 x 18 km) orbiting fast close to Mars. In its anti- and sub-Martian points \( (dq+dg)/g = 0.108 \).

![Figure 5](image)

Figure 5. Phobos, image by MEX HRSC. Courtesy of ESA.

We acknowledge help of V.V. Pupyshev.

References:
Introduction: Geologic evolution of Claritas Fossae (CF) on the SSE slope of the Tharsis bulge was characterized by various energy releases. The fluvial history of southern CF has had a paleolake phase [1,2]. There is a need to study the fluvial, erosion and sedimentary features of the CF area and incorporate them to provide an understanding of their interplay with the rift structures. Our goal is to add more details to the understanding of the general development of the area as part of the evolution of the Martian geology and climate. The concept of tectonics that co-acted with climate-related events provides a framework to understand some geologic aspects of this area. Morphologic details help to identify the interplay between the processes and the related basin morphology development, and the valley deformation most probably connected to climate and tectonics. Melting of snow and ice due to aerothermal heat was already proposed [3] to have been responsible for the channel formation on the regional slopes, but the other factors may have been as important.

Data: The CF area is covered by the maps MC-17 and MC-25. The MEX HRSC data set covers the entire CF area with a 12 to 50 m/pixel resolution [4] and provides detailed 3D views of tectonics, structures and surface color. The topography was derived from the 16 bit 128 pixel/degree MGS MOLA DTM data [5]. MO THEMIS provided image resolution of 18 m/pixel (VIS) and 100 m/pixel (IR), respectively [6]. The accurate (= 4 m/pixel) MOC NA images [7] were used to find small-scale features. The very first MRO HiRISE images available over smallish areas were also incorporated [8].

Tectonics vs. hydrology: Tectonics of CF has included several deformation phases (Fig. 1). The E-W grabens belong to the oldest phase. They are seen on the elongated first-order NWW-SEE antiforms that are associated with the N-S Claritas Rupes (CR) fault on its western side. The wide set of conjugate N-S, NNE-SSW and NNW-SSE grabens are formed in several deformation events. The CR fault and the CF grabens form a rift zone on the CF bulge.

The multi-temporal tectonic events were accompanied by changes in climate and hydrology over a period of time (Figs. 2, 3) as seen from the fact that some channels pre- and other post-date the faults of the very same set. Channels were frequently re-arranged by tectonics. Only some basins on CR (Fig. 1) provided temporal volatile reservoir.
Sources of the CF water: The hill slope alcoves (Fig. 4) resembling glacial amphitheatres indicate ice accumulation areas. Glacial-type U-valleys lead down from them. Release of water from the volatile-rich hilltops eroded the lower slopes and resulted in faint channels originating from the deposits. An amount of water penetrated to the ground and resulted in permafrost, ground ice and groundwater that led to further conduit formation along faults and to sapping events. This was repeated along the climate change cycle and resulted in frequent hydrology events that were correlated with tectonics.

Further consideration: The interwoven activity phases of CF includes the rift development that had its driver in the Martian interior. Tectonics was complicated by volcanism and hydrology. The faults provided aquifers for a substantial part of the water that accumulated on the high mid-latitude hills, and even water from the Tharsis volcanoes [11,13] may have utilized the CF rift. The broken uppermost surface allowed the water to erode flow channels and channel networks while the groundwater followed faults and weak zones carving conduits and welling in places into the surface to form sapping structures. The aquifer system activated repeatedly due to climate changes and it resulted in new erosion and deposition events. Additionally, temporal groundwater and -ice may have affected faults by erosion and fault lubrication.

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MICROPHYSICS OF TITAN ATMOSPHERIC AEROSOL. A. V. Rodin1,2, Yu.V.Skorov3,4, H.U.Keller4, and M.Tomasko5, rodin@irn.iki.rssi.ru, 1Moscow Institute of Physics and Technology, Dolgoprudny, 141700, Russia, 2Institute for Space Research, RAS, Moscow, 117997, Russia, 3Keldysh Institute for Applied Mathematics, 119991Moscow, Russia, 4Max-Planck Institut fur Sonnensystemforschung., Katlenburg-Lindau, Germany, 5University of Arizona, Tuscon, AZ, USA.

Introduction: Comprehensive analysis of data retrieved by the Descent Imaging Spectral Radiometer model to equilibrate, with resulting distribution in size (DISR)[1], a multichannel camera and spectrophotometer package operating in spectral range from 350 to 1700 nm, during the descent in the Titan atmosphere, suggests that the tholin haze is composed of particles with complex fractal structure and broad distribution of parameters [2]. The challenging aspect of modeling is the capture key microphysical processes that govern the formation of haze with given microphysical parameters. In order to evaluate those processes, a comprehensive microphysical model has been developed and the results were compared with DISR retrievals. Here we consider the most important processes, namely coagulation controlled by electric charging.

The model: A time-dependent, 1D model calculates the distribution of tholin particles in height and size. Vertical motion of haze material is controlled by sedimentation and eddy mixing, whereas their size distribution results from the balance of coagulation and removal in the vertical, that mainly concerns larger particles. As electric charge acquired by particles is believed to play a major role in shaping the tholin haze size distribution[4], the model interactively calculates distribution of particles in charge and simulates essential charge exchange processes, such as collisions with ions and electrons as well as photoelectric effect. Based on this quasi-3D distribution calculated on the mesh with resolution of 100x60x30 nodes, an effective coagulation kernel assuming Brownian and ballistic coagulation, is calculated and applied to the current size distribution, so the charge exchange due to coagulation is neglected compared to faster processes. The only input control parameters of the model remain the location, vertical extent and intensity of the assumed photochemical source, which was specified similar as in [3], and eddy diffusion coefficient. The latter was assumed to have monotonously increasing vertical profile and varied in a wide range from 3x10^2 to 10^6 cm^2/see everywhere except for the altitudes with strong thermal inversion around 80 km, where it fades by several orders of magnitude.

Charge distribution is calculated assuming the dynamic equilibrium between acquisition and loss of charge[5]. Photoelectric effect is calculated on the basis of the self-consistent UV radiative transfer calculations, taking into account the evolving haze distribution. As the charging conditions are dependent on the local time, solar angle in the model varies according to Titan’s sol equal to 16 days.

Results: Typically it takes about 100 yr for the model to equilibrate, with resulting distribution in size and height comparable to DISR retrievals. On the night side, the charge distribution is mainly controlled by ion and electron number density and the particles are charged negatively. For particles larger than 0.5 μm, mean particle charge is approximately proportional to its radius, as shown in Figure 1. This is consistent with assumption of the early models [6,3] and proves the validity of coagulation models assuming specified charges. However, for smaller particles the distribution reveals a plato, with domination of both neutral particles and those possessing a single electron. There is a strong indication that it is the elementary charge that provides threshold for Brownian coagulation of small particles possessing single electrons, and determines the effective monomer size of approximately 0.05 μm.

Figure 1. Typical nighttime mean charge vs. particle radius for different altitudes. Fractional values imply the presence of neutral population.
Figure 2. Time-average log charge distribution in (color map) and tholin particle density in size and height (contours).

Figure 2 demonstrates a typical equilibrated distribution of particles in all three model dimensions. In this simulation a photochemical layer is located at 275 km with vertical extent of 50 km. It is evident that in the lower atmosphere, largest clusters possess charge of hundreds electrons that effectively stops subsequent coagulation with like clusters. However, there is a significant resource of growth for such particles by accretion of fine fraction, which is either neutral or possess positive charge. As these particles sediment extremely slowly, the intensity of their transport to the lower atmosphere depends on the intensity of mixing.

Fading eddy mixing near 80 km results in the separation of population by the non-mixing inversion layer characterized by high static stability. This is why below the tropopause particle growth essentially stops, and the model generates size distribution relatively constant with height.

Conclusions: the self-consistent microphysical modeling of tholin haze suggests that charging is a key mechanism forming particle size distribution. As the charging parameters of tholin are poorly known, comparison of the modeling results with observations may shed additional light to the nature of this substance. Direct quantitative comparison may only be accomplished by means of the general circulation model.

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References:
JEZERO CRATER DELTAS: INSIGHTS FROM TERRESTRIAL ANALOGS. S. C. Schon, C. I. Fassett, and J. W. Head. Dept. of Geological Sciences, Brown University, Providence, RI, 02912 USA; samuel_schon@brown.edu.

Introduction: Fassett and Head [1] first identified deltaic deposits associated with two valley networks debouching into (recently named) Jezero crater in the Nili Fossae region. Their analysis constrained the development of the crater lake and concluded that deltaic deposition was primarily subaqueous. Recent high-resolution datasets from the Mars Reconnaissance Orbiter mission provide new insights into the depositional history of these deposits and support continued analysis and reevaluation of earlier interpretations [2]. In this contribution we focus on terrestrial analogs for fluviosedimentary architectures observed in the western delta that are indicative of stable baselevel. We also classify the delta system as a sediment-dominated, lobate, Gilbert-type delta based upon the relative influence of constructive and destructive processes as observed terrestrially.

Meanders, scroll bars, & point bar sequences: HiRISE data contain evidence of meander development and truncation within the stratigraphy of the delta. Meander cut-offs define the evolution of the meander belts, while dark sands fill swales --highlighting the scroll bars of the lateral accretion topography. Point bar sequences are well-characterized terrestrially as time-transgressive, laterally continuous deposits that are indicative of meandering systems. These sequences are prograding, diachronous, fining upward (channel lag → gravel → sand → mud) sequences which form at the inner bank of meanders; see Figure 1. During flood stages, point bar sequences may be inundated and scoured forming chute bars and clay plugs.

Channel-levee deposits: In contrast to braided fluvial systems, stable meandering systems can develop natural levees. These wedge-shaped ridge deposits confine active channels and are composed of coarser-grained sediments that accrete vertically during flood episodes. These deposits are less prone to compaction than finer grained floodplain deposits terrestrially. Multiple hypotheses for the formation of the delta’s upper units have been proposed [2]; channel-levee deposits’ enhanced resistance to eolian deflation may be responsible for the preferential preservation of this material.

Delta Classification: Terrestrial deltas are commonly characterized based upon several non-exclusive parameters. Tidal energy, wave energy, sediment supply, longshore drift, and structural control are all influential in determining delta architectures. These factors can be considered primarily fluvial (sediment supply and composition) [constructive] or marine (waves, tides, and longshore drift) [destructive] processes. Therefore, ternary diagrams are useful representations of the relative influence of these factors; see Figure 3. In Jezero where meander belt formation indicates stable baselevel during at least significant periods of deposition, sediment supply and composition are interpreted to be dominant factors controlling the delta’s architecture and lobate/elongate planform.

Implications & Predictions: New datasets from MRO indicate that Jezero crater was a basin with significant deltaic deposition during a period(s) of stable baselevel. Lower in the stratigraphy are lateral accretion deposits indicative of subaerial deposition in a delta plain setting, while higher in the stratigraphy channel-levee deposits may be preserved [2]. Based on terrestrial analogs, these associations indicate a stable baselevel, general coarsening upward trend in the delta deposit, significant eolian deflation of the original delta surface, and backwasting of the delta front.

Figure 3: A ternary diagram classification of delta morphology illustrates dominant controls and the continuums of morphology that exist between endmembers. Preliminary analyses indicate that Jezero crater delta deposits are lobate to elongate and formed in a fluviually-dominated environment characterized by baselevel stability (after [6,7]).
LAYERED MORPHOLOGY OF THE LATITUDE-DEPENDENT MANTLE. S. C. Schon1, J. W. Head1 and R. E. Milliken2. 1Dept. of Geological Sciences, Brown University, Providence, RI, 02912 USA; samuel_schon@brown.edu. 2Jet Propulsion Lab, 4800 Oak Grove Dr., Pasadena, CA 91001 USA.

Introduction: Recent high-resolution imaging has confirmed earlier observations of latitude-dependent morphologies associated with ice-rich mantling deposits. Using Mars Orbiter Laser Altimeter (MOLA) data, [1] documented systematic latitudinal variations in surface roughness. Within the mid- to high- latitude region of both hemispheres, higher latitude terrains were found to be smoother at short baselines. Using Mars Orbiter Camera (MOC) images, [2,3,4] presented morphological observations of young surface textures ranging from smooth and continuous to highly degraded, viscous flow features, and gullies which they interpreted as consistent with the recent emplacement of ice-cemented loess undergoing desiccation/degradation. Numerous analyses (e.g., [5,6]) have supported the stability of near-surface ground ice in this latitude regime in addition to gamma-ray spectroscopy results of abundant hydrogen (e.g., [7,8]). Stratigraphic analysis of layering within mantle deposits is a means of assessing formation hypotheses such as vapor diffusion (e.g., [9]) and airfall deposition that may be correlative with geologically recent obliquity perturbations; see Figure 1.

![Figure 1](image1.png)

**Figure 1:** A) Schematic illustrating the difference between strictly vapor diffusion (left) and obliquity-driven surface emplacement of mantling materials (right). B) High obliquity expands the zone of mantle stability and mobilizes volatiles from higher latitude reservoirs for mantle deposition.

Observations: Layering within the mantle is observed symmetrically within the mid-latitudes of both hemispheres, but outcrops are more numerous in the southern hemisphere where the topographic variability of the basement more frequently generates favorable slopes with orientations that expose layering. Generally, mantle surface texture varies from smooth and continuous at higher latitudes to discontinuous, sublimation pitted, degraded mantle textures at lower mid-latitudes.

**Latitude bands.** Layering outcrops are concentrated in the transitional zone between these textures (~35°-40°); most MOC and HiRISE images in which layering was observed individually contain both smooth and degraded mantle textures. The global MOC catalog of ~13,000 images compiled by [3] contains 101 southern hemisphere images interpreted as mantle layering outcrops located between 28.9°S and 49.44°S with a median of 37.9°S (x=37.9°S). This latitudinal range is commensurate with a band of strong slope asymmetry attributed to obliquity-controlled insolation geometry that favored downslope movement on pole-facing slopes [10] as well as the occurrence of young gullies [11]. All available HiRISE images between 20°S and 50°S were also analyzed: 42 of 210 images contained evidence of mantle layering with a median of 37.9°S (x=37.8°S).

![Figure 2](image2.png)

**Figure 2:** This portion of MOC M0402834 (40.5°S, 190.4°W) illustrates asymmetric mantle textures associated with a (~340m) crater: A) Smooth equator-facing crater rim, B) Degraded pole-facing crater interior; C) Smooth equator-facing crater interior, D) Degraded pole-facing crater rim.

Slope and Orientation. Mantled craters, valleys, and scarps are common in this subset because of their association with slopes that are advantageous for observing layering in cross-section. Smooth mantle textures are observed preferentially on equator-facing

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1 Stereo pairs are considered a single observation for statistical purposes.
slopes, while degraded mantle textures exhibit a preference for pole-facing slopes. Asymmetrically mantled craters illustrate this phenomenon and are common in the latitudinal band between smooth and degraded textures where layering is most commonly observed: smooth mantle morphology dominates the equator-facing interior wall, while the pole-facing interior wall is degraded; see Figure 2.

Commonly crater walls are too steep and altered (e.g., gullies, slumps, viscous flow features) to be conducive to mantle layering outcrop exposure, but pole-facing slopes from raised crater rims are frequently observed locations for outcrops of layered mantle units. This insolation control of preserved mantle texture is highlighted by observations where a pole-facing, steep slope is dominated by degraded texture, but benches of much gentler slope have smooth mantle texture.

Layer Morphology. Individual layers are interpreted to be of relatively uniform and consistent thickness (on the order of several to tens of meters). Cross-bedding relationships are not observed and correlation of layered units is possible within a single MOC image when surface features and outcrop geometries are similar (e.g., M1900888). Therefore, layers are interpreted to be of wide aerial extent. Approximately tripartite layering with homogeneous lithologic facies (degraded textures) is the most common style of layering observed in outcrop; see Figure 3.

Figure 3: This portion of MOC M0200544 (38.84°S, 112.05°W) contains an outcrop of mantle layering associated with a small TROUGH south of Claritas Fossae; MOLA track for reference.

However, finer layering is observed on more gentle slopes where ≥8 individual layers are sometimes distinguishable; see Figure 4. These homogeneous units are discernable by slight variations in texture and albedo near the limit of resolution in MOC data. Within the sequence, smooth textured lower albedo surfaces separate higher albedo, raised relief, blocky and segmented surfaces. These blocky segments are interpreted as semi-consolidated dust-rich lag deposits, which are more resistant and provide the strength to support greater relief.

Figure 4: This portion of MOC M0204329 (38.01°S, 113.59°W) contains an outcrop of fine mantle layering associated with a gentle pole-facing slope.

Implications: Layered mantle outcrops provide genetic evidence of syndepositional layering interpreted to be the result of cyclical deposition of an ice-rich eolian dust material, which is inconsistent with a strictly vapor diffusion model of high latitude terrain softening. Individual units are hypothesized to represent geologically recent obliquity excursions. In this model, obliquity excursions lead to the mobilization of volatiles that are deposited in mid-latitude regions during the return phase of an excursion. This deposition is expected to be most ice-rich initially then become increasingly dust-rich, leading to a reverse grading pattern in these units. Therefore, one smooth texture low albedo unit and one blocky high albedo unit together could be the result of one obliquity excursion, but more work remains to explore the relationship between obliquity variation and mantle stratigraphy.

Observation of the north-south asymmetry of polarization: As known, ground-based and cosmic polarimetric observations of Jupiter in visual spectrum range show the dependence of linear polarization degree P on phase angle and polarization increasing with latitude (even at zero orbital phase angle): polarization degree increases from zero (equatorial regions) to 7-8% (polar regions). Also it is known, that there is a north-south asymmetry of linear polarization at Jupiter [1-4]. To explain these observational facts, we have started regular polarimetric observations of Jupiter in 1981. On the basis of our Jupiter photopolarimetric observations in opposition at blue light during 1981-1999, seasonal variations of north-south asymmetry (PN-PS) of linear polarization P in polar regions and anticorrelation between PN-PS and insolation have been found [1]. Parameter of asymmetry PN-PS is defined as a difference between values of linear polarization degree on north and south at the latitudes ±60° at the central meridian. PN-PS data are well organized if plotted in accordance with Jupiter’s orbital location and there is some relation between PN-PS and insolation [1]. We are continuing our studying: 1) our new observations were used; 2) our old data (1981-1998) have been reprocessed using new improved technique; 3) Hall and Riley data [2] (1968-1974) are involved for analysis. New variant of P-asymmetry dependences on Jupiter’s orbital location are presented in the Fig.1. To investigate the nature of the dependence the approximations have been made using different functions. The best approximation was shown by one-periodic function (Fig.1, dashed line).

Correlation coefficient between PN-PS and IS/IS is -0.7, i.e. there is significant anticorrelation. Our earlier assumption [1] is confirmed by new data and we can speak about seasonal variations of polarization.

Causes of seasonal variations: As known, data of polarimetric observations in visible, infrared and ultraviolet range are sensitive to presence of stratospheric aerosols’ haze in Jovian atmosphere, observed at top levels of Jupiter stratosphere at high latitudes (on p ~ 0.1-1 mbar pressure level) [4]. We suppose that the main cause of these effects is change of aerosol concentration. Aerosols may be unstable, and temperature changes may influence upon generation and destruction of aerosol particles. According to [5], the observed aerosol haze is consists of benzene and polycyclic aromatic hydrocarbons (PAH) like naphthalene, phenanthrene, pyrene.

Thus, possible scenario of appearance of north-south asymmetry of polarization is: seasonal variations of insolation \( \rightarrow \) seasonal variations of temperature \( \rightarrow \) changes of activity of aerosol generation \( \rightarrow \) aerosol concentration changes \( \rightarrow \) polarization changes \( \rightarrow \) changes of north-south asymmetry of linear polarization.

Temperature effect on aerosol haze formation: Average temperature in polar regions of Jovian stratosphere is about 150 K [6]. This temperature is lower than triple points of naphthalene and benzene (359 K and 278 K, respectively [7]), so they may produce crystal nucleus from gaseous phase. Let’s consider homogeneous particle nucleation. Such process proceeds without additional condensation centers. Equilibrium condition for nuclei of a crystal with radius \( r \) and surrounding gas is defined as follows [8]:

\[
    r = r_c = \frac{2\alpha\Omega}{\Delta\mu(T, \xi)} ,
\]

where \( r_c \) - critical radius (nuclei with smaller radius evaporate, and bigger ones grows); \( \Omega \) is specific volume of molecule in crystal; \( \Delta\mu = kT\xi \) is chem-
Mechanism of the effect of solar cosmic rays on aerosol haze: Earlier (Fig. 3 in [3]) we have studied the relation between parameters of solar activity and polarization of polar regions, and possibility the influence of high energy protons (solar cosmic rays) on polarization values for years 1998, 2000 and 2001 has been found. So, the first, high-energy protons may increase concentration of ions that participate in chemical reactions, which can enhance synthesis of source material (PAH molecules) for aerosol formation. Second, the ions may serve as additional condensation centers of aerosols. At last, chemical reactions stimulated by additional ionization of the atmosphere occur with heat release or absorption, which may result in temperature change at high altitudes (similar effect is well known for the Earth stratosphere). This can change aerosol concentrations and, consequently, polarization values at both poles. Because of nonlinear dependence of vaporization-condensation processes upon temperature, the stratosphere aerosol concentration is different in both polar regions, which may produce polarization asymmetry. Only second mechanism (nucleation in gas containing ions) can be described quantitatively. Gibbs potential in this case is:

$$\Delta G(\xi, q, r) = \frac{4}{3} \pi r^3 \frac{\Delta \mu(T, \xi)}{\Omega} + 4 \alpha \pi r^2,$$

where $G$ is Gibbs potential. Homogeneous nucleation takes place when radius of critical nucleus is close to molecular sizes; at the same time, supersaturation $\xi$ is about or larger than 1.

To study the effect of temperature changes on PAH formation we have used altitude concentration profiles from [5]. Season changes of temperature in north and south Jupiter polar regions amount to ±30 K [9].

Mechanism of the effect of solar cosmic rays on aerosol haze: Earlier (Fig. 3 in [3]) we have studied the relation between parameters of solar activity and polarization of polar regions, and possibility the influence of high energy protons (solar cosmic rays) on polarization values for years 1998, 2000 and 2001 has been found. So, the first, high-energy protons may increase concentration of ions that participate in chemical reactions, which can enhance synthesis of source material (PAH molecules) for aerosol formation. Second, the ions may serve as additional condensation centers of aerosols. At last, chemical reactions stimulated by additional ionization of the atmosphere occur with heat release or absorption, which may result in temperature change at high altitudes (similar effect is well known for the Earth stratosphere). This can change aerosol concentrations and, consequently, polarization values at both poles. Because of nonlinear dependence of vaporization-condensation processes upon temperature, the stratosphere aerosol concentration is different in both polar regions, which may produce polarization asymmetry. Only second mechanism (nucleation in gas containing ions) can be described quantitatively. Gibbs potential in this case is:

$$\Delta G(\xi, q, r) = \frac{4}{3} \pi r^3 \frac{\Delta \mu(T, \xi)}{\Omega} + 4 \alpha \pi r^2,$$

where $r$ is the radius of charged sphere, $r^*$ is the ion radius; $q$ is ion charge, and $\varepsilon$ is dielectric permeability of nucleus. The new term on the right in (3) describes screening of a charge $q$ by growing particle. As shown our calculations ($\varepsilon = 2.3$, $r^* = 2$ A), additional domain of stability (local minimum of function $\Delta G(\xi, q, r)$) does not appear in the range of our interests (particles sizes ~ 1 μm) even for unrealistically great charges. At real values of charges (1-2 charge of electron), stability appears only very close to molecular sizes, i.e., only charged molecular clusters (not particles) can be stable (not evaporating and not growing). Thus, mechanism of aerosol particles formation on charges is not effective.

Conclusions: (1) There is a correlation between polarization asymmetry and insolation. (2) Seasonal variations of insolations (through variations of temperature) is the principal cause of variations of north-south asymmetry of polarization. (3) Jovian stratospheric haze which consists of crystal PAH (naphthalene, phenanthrene) particles may be formed by homogeneous nucleation. (4) Temperature variations in jovian stratosphere have strong influence on PAH condensation. (5) Flux of solar cosmic rays may effect upon concentration of aerosol only through series of chemical reactions that produce source material for aerosol formation.

TOPOGRAPHICAL SURVEY OF AITKEN CRATER ON THE FAR SIDE OF THE MOON

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It is young impact crater about 130 km in diameter. Aitken is located at central part of the far side of the Moon: longitude 173E, latitude 17S. The crater has an appreciable central peak (Fig.1).

Fig.1 The Aitken crater view: left - from distance about 1000 km, Zond-8, 1970; right - from distance about 100 km, Apollo-17, 1972.

The position of Aitken attracts an attention. Firstly-the crater is located near lunar equator. It means unlike the polar craters the illumination intensity and heating of the crater bottom achieves of maximum values during lunar day. Secondly – the crater is practically located on the outer wall of the “South Pole – Aitken” basin. Besides that Aitken crater region is the single lunar territory where the orbital topographical observations were made on a film and delivered to the Earth by Zond and Apollo missions.

Topographical data for Aitken crater region are presented at the Lunar Topographic Orthophotomap [1]. It is 2 sheets: Australis (south part of Aitken region) and Borealis (northern part). These data were obtained on the basis of Apollo images. The LTO scale is 1:250000 and contour interval 100 meters.

Lunar coordinate system created by Aleksashin’s group [2] allows to get some data on topography of Aitken crater also. This system was based on Zond-8 images and contains spatial coordinates for 881 points. Part of them are belong to Aitken crater region.

Height profiles of the crater are shown on Fig.2. The elevation numbers are given relatively level of 1730.0 according to LTO system. Both profiles are the sections over the crater center. The first profile gives the height changes in the lunar meridian direction 173E and the second one – along the lunar parallel of 17S.

The general elevation difference equals 5.8 km (profile 173E). Some points in the Aitken vicinity have the elevation numbers more than 11.2 km and the most low sites lie at the level 4.0 km. One point (in Haviside crater direction) achieves 14010 meters! Thus the maximum elevation difference in Aitken region equals more than 7.0 km.

Fig.2 Aitken height profiles.

The task of the joint processing of the Zond and Apollo images appears rather attractive. These images supplement to each other successfully and in our opinion could get new additional topographical data for Aitken region. Moreover the joint photogrammetry analysis of mentioned images could facilitate the decision of the problem of creation the uniform coordinate system on the Moon for near and far side.

We have started revision the Zond and Apollo observation. At present we busy with the images measurements and the construction of single stereomodels.

Some problems takes much more time than we expected. For example usual household personal computers are not suitable for processing of large graphical images and we have forced to devide the image in several parts. We have no full data on Apollo images (cameras calibration data). Certainly it constrains the job in general.

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TO THE PROBLEM OF PREARCHEAN HISTORY OF THE EARTH (ON THE
BASIS OF COMPARATIVE PLANTTOLOGY DATA). V.I. Sirotin, Voronezh State
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The most ancient time interval in the Earth's history (4.55-3.85 mil. years ago) has not been studied in details. That's why it is sometimes called a “dark” one. However recent achievements in comparative planetology allow to distinguish the following stages in the early history of the Earth:

1. The “hot Earth” stage: 4.55-4.45 mil. years; in the course of the Earth's mass gaining approximately by 99.9 % it was differentiated into the core, the mantle and the primary earth's crust. The following scenario of origin and formation of the earth's crust is the most probable. According to greater gravity on the Earth (compared to the Moon) and during crystallization differentiation pyroxene and olivine gained great amount of anortite in sinking thus excluding the formation of “anortozitic” (as on the Moon) crust. It is quite possible that differentiation of magnetic ocean, poor in iron, magnesium (bonded in olivine and pyroxene) and calcium (bonded in anortite), led to formation of more “acidic” – “granitic” layer some kilometers thick when cooling [3]. This “acidic crust” was underlain by magmatic layers corresponding in composition to basalts and gabbro. In its turn, this layer of magma was blocked by the layer of eclogite, formed due to sinking of pyroxene crystals and crystallization of garnets of almandine row. Formation of planetary eclogitic layer blocked the growth of the earth’s crust thickness [3,6]. The primary crust has not survived on the Earth. Thick carbonaceous – vapour atmosphere formed simultaneously with primary crust (with the admixture of methane, ammonium, hydrogen etc) its pressure being over 350 atm (water vapor formed 300 atm, CO₂ – 45 atm, HCl and other volcanic gases – 10 atm). The reaction of such atmosphere with the newly formed earth’s crust was violent [1] and it led to formation of primary mass of sedimentary rocks in the form of planetary crust of weathering [2] that contains a great number of clayey minerals. By the end of the stage (by 4.45 mil. years ago) organic prebiological structures could have originated. Their origin could be facilitated by strong thunder storms in the earth’s atmosphere, while their conservation and evolution – by adsorption in beddings of clayey smectite minerals. Tectonic style of the Earth was characterized by “herds ” of small lithospheric plates, plume tectonics, obduction, tessera formation and possibly conception of rudiments of sialic crust in composition of lithospheric plates. Tension of the Earth’s tectonic style was aggravated by gravity in the system Earth – Moon, (according to impact mechanism formation. The Moon was in the Rocher zone) [2, 5, 6].

2. The stage of considerably cooled “warm”, “cool” Earth: 4,45 – 4,20 mil years ago. Many scientists call this interval Prearchean Era. The upper boundary of this interval is determined according to the age of the beginning of the planets’ bombardment by large meteorite-asteroid bodies – according to the age of the most ancient impact basin Nectaris on the Moon – 4,20 mil. year. The stage is characterized by the absence of intensive bombardment. However the stage was not calm because of strong earth’s volcanism, that was supported by the influence of the Moon’s proximity [5,6]. By 4,45 mil.years the primary anortozite crust of the Moon was completely formed while KREEP association that may be considered as the final act of planetary magmatic system crystallization was formed by 4,35 mil.years. Taking into account the Earth’s mass and its considerable energy potentiality it may be supposed that primary crust was of great heterogeneity, while the Earth’s variant of KREEP association may
be considered as further development of sialic crust [6]. Theoretic calculations show that if the mantle differentiation had occurred completely the thickness of the earth’s crust would have been about 200 km [6] while it is up to 20 km in all earth type planets. It means that either there was no entire magmatic ocean on the Earth, or the crust substance repeatedly returned into the mantle in the result of obduction, sagduction and subduction during this stage.

3. The stage of intensive bombardment of the Earth and other planets by large asteroid – meteorite bodies: 4,20-3,85 mil. years. The upper boundary of the stage is dated by the age of Kaloris basin on Mercury. Cratered surfaces of the Moon and Mercury were formed during this stage. The ideas of “dark history” of the Earth given above are also confirmed by the data on the Earth’s oxygen isotopy (by zircons) and seismic holography [4,6], that testify to early origin of the Earth’s hydrosphere and existence of convergent zones extending up to the earth’s core as well as to the circulation of substance in the Earth’s interior. Study of komatiites of Archean greenstone belts point to the great role of mantle-plume and subduction processes in formation of the early continental earth crust. The Earth “has invented” its own mechanism of extra heat release [6]; which is characterized commonly by three geodynamic environments: 1) oceanic lithosphere that is constantly recycled (in spreading and subduction zones); 2) continental lithosphere, floating continents that are constantly destructed and restored, but always remain at the surface; that is why the Earth's crust could not be thicker. As soon as the thickness of the crust is 60 km, crust minerals with low density transform into thicker ones; 3) interacting continents in the collision zones where they are subjected to the action of lithospheric subcrustal mantle and adjacent plates. Thus, continents are constantly rejuvenated due to compression and uplifting of substance in the collision zones or in the result of island arc elimination due to their joining to the continent (in the subduction zones) or in the result of intraplate eruption of basalts on or under continental plate, or in the result of intrusion into it. It is reasonable to suppose that all geodynamic processes of the modern Earth originated in its early history.

SELF-GRAVITY AND RHEOLOGY OF SMALL SOLAR SYSTEM BODIES. E. N. Slyuta, Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow, Russia. slyuta@mail.ru.

Introduction. Estimation of the transition parameters between the small and the planetary bodies with well known “hydrostatic pressure” equation [1, 2] may be correct only in a case of viscous liquid rheology and in a case of strengthless material. It is necessary to take into account also, that hydrostatic pressure alone does not produce plastic deformation. An analysis of rheology properties of Solar system bodies has been carried out with a rheology model [3,4], which uses the elastic theory with ultimate strength for a three-dimensional self-gravity body, and allows receiving the exact solution of differential stresses in a solid elastic body and to carry out their analysis.

Icy and rocky bodies. An obtained yield strength for ice pair the Hyperion (small body) – Mimas (planetary body) is within 0.38<$\sigma$<$1.4$ MPa [4]. Experimental data for pure water ice is 0.1<$\sigma$<$2$ MPa at 203°K temperature for the upper limit [4]. For rocky pair a Pallas (small body) - Ceres (planetary body) this range is within 30.3<$\sigma$<$60.2$ MPa [4, 5]. To reaching of a threshold value of the stress deviator and transition to the category of planetary bodies the mass of Vesta asteroid should be larger at least by order of magnitude [4]. For L-5 ordinary chondrite (Tsarev meteorite) the critical radius is estimated to be 756 km [4]. For comparison, for terrestrial basic and ultrabasic rock depending on its rheology properties the critical radius settles down in a range 582<$R_c$<1038$ km [4]. As the yield strength depends not only on structure, but on temperature also, a body critical radius varies accordingly. For example, the critical radius of a metallic body similar Sikhote-Alin iron meteorite, at 200°K temperature (asteroid belt) is estimated to be 259 km, at 77°K (Saturn system) - 337 km and at 4.2°K (deep space) - 389 km [4]. Ceres composition on the rheology properties obviously differs from the considered examples and, probably, corresponds to carbonaceous chondrites [4, 5] with lower strength properties in comparison with other rocky bodies.

Cometary nuclei: Review of the observed, analytical and experimental data on strength properties of cometary nuclei shows that a cometary material tensile strength is about 2×10^4$ dyn cm^-2 [6]. The stress deviators obtained for some cometary nuclei are small and are lower by two orders of magnitude than the cometary material tensile strength [7]. Cometary nucleus size at which the stress deviator is equal to the cometary material tensile strength (2×10^4$ dyn cm^-2), is estimated to be 54 km [7]. That is Hale-Bopp cometary nucleus size within its size estimation uncertainty [8]. It means, that cometary nuclei less than 54 km in size (it is practically all known comets because Hale-Bopp is the largest comet excluding 2060 Chiron) have a constant tensile strength of about 2×10^4$ dyn cm^-2, which is determined by structure only. Effective tensile strength of the bodies more than 54 km in size depends on a body mass and shape parameters and increases under the square-law depending on a body size [4] (at constant density) (Fig. 1). Such increase of the tensile strength can explain a lack of cometary nuclei more than 54 km in size (size gap). At least, such dependence would be influenced on a secondary population, which is formed as a result of a destruction of parent bodies.

When self-gravity dominates a tensile strength (i.e. small bodies of >54 km in size), fracture starts at the surface and the object erodes inward [9], while in small bodies of <54 km in size fracture begins in the center [9, 10]. Similar character of destructions is one of the strong conclusions of critical mass theory (CMT) [4] and it depends not only on mass, but on shape parameters also. Though a radial gradient of the stress deviator is insignificant [4], it is present, and at shape parameters of $a/c$<1.75 the stress deviator on the body surface is more than in the center (Fig. 2). Hence, any elastic or plastic deformation will develop from a surface to the center of a body. At the shape parameters of $a/c$>1.75 a radial gradient of the stress deviator varies on opposite and the mechanism of destruction of a body will be the same, as well as for bodies of <54 km in size.

Phoebe. If comets are small bodies with an irregular shape, Phoebe is enough large body with a sphere-like shape. The mean radius of the satellite is 106.6 km [11]. As the satellite has very low albedo (0.06) [12] was considered, that Phoebe is a rocky body. But among small rocky bodies Phoebe differed by anomalous shape parameters [4]. An explanation which followed from CMT, consist that Phoebe has another composition than rocky and even icy bodies. According to the last data, Phoebe’s composition is like to that of Kuiper Belt objects [13]. The orbital properties of Phoebe are allowed to suggest that it was captured by the Saturn’s gravitational field. On the shape parameters Phoebe belongs to planetary bodies [14]. It means that the minimal stress deviator in a case with Phoebe exceeds yield strength of a material and gravitational deformation has already taken place. Gravitational deformation is accompanied by gravitational densification and gravitational strengthening of a material at the entire body due to volumetric gravitational compression accompanied by two basic mechanisms of plastic deformation as minimum [4, 14].

Cometary nuclei are characterized by high porosity and low density [9, 15-18]. The maximum mean density of a fully packed cometary nucleus would be ≈ 1.65 g cm^-3 [18]. Phoebe’s density is 1.63 g cm^-3 [11] and corresponds to the above-stated value for a fully packed cometary nucleus. The stress deviator caused in shape parameters and mass of Phoebe is 0.9 MPa [7]. Hence, the yield strength of a material of Kuiper objects to which comets and Phoebe belongs [13, 19, 20], is within the range 0.002<$\sigma$<0.9$ MPa, where the minimum value corresponds to a cometary nucleus tensile strength. It is necessary to note, that if to take a yield strength of 0.9 MPa, the mean
radius of a small body not exposed to gravitational deformation (i.e. with 300 kg m$^{-3}$ density), would reach 570 km.

**Conclusion.** Kuiper Belt objects are characterized by the lowest value of yield strength among Solar system bodies (Fig 3). It follows from a composition, of course. In comparison with water ice (about 30%) the share of exotic ices (CO, CO$_2$, CH$_3$OH, CH$_4$, H$_2$CO and others) in cometary nuclei is rather small (about 12%) [21]. Thus, obtained data on rheology of Kuiper objects shows that even the subordinated amount of exotic ices can result in essential change of rheology properties of material. It is to be noted, that the strength properties of solid Solar system objects in depending on their composition vary within extremely broad range – from 0.002 up to 350 MPa (Fig. 3). Such dependence of transition parameters on composition can serve as a good indicator of distinction in composition at remote researches of numerous Kuiper and transplutonian objects. As example of it the case with Phoebe can serve.

![Fig. 1. Dependence of a tensile strength on size of a cometary nucleus (at constant density). Tensile strength of cometary nuclei of <27 km in radius is constant (2×10$^4$ dyn cm$^{-2}$) and is determined by a structure only. Tensile strength of cometary nuclei of >27 km in radius is determined by a body mass and shape parameters according to the stress deviator equation [4].](image1)

![Fig. 2. Dependence of a radial gradient of the stress deviator on shape parameters. Radial gradient is determined by ratio of stress deviators in the center and on the surface of a body [4]. $\sigma^c$ - a stress deviator in the center; $\sigma^s$ - a stress deviator on the surface; $a$ - the largest semi axis of a body; $c$ - the smallest semi axis of a body.](image2)

![Fig. 3. Dependence of a transitional critical radius of Solar system bodies on yield strength of a material on the observed and experimental data. [Kuiper Belt objects][Hyperion–Mimas][Pallas – Ceres][L-5 ordinary chondrite][Terrestrial basic and ultrabasic rocks][Meteoritic iron, 300°K][Meteoritic iron, 77°K][Meteoritic iron, 4.2°K]. Five basic groups of objects depending on rheology properties are allocated. The first (I) is Kuiper Belt objects which are characterized by the lowest yield strength. Ice bodies parameters (Hyperion–Mimas) form independent, separate group (II) and are not crossed with ones of Kuiper objects. The third group (III) is rocky bodies which are limited to Ceres parameters [4]. Probably, that is a field of carbonaceous chondrites [4] and similar rocks characterized by week mechanical properties. The fourth group (IV) includes L-5 ordinary chondrite and material similar to terrestrial basic and ultrabasic rocks [4]. These rocky objects are characterized by much stronger mechanical properties. The fifth (V) is a field of the metal bodies (meteoritic iron) at temperatures of 300°K, 77°K and 4.2°K. These objects are differd by a high density and high strength properties [4].](image3)

Introduction. The main feature of atmosphereless small and planetary bodies, including the Moon, is a regolith layer generated as a result of meteoric bombardment during all geological history of these bodies, i.e. or from the body formation (asteroids, small satellites, comets), or from the underlying bedrocks (formation of the Moon and other planetary bodies). Stratified core sampling of lunar regolith and underlying bedrocks is actually all geological history of the Moon limited to age of underlying bedrocks. For lunar highlands it may be more than 4.0 b.y.

The main project scientific tasks.
1. Research of a lunar heat-flow lateral distribution on the Moon near and far side.
2. Research of the Moon interior structure and distinctions in the lunar crust structure on the Moon near and far side. Research of a structure and thickness of the lunar crust top layers - megaregolith and regolith. Research of the megaregolith structure within mascons areas.
3. Research of distinction in lunar mare basalts compositions on the Moon near and far side. Research of composition and distribution of lunar mare rocks of different ages – Upper Imbrian series (3.8-3.2 billion years), Eratosthenian system (3.2-1.1 billion years) and Copernican system (1.1 billion years and is younger), and of different spectral classes (from low-titanium up to high-titanium basalts). Research of volcanic rocks within volcanic provinces at Oceanus Procellarum.
4. Research of composition and age of lunar highland rocks of the Pre-Nectarian (> 3.92 billion years) and Nectarian (3.92-3.85 billion years) systems. Search of the oldest lunar highland rocks (> 4.0 billion years).
5. Research of physical and mechanical properties of lunar regolith and rocks in situ.
6. Search in lunar highlands of terrestrial meteorites which have arrived from the Earth more than 4 billion years ago during formation of the largest impact basins on the Moon and, accordingly, on the Earth.
7. Correction of instrumental (seismic prospecting and electrical survey), remote (radar-location) and statistical (the crater morphology analysis and crater population characteristics) methods of regolith thickness determination in a direct measurement point of one (a basic network borehole).
8. Testing on basic network of well-logging boreholes of any perspective remote methods of an estimation of any necessary elements abundance in lunar regolith.

The project description. According to the basic scientific problems the project provides creation of a basic network of well-logging core boreholes and long term monitoring automatic mini stations on the Moon. Creation of the boreholes basic network is made with two types of automatic drilling rigs - stationary (mainly on the Moon far side) and self-propelled (mainly on the Moon near side). The basic network of boreholes can be created stage by stage within several or more years and should include from 3 up to 10 boreholes on the Moon near and far side. Each borehole of stationary drilling rigs and one of boreholes of each self-propelled drilling rig should be equipped with long-term automatic scientific station. The stratified core of lunar rocks from a borehole is delivered to the Earth. Borehole depth depends on regolith thickness and in lunar maria one should be 10-15 m, in lunar highlands - up to 30 m.

Lon-term scientific mini station. The scientific station should include the following basic components:
1. A stationary temperature probe for long-term measurements of a lunar heat-flow. The probe falls to a borehole after all works on drilling and remains there. Gauges of temperature measurements in both lunar maria and highlands are established on depth of 0.3-0.5 m, 1.5 m, 3 m and on the maximal depth of a borehole.
2. The temperature gauge on a lunar surface.
3. A seismometer for long-term monitoring and registration of a seismic activity caused or with falling of meteorites and endogenic activity of the Moon, or with carrying out of active seismic experiments.
4. A survey videocamera for research of local geomorphology, change of light exposure, and so forth.
5. A compact angular reflector for a laser location from the Earth for measurement of exact distance and coordinates of the stations on the Moon near side.
6. The central power station.
7. Radio equipment for communication with an orbital lunar satellite and with the Earth.

The complete set of the scientific equipment depending on additional scientific problems can be expanded.

The main scientific tasks of the automatic stationary drilling rig:
1. Drilling a borehole with core sampling and well-logging researches on depth up to 15 m in the lunar maria and up to 30 m in the lunar highlands (Table 1).
2. Deploy in the borehole of the temperature probe.
3. Deploy on a lunar surface of seismic station and other scientific and service equipment.
4. Delivery of lunar rock samples to the Earth.

<table>
<thead>
<tr>
<th>Type</th>
<th>Semi-automatic with remote command - program management of disposable action.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Way of drilling</td>
<td>Rotary on regolith, percuss-ion-rotary on solid rocks</td>
</tr>
<tr>
<td>Technology of drilling</td>
<td>Short runs with mechanical clearing of bottomhole</td>
</tr>
<tr>
<td>Depth of drilling and logging</td>
<td>10-15 m</td>
</tr>
<tr>
<td>Borehole diameter</td>
<td>60 mm</td>
</tr>
<tr>
<td>Sampling core diameter</td>
<td>41 mm</td>
</tr>
<tr>
<td>Logging probe diameter</td>
<td>40 mm</td>
</tr>
<tr>
<td>Run length</td>
<td>300 mm</td>
</tr>
<tr>
<td>Technological operating time</td>
<td>160 hours</td>
</tr>
<tr>
<td>Dimensions of drilling rig</td>
<td>4000×625×440 mm</td>
</tr>
<tr>
<td>Weight</td>
<td>243.5 kg</td>
</tr>
</tbody>
</table>

"LB10 drilling rig is an improved version of LB09 drilling rig which has successfully fulfilled at station "Luna-24". The perspective version of LB-10 well-logging drilling rig essentially differs from the basic version and allowing drilling up to 30 m.

The main project scientific tasks.

1. Drilling a borehole with core sampling and well-logging researches on depth up to 15 m in the lunar maria and up to 30 m in the lunar highlands.
2. Deploy in the borehole of the temperature probe.
3. Deploy on a lunar surface of seismic station and other scientific and service equipment.
4. Delivery of lunar rock samples to the Earth.

Table 1. The characteristic of LB-10 stationary well-logging drilling rig.
The main tasks of the automatic self-propelled drilling rig (Base version):
1. Performance of a geological route in length up to 400 km in preset area (Table 2).
2. Well-logging drilling up to three boreholes with core sampling on depth up to 15 m.
3. Deploy in one of the boreholes of a temperature probe.
4. Deploy of seismic station and other equipment.
5. Documented sampling (the image before and after sampling and coordinates) of regolith and fragments of lunar rocks with a manipulator.
6. Deploy and undermining of explosives on a route under the set program with the purpose of carrying out of active seismic experiment.
7. Documented sampling of weakly bound inert gases in regolith on a route with mechanical loosening of lunar regolith on depth up to 20-30 cm (for example, with a screw) under sealed dome and definition of their isotopic composition with an onboard mass spectrometer.
8. Research of a magnetic field on the all route length.
9. Docking with automatic spacecraft returned to the Earth (an automatic cargo rocket, etc.) in a final point of a route and transfer on its board of the collected lunar rock and gas samples.

Reconnaissance version differs from the base version by the drilling only one deep borehole up to 15 (in lunar maria) or up to 30 m (in lunar highlands) and additional well-logging drilling several (up to 5-6) shallow (up to 3-5 m) boreholes with core sampling on the route.

Table 2. The characteristic of self-propelled automatic well-logging drilling rig on the automatic moon rover basis (AL).

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time of active functioning</td>
<td>0.5-1 year</td>
</tr>
<tr>
<td>Distance of a movement - not less</td>
<td>400 km</td>
</tr>
<tr>
<td>The maximal speed of movement</td>
<td>5 km h⁻¹</td>
</tr>
<tr>
<td>Average speed of movement</td>
<td>2-3 km h⁻¹</td>
</tr>
<tr>
<td>It is long overcoming slope</td>
<td>25°</td>
</tr>
<tr>
<td>Angle of longitudinal dynamic stability</td>
<td>32°</td>
</tr>
<tr>
<td>Angle of cross dynamic stability</td>
<td>38°</td>
</tr>
<tr>
<td>Overcoming height of a threshold obstacle</td>
<td>250 mm</td>
</tr>
<tr>
<td>Number of videocameras</td>
<td>5</td>
</tr>
<tr>
<td>Range radius of the manipulator</td>
<td>3 m</td>
</tr>
<tr>
<td>Gross weight of the scientific equipment</td>
<td>400-450 kg</td>
</tr>
<tr>
<td>Dimensions of AL with drilling rig</td>
<td>5260×2440×2480</td>
</tr>
<tr>
<td>Gross weight of AL with drilling rig</td>
<td>2000 kg</td>
</tr>
</tbody>
</table>

The main tasks of well-logging researches. Researches of the basic geophysical properties of lunar rocks, which can be measured only in situ, are made with well-logging device during drilling:
1. Gamma-gamma logging for measurement of regolith density. In lunar maria it is measured in 4 points - on depth of 0.3 m, 1-1.5 m, 3 m and on the maximal depth of a borehole, and in lunar highlands - on depth of 0.3 m, 1-1.5 m, 3 m, 10 m and on the maximal depth.
2. Magnetic logging for measurement of a magnetic field.
3. Electromagnetic well-logging for measurement of a magnetic susceptibility and dielectric permeability of lunar rocks.
4. Borehole spectrometry for measurement of the radioactive elements abundance.

The landing areas of stationary drilling rigs:
1. Mare lunar rocks in Tsiolkovsky crater on the Moon far side. Drilling on depth up to 15 m.
2. Mare lunar rocks in Mare Orientale. Drilling on depth up to 15 m.
3. The oldest highland lunar rocks of Pre-Nectarian system in the field of the South Pole - Aitken on the Moon far side. Preliminary coordinates of a landing place is 60°N and 180°W. Drilling on depth up to 30 m.
4. Mare lunar rocks in Mare Moscow on the Moon far side. Drilling on depth up to 15 m.
5. The oldest highland lunar rocks of Pre-Nectarian system outside borders of impact basins on the Moon far side in area with coordinates: 5°S - 5°N, 175°W - 175°E. Drilling on depth up to 30 m.
6. Reserve variant.

The landing areas of self-propelled drilling rigs:
1. Oceanus Procellarum on the Moon near side. Drilling up to 3 boreholes on depth up to 15 m (base version) or one boreholes on depth up to 15 m (reconnaissance version). A route length is about 400 km (approximately 13° on latitude). The preliminary beginning point coordinates of a route are 44°N and 62°W, the preliminary endpoint coordinates are 38°N and 50°W. The route crosses two age divisions of mare lunar rocks of Upper Imbrian system, a volcanic province rocks, mare lunar rocks of Eratosthenian system and mare lunar rocks of Copernican system. The route crosses mare lunar rocks of several spectral classes, including low-titanium and high-titanium basalts.
2. Mare Humboldtianum and lunar highland to the north from the Mare. Drilling up to 3 boreholes on depth up to 15 m (base version) or one boreholes by depth up to 30 m (reconnaissance version). A route length is about 400 km (approximately 13° on latitude). The preliminary beginning point coordinates of the route are 58°N and 80°E in Mare Humboldtianum, the preliminary endpoint coordinates are 70°N and 80°E in lunar highland. The basic part of a route passes on area of distribution of Nectarian system highland lunar rocks.
3. Mare Imbrium of the Moon near side. Drilling up to 3 boreholes on depth up to 15 m (base version) or one boreholes on depth up to 15 m (reconnaissance version). A route length is about 400 km. The preliminary beginning point coordinates of a route are 24°N and 15°W, the preliminary endpoint coordinates are 34°N and 25°W. The route repeatedly crosses a minimum two age divisions of mare lunar rocks of Upper Imbrian system, mare lunar rocks of Eratosthenian system and area of ejected material from Lambert crater. The route crosses mare lunar rocks of several spectral classes.
4. Reserve variant.

Conclusion. According to the basic scientific problems the completed project assumes creation of 8 well-logging boreholes and stationary automatic scientific stations of long-term monitoring. Creation of a basic network of boreholes and functioning of scientific stations is provided only with automatic systems. The project concept takes into account experience and productivity of all previous lunar programs, especially the "Apollo" program, "Lunokhod" program and successful experience on a lunar soil intake at «Luna-24» automatic station. The project provides the newest data, since the very first borehole in Tsiolkovsky crater or in another landing place. But only the basic network as a whole can ensure the regular data according to the put basic scientific problems.

Acknowledgments. We thank A.T. Basilevsky for useful remarks.
PRELIMINARY DATA ON PHYSICAL AND MECHANICAL PROPERTIES OF SAYH AL UHAYMIR 001 METEORITE. E.N. Slyuta¹, S.M. Nikitin¹, A.V. Korochantsév¹, C.A. Lorents¹, Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow, Russia. slyuta@mail.ru.

Introduction. As against of the data on chemical, mineralogical and isotopic composition of meteorites the physical and mechanical properties of ones of different classes and types were not investigated almost. Without knowledge of meteorite-analogue strength properties the age of an asteroid by the data on a saturation of impact craters sometimes even approximately does not yield to estimation [1], or the minimal and maximal value may differ by two (!) orders of magnitude [2]. It is also a problem of meteoroid breakup height in the Earth upper atmosphere and an estimation of a meteorite shower area [3]. It is a problem of tidal asteroid destruction within the Earth gravitation field and an adequate estimation of asteroidal hazard [4]. It is a problem of creation of reliable engineering models of space objects for their research with spacecrafts. Parameters of transition between the small and the planetary Solar system bodies depend on strength and rheology properties of material [5, 6]. But unfortunately, strength and rheology properties of material practically do not yield gracefully to theoretical investigations and their almost unique source is the experimental data.

Sayh al Uhaymir 001 meteorite. Sayh al Uhaymir 001 (SAUH 001) is a stony meteorite shower found March 16, 2000 and is one of Oman’s largest known meteorite showers [3]. More than 2670 samples weighing more than 450 kg have been collected. On composition the meteorite is ordinary chondrite of L4/5 petrographic type (Fayalite - 24.7 mol%; Ferrosilite - 21.4 mol%) with S2 shock stage. It is ordinary chondrite of L4/5 petrographic type (Fayalite - 24.7 mol%; Ferrosilite - 21.4 mol%) with S2 shock stage.

Technique of experimental researches. Physical and mechanical properties of a meteorite were investigated by a standard technique according to working instruction on research of rocks and ores at carrying out of geological engineering survey [7, 8]. The sample of a meteorite of approximately 9×10×12 cm in size has been cut on three perpendicular each other plates, everyone by thickness of 2 cm, and on one cube of 4×4×4 cm in size with the sides parallel to all three plates (Fig. 1). Such technique allows to investigate volumetric anisotropy of physical and mechanical properties in a sample.

The meteorite density was measured with a method of hydrostatic weighing on densitometer DG-1 with accuracy of weight measurement up to 0.001 g. Accuracy of density measurement in test is 0.01 g cm⁻³. 24 of density measurements have been executed. The density average value is 3.45 g cm⁻³ (Table 1).

Speed of P-waves was measured with a dynamic method which is based on excitation in a sample of wave fluctuations and registration of the first introduction moment by the ultrasonic receiver. The cross size of a sample was measured with a micrometer with accuracy of 0.01 mm at the moment of passage of an acoustic wave. For acoustic contact on opposite sides of samples the thin layer of Vaseline was applied. The measurement accuracy is 15 m c⁻¹. Speed of P-waves was measured for each of three sample plates in three, perpendicular each other, directions and in several places on each direction. Speed of P-waves in cube was measured in three perpendicular directions, and also in a direction of all cube corners. In total 216 measurements have been made (Table 1).

Compressive and tensile strength was measured on air-dry samples by CD-100 and CD-10 testing machines (VEB Werkstoffprüfmaschinen, Leipzig, Germany) which allow to carry out proportional loading in a range of up to 100 and 10 tonnes accordingly. The splitting method was applied to measure of tensile strength with measurement of the enclosed load and destroying effort [7]. The length of split was measured with a margin error no more than ±0.5 mm at length not less than 20 mm. Each plate depending on its sizes broke up from 4 up to 10 samples of the semi regular shape of (20-30)×(20-30)×(20-30) mm in size. At calculation of tensile strength it was used Hertz equation. In total 27 measurements have been made. Average value of tensile strength is 16.12 MPa (Table 1).

Compressive strength was measured with crushing the cube and the samples of semi regular cubic shape received during splitting of plates after tensile strength measurements [7, 8]. In total 28 measurements have been made. The average value of compressive strength is 101.6 MPa. Compressive strength of the cube is 112.44 MPa.

Deformation characteristics (Young modulus and Poisson coefficient) were measured at static loading both in tests for compression, and for a tension. Registration of deformation characteristics was made with three electronic instruments such as "MKe" and "MKe-C" (Ing. Bernhard Holle Feinmechanische Werkstätten, Magdeburg, Germany) and the electronic dynamometer on the basis of EMP-109 IMZ automatic potentiometer. The basic error of deformation records are no more than 0.01%, and of load records are no more than 0.5%. Deformation of samples was made in a mode of continuous proportional loading. The deviation from proportionality nearby ultimate strength point did not depend on the test specifications determined by the operator (Fig. 3). Processing of diagrams in coordinates of load - time and load - movement (deformation) has been executed with use of special methods of the random processes schematization [7, 8]. Young modulus and Poisson coefficient was calculated with equations for uniaxial load at registration of longitudinal and cross deformations in two directions, both in a compression mode, and in tension. The total of Young modulus measurements is 45. Average value of Young modulus is 0.171×10⁵ MPa. The total of Poisson coefficient measurements has been made 39. Average value of Poisson coefficient is 0.33 (Table 1).
Summary. Basically igneous terrestrial rocks are characterized by much higher strength properties than the meteorite. Similar on strength properties terrestrial rocks to the meteorite are few of them - troktolite gabbro-diabase, olivine gabbro, dolerite [9], fine-grained gabbro and serpentinous dunite [10]. But these rocks are characterized by lower values of Poisson coefficient and much higher values of Young modulus. P-waves speed of the compared terrestrial rocks almost twice are larger, and for dunite is three times larger than for the meteorite. On set of physical and mechanical properties of the meteorite (Table 1) analogues among terrestrial igneous and sedimentary rocks and ores are not present [10].

The used technique of researches allows to receive enough statistics of measurements and, accordingly, reliable enough data on rather small volume of a material. Besides the given technique allows to investigate also volumetric anisotropy of physical and mechanical properties and its dependence on structural, mineralogical and petrographic characteristics of material. The data on anisotropy are in process. Researches of specific density, porosity and electromagnetic properties of the SAUH 001 meteorite proceed.

Table 1. Physical and mechanical properties of SAUH 001 meteorite.

<table>
<thead>
<tr>
<th>Name</th>
<th>Quantity of measurements</th>
<th>Average value</th>
<th>Range of measurements</th>
<th>Variation coefficient, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density, g cm⁻³</td>
<td>24</td>
<td>3.45</td>
<td>3.37 – 3.46</td>
<td>0.6</td>
</tr>
<tr>
<td>Compressive strength, Mpa</td>
<td>28</td>
<td>101.6</td>
<td>47.5 – 173.9</td>
<td>31.1</td>
</tr>
<tr>
<td>Tensile strength, Mpa</td>
<td>27</td>
<td>16.12</td>
<td>6.46 – 27.38</td>
<td>32.5</td>
</tr>
<tr>
<td>Young modulus, 10⁻⁵ MPa</td>
<td>45</td>
<td>0.171</td>
<td>0.102 – 0.185</td>
<td>12.1</td>
</tr>
<tr>
<td>Poisson coefficient</td>
<td>39</td>
<td>0.33</td>
<td>0.18 – 0.46</td>
<td>21.2</td>
</tr>
<tr>
<td>P-wave, m c⁻¹</td>
<td>216</td>
<td>3121.9</td>
<td>2161.7 – 4718.9</td>
<td>20.4</td>
</tr>
</tbody>
</table>

Fig. 1. The cutting plan of a meteorite. The meteorite has been cut on three perpendicular each other plates and on one cube with sides parallel to all three plates.

Fig. 2. The deformation-loading characteristic of compression of the meteorite cubic sample.

ICE ON THE MOON: REANALYSIS OF THE ORIGIN AND SURVIVAL CONDITIONS.

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Introduction: In the permanently shadowed areas of polar regions of the Moon and Mercury ice deposits were supposed to occur [1-3]. The upper temperature of survival $T_{\text{sur}}$ is usually defined as the temperature of evaporation of 1 m thick ice layer during 1 Gyear. For bare solid ice in vacuum $T_{\text{sur}} \approx 114$ K. If higher evaporation rate ($1m/100$ Myears) is admitted, $T_{\text{sur}} \approx 119$ K. Two factors can slow down ice evaporation: (1) burial of ice deposits by crater ejecta and (2) return of evaporated molecules back to cold polar traps. The first factor is trivial but was not calculated; the second was never discussed before. Here we calculate ice evaporation rates and $T_{\text{sur}}$ for both slowing mechanisms and consider constraints on the origin of ice on the Moon.

On the origin of ice on atmosphereless bodies: A few mechanisms of ice origin on the Moon and Mercury were proposed [1,2]. Such mechanism as degassing of the lunar interior is hardly probable from geological point of view. Water production from silicates by solar wind protons is hardly probable too, because $H_2O$ molecules penetrating to silicate matrix dissociate into OH and H, the latter being bonded to a neighbor oxygen atom by hydrogen or covalent bond [4]. So OH groups that form in silicates in proton bombardment [5] do not tend to associate with implanted H to form $H_2O$ molecules.

Among the mechanisms of ice origin only water accumulation in cometary impact cannot be rejected, though cannot be validated. In attempts to validate it [6,7], ice accumulation in cold polar traps by random ballistic hops of water molecules in the lunar or Mercurian exospheres was considered. In such hops, molecules contact only with planetary surface and not collide with each other. Let us show that such collisionless temporary atmosphere contains small amount of water molecules, which can provide only thin ice deposits ($<<150\ \AA$).

To collide only with lunar surface, molecules should have free path in gaseous phase $\lambda$ much greater than typical hop length $\lambda_h \approx 200$ km. Density of such atmosphere is $n = 1/\lambda^3 << n_{\text{max}} = 1/\lambda_{\text{max}}^3 \approx 4 \times 10^8\ \text{cm}^{-3}$ ($\lambda$ is molecular size). Height scale of a cold atmosphere $H_{\text{si}} = kT_i/mg \approx 100$ km (k is Boltzmann constant, $T_i$ is typical recoil temperature, $m$ is the mass of water molecules, and $g$ is lunar gravity). Then the number of molecules in collisionless atmosphere $N << n_{\text{max}} \approx 4\pi R_i^3 H_{\text{si}} \lambda_{\text{max}}^3 \approx 4 \times 10^{12}$, their total volume in solid state $<< 4.4 \times 10^6\ \text{cm}^3$. They may originate from a cometary nucleus as small as $<30$ m (half of the parent nucleus is assumed to be water ice and $1/4is$ supposed to escape in impact).

Since mass contribution from impacts $<30$m is almost entirely due to microimpacts ($<1cm$) [8,9], the analysis [6,7] is valid rather for water-containing meteorites and interplanetary dust than for comets. If $w\%$ of the molecules of collisionless atmosphere are deposited in cold polar traps that take $p\%$ of lunar area, the depth of ice deposits is $h_d << h_{\text{max}} = a(H_{\text{si}}/\lambda_{\text{max}})(w/p)$.

From estimates on the base of crater statistics [10], $p = 1\%$; from comparison of the recent south pole shadow model [11] with a previous model [12] for total lunar surface, $p = 0.05\%$. Our calculations with improved analytical version of the numerical models [6,7] give $w$ from 5.6% to 36% for $p$ from 0.05% to 1% of the lunar area. I.e., $h_{\text{max}}$ is from $\approx 40$ Å to $\approx 130$ Å. Deposition of this layer takes about a lifetime $\tau_w$ of water molecules after impact ($\tau_w = 6.7 \times 10^3$ s in lunar exosphere), i.e., much shorter than the time interval between meter-scale impacts. Total contribution of small impacts ($<30$ m) to ice deposition is $v_d \approx (3/4)N_{\text{v}}\xi(w/p)$, (1)

where $N_{\text{v}} \approx 3 \times 10^6\ \text{cm}^{-2}\ \text{s}^{-1}$ is the number of molecules in the total flux of impactors <30m calculated from [8,9], $\xi$ is the fraction of water molecules among them. $\xi$ can be evaluated from concentration of $H_2O$ molecules in lunar exosphere

$n_{\text{RO}} \approx (3/4)N_{\text{v}}\xi(w/H_2O)$, (2)

Comparison with the observed value $n_{\text{RO}} \approx 0.5\ \text{cm}^{-3}$ [13] yields $\xi \approx 3 \times 10^{-5}\%$. Substituting (2) to (1), obtain

$v_d \approx \xi(w/p)$, (3)

which is 72 $\mu$m/Gy for $p = 0.05\%$ and $\approx 9.8\ \mu$m/Gy for $p = 1\%$, the total ice volume rate being $1.4 \times 10^{12}$ and $8.9 \times 10^{12}$ cm$^3$/Gy. These deposition rates exceed the evaporation rates of bare ice in cold traps (see below) at $T < 98$ K and $T < 100$ K, respectively.

Large cometary impacts. They form extremely nonstationary, inhomogeneous and rather dense temporary atmosphere. Models of its expansion and cooling require information about the optical constants of hot vapor in wide temperature, density, and wavelength ranges. No ice deposition can occur, if the expanding vapor that reaches polar regions is too hot. So we may only assume that a part of the water from comet nucleus is deposited in permanent shadows. E.g., if 1% of $H_2O$ molecules from a cometary nucleus is deposited in permanent shadow, $\tau_{\text{d}} \approx 10^3$ s. 

The main doubt about the origin of water from impacts, either cometary or micrometeorite, is that we do not know if water molecules survive impact or they are irreversibly destroyed by ionization and dissociation in hot plasma which expands so quickly that recombination can be prevented. Let us estimate the survival conditions of ice deposits, if any.
Evaporation of ice from under regolith: Such process includes evaporation of bare ice and diffusion of vapor through regolith. At temperatures $T < 150$ K evaporation rate $v_0$ of bare ice in vacuum is so low that heat flow from the lunar interior is enough to supply heat for evaporation and it becomes temperature-controlled:

$$v_0 = \frac{p_0 \omega (2 \pi m k T)^{1/2}}{p_0 \omega Dq}, \quad (4)$$

where $p_0 = 2 \times 10^{-7} \exp(-6052/T)$ dyn/cm$^2$ is water vapor pressure at temperature $T$, $\omega = 3.2 \times 10^{-23}$ cm$^3$ is the volume per molecule in ice. Evaporation rate of ice at a depth $H$ under regolith is

$$v_H(H) = \frac{v_0}{1 + v_0 k T H / p_0 \omega Dq}, \quad (5)$$

where $q$ is the volume fraction of pores, and $D = h/3$ is diffusion coefficient of water vapor in regolith with pore size $l$, $v = 8(8kT/\pi m)^{1/2}$ being the average thermal velocity of molecules of mass $m$. At large $H$ evaporation becomes diffusion-controlled and (5) is reduced to

$$v_H = v_0 (4lq/3H) \quad (6)$$

Ice deposition from cometary impact takes up to tens of hours; then gradual burial of ice by ejecta from craters occurs, with burial depth increasing with time $t$: $H = v_d t$. The thickness $h$ of evaporated ice is derived from the differential equation:

$$dh/dt = v_B(H), \quad (7)$$

where $v_B$ is described by (5). Integration of (7) gives

$$h = \left(\frac{p_0 \omega Dq}{v_0 k T h}\right) \ln \left[1 + \frac{v_0 k T h / p_0 \omega Dq}{v_0 k T / p_0 \omega Dq} h\right] \quad (8)$$

The results of calculations at burial rate $v_B = 10$ m/Gy are presented by dot lines in Fig.1. As shown, burial of ice by regolith can raise $T_m$ by no more than 15-25 K. Increasing or decreasing of $v_0$ by a factor of N shifts the upper parts of the curves in Fig.1 approximately by N degrees to the right or left, respectively.

**Fig.1.** Thickness of ice layer evaporated in 0.5 Gyear at different temperatures: bare ice (green lines), ice gradually covered with 5 m-thick regolith of porosity 0.5 and pore size 2 mm (blue) or 50 mm (red). The case of no return of the evaporated molecules back to cold traps is presented by dot lines; dashed and solid lines show the case of return to traps of 0.05% and 1% of the lunar area, respectively.

Return of evaporated molecules back to cold polar traps: Thermal velocities of the molecules evaporating even from the equatorial zones of the lunar or Mercurian surfaces are too low for escape from the gravity of these planets. The loss mechanisms for water molecules is photodestruction (dissociation or ionization). Since its characteristic time $\tau_d$ is much longer than the time of typical jump of an evaporated molecule over the surface of a planet ($\sim 100$ s), many jumps are allowed to the molecules. All molecules that are not destroyed during jumps are captured again in permanent shadow.

Using analytical model of ballistic hops over a planet, we calculated the probability of repeat capture $w_e$ and the coefficient of “true” evaporation (the fraction of irreversibly lost molecules) $w_r = 1 - w_e$ for various areas and temperatures $T$ (K) of permanent shadows. For the Moon,

$$w_e,0.05\% = 4.53 \times 10^{-3}(T - 15), \quad (9a)$$

$$w_e,1\% = 4.53 \times 10^{-4}(T - 48)^{0.315}, \quad (9b)$$

where the subscripts denote the fractions of the permanent shadows, To take the return of molecules to cold traps into account, $v_0$ in (4) and $h$ in (8) should be multiplied by (9a) or (9b). The results are shown in by dashed and solid lines in Fig.1.

**Conclusions:** (1) The only probable hypothesis of ice origin on silicate atmosphereless bodies is contamination from comets and water-containing meteoroids, but it is still not validated enough.

(2) Ice burial under regolith together with return of most of the evaporated water molecules back to cold traps cannot increase survival temperature of ice more than by 20-35 K. The resulting $T_m < 150$ K, which does not enable ice to survive on sunlight areas even under deep regolith cover.

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In addition to the previously widely used methods of analysis - neutron - activation (NAA) and mass-spectrometric (MS) is proposed the version of the atomic-emissive determination of platinum metals, silver, gold, molybdenum, tungsten and other elements, which play important role in the study of the evolution of terrestrial and cosmic matter [1, 2].

We used a device, which includes spectrograph PGS-2, two-jet arched plasmatron (as the source of spectra excitation), sample input unit, ), photoelectronic cassettes on linear CCD ("MORSE" Ltd.) for spectra registering and computer (with the program of processing spectra and automatic calculation of the content of elements). The temperature of plasma is 8000K, what made it possible to determine, for example, noble metals, molybdenum and tungsten at the level of concentrations n (10^{-5} -10^{-6})%.

Sample mass was - 30 -70 mg (mixed with the carbon powder), exposure - 5 s, error - 5-10%, the calculation of contents was made according to the reference materials.

The obtained results are comparable with the data of NAA and MS. The procedure was used for analysis of samples of rocks and some meteorites.

Coronae and Arachnoids of Venus Revisited: Sizes and Topographic Characteristics.

Introduction: We presented the first findings about our survey of coronae and arachnoids on Venus in the Vernadsky-Brown Microsymposium 44 in 2007 [1]. We suggested then that there probably was even more coronae and arachnoids on the surface. Here we present results of the re-survey that we have done from winter 2006 onwards in order to verify that all possible coronae and arachnoids were recognized from the Magellan data. Here we report on the corona and arachnoids numbers, sizes (maximum widths), topographic characteristics as well as a few preliminary results of the statistical analyses of the topographic measurements.

Methods: We re-surveyed the global distribution of coronae and arachnoids on Venus based on existing catalogs [2-4] of coronae and related volcano-tectonic structures and by carefully and systematically studying Magellan radar images and topography data with the help of the ArcGIS 9™, the USGS Pigwad server and the USGS Map-a-planet service. In this survey we made sure that all areas of the planet covered by the Magellan SAR images were systematically checked.

The maximum width (diameter) of each feature was either taken from published sources (as the largest value if the listed values were different in each catalog) or measured from Magellan images or topography map with the ArcGIS. Topographic characteristics (maximum and minimum elevation, maximum rim height, minimum and maximum interior elevation and basal altitude) of each corona and arachnoid were also measured again from the improved Magellan topographic map [5]. We plan to measure the maximum widths of all these structures also ourselves to get more systematic and independent size measurements.

Numbers: We found a total of 857 (707 coronae and 150 arachnoids; 576 of type 1 and 281 type 2) in our current survey (Table 1). This is a remarkable 67% increase compared to e.g. [2]. The maximum diameters range from a few tens of kilometers to well over 1000 km (2600 km for Artemis), but majority of coronae have diameters between 100 and 270 km and arachnoids between 80 and 200 km (Table 2). Geometric means of Type 1 and Type 2 coronae (and arachnoids) may be statistically different in contrast to what was shown in [6]. This is probably a consequence of a larger number of small Type 2 coronae and arachnoids that were recognized in this survey. This does not necessarily mean that Type 1 and Type 2 coronae have different formation mechanisms. The geometric mean width of arachnoids is significantly smaller than that of coronae (both for Type and Type 2).

Topographic characteristics: We classified all the coronae and arachnoids into the 9 different topographic groups defined by Smrekar and Stofan [7]. We included a new class of 3c into the classification because of the difficulties in placing some coronae into the previous classification and discrepancies we noticed in comparing topographic groups of a few coronae listed in [8 ,9] with our results.

Overall there are not any major differences between group frequencies for Type 1 coronae, but there is a big difference in the proportion of Type 2 coronae in groups 4 and 7. This may be just due to difficulties in interpreting the topographic form and profiles of some coronae, which are placed on slopes and which therefore can have properties that fit both group 4 and group 7 (on one side of the corona the topography of the immediate surroundings is lower than on the other, and thus the reference level is different on different sides of the structure). We are currently re-checking the Type 2 corona topographic forms, so that we can be more confident in our numbers.

We have started to analyze the topographic measurements in detail. Preliminary statistical results suggest that the basal altitude of the Type 1 coronae and arachnoids may be different than of the Type 2. Type 2 coronae may be located on somewhat lower elevations. For Type 1 coronae and arachnoids it also seems that topographic group 8 structures (rimless depressions) are situated on higher elevations than group 4 and 7. Other groups overlap both group 8 and groups 4 and 7. Also there is some indication that group 8 corona (Type 1) are deeper than group 3b, 3c or 4 corona. The implications of these findings are not clear but this may suggest that coronae and arachnoids, which are rimless depressions, may have more than one formation mechanism [e.g. 10]. We are currently making other statistical comparisons (e.g. comparing maximum rim heights between Type 1 and 2 structures and between topographic groups 3a-c, 4 and 7) and will present a few of the more relevant results in the Microsymposium.

Table 1. Numbers of coronae and arachnoids found in this survey compared to earlier lists.

<table>
<thead>
<tr>
<th>Class of structure</th>
<th>This survey</th>
<th>Stofan et al., 2001</th>
<th>Crumpler &amp; Aubele, 2000</th>
<th>Kostama &amp; Aittola, 2003</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coronae</td>
<td>707</td>
<td>513</td>
<td>209</td>
<td>350</td>
</tr>
<tr>
<td>Type 1</td>
<td>483</td>
<td>406</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Type 2</td>
<td>224</td>
<td>107</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Arachnoids</td>
<td>150</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Type 1</td>
<td>93</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Type 2</td>
<td>57</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Coronae &amp; Arachnoids Type 1</td>
<td>576</td>
<td>406</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Coronae &amp; Arachnoids Type 2</td>
<td>281</td>
<td>107</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

857 513 474 446

Table 2 a) Summary statistics from [6].

<table>
<thead>
<tr>
<th>2 a)</th>
<th>Coronae Type 1</th>
<th>Coronae Type 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>N</td>
<td>406</td>
<td>107</td>
</tr>
<tr>
<td>Mean [km]</td>
<td>256.8</td>
<td>236.8</td>
</tr>
<tr>
<td>Geometric mean [km]</td>
<td>220.7</td>
<td>208.0</td>
</tr>
<tr>
<td>St. Dev. [km]</td>
<td>150.9</td>
<td>127.8</td>
</tr>
</tbody>
</table>

Table 2 b) Summary statistics from this survey (excl. Artemis Corona).

<table>
<thead>
<tr>
<th>2 b)</th>
<th>Coronae Type 1</th>
<th>Arachnoids Type 1</th>
<th>Coronae Type 2</th>
<th>Arachnoids Type 2</th>
<th>All coronae</th>
<th>All arachn.</th>
<th>All Type 1</th>
<th>All Type 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>N</td>
<td>482</td>
<td>93</td>
<td>224</td>
<td>57</td>
<td>706</td>
<td>150</td>
<td>576</td>
<td>281</td>
</tr>
<tr>
<td>Minim. [km]</td>
<td>35</td>
<td>30</td>
<td>40</td>
<td>32</td>
<td>25</td>
<td>30</td>
<td>25</td>
<td>32</td>
</tr>
<tr>
<td>Maxim. [km]</td>
<td>1060</td>
<td>500</td>
<td>700</td>
<td>328</td>
<td>1060</td>
<td>500</td>
<td>1060</td>
<td>700</td>
</tr>
<tr>
<td>Mean [km]</td>
<td>238.6</td>
<td>124.0</td>
<td>202.2</td>
<td>127.5</td>
<td>228</td>
<td>125.4</td>
<td>220.1</td>
<td>189.5</td>
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<tr>
<td>Geom. mean [km]</td>
<td>199.8</td>
<td>108.4</td>
<td>177.0</td>
<td>111.0</td>
<td>205.0</td>
<td>109.4</td>
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<td>161.0</td>
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<tr>
<td>St. Dev. [km]</td>
<td>151.9</td>
<td>69.7</td>
<td>115.3</td>
<td>66.9</td>
<td>191.2</td>
<td>68.4</td>
<td>148.0</td>
<td>111.6</td>
</tr>
<tr>
<td>Median [km]</td>
<td>200</td>
<td>100</td>
<td>180</td>
<td>120</td>
<td>190.5</td>
<td>108</td>
<td>180</td>
<td>170</td>
</tr>
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</table>

Table 3. Topographic groups of Type 2 coronae and arachnoids (* from [2 ]).

<table>
<thead>
<tr>
<th>3</th>
<th>Type 1 Coronae</th>
<th>Type 1 Coronae and Arachnoids (This survey)</th>
<th>Type 2 Coronae</th>
<th>Type 2 Coronae and Arachnoids (This survey)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 – Domes</td>
<td>7%</td>
<td>6%</td>
<td>0%</td>
<td>6%</td>
</tr>
<tr>
<td>2 – Plateaus</td>
<td>13%</td>
<td>15%</td>
<td>7%</td>
<td>5%</td>
</tr>
<tr>
<td>3a – Rimmed plateaus</td>
<td>9%</td>
<td>11%</td>
<td>6%</td>
<td>7%</td>
</tr>
<tr>
<td>3b – Rims with central high</td>
<td>16%</td>
<td>8%</td>
<td>4%</td>
<td>5.5%</td>
</tr>
<tr>
<td>3c – Rims with int. dome</td>
<td>-</td>
<td>7%</td>
<td>-</td>
<td>5%</td>
</tr>
<tr>
<td>4 – Rimed depressions</td>
<td>28%</td>
<td>25%</td>
<td>23%</td>
<td>41%</td>
</tr>
<tr>
<td>5 – Outer rise, trough, inner high</td>
<td>5%</td>
<td>1%</td>
<td>0%</td>
<td>0%</td>
</tr>
<tr>
<td>6 – Outer rise, trough, inner low</td>
<td>1%</td>
<td>2%</td>
<td>0%</td>
<td>0%</td>
</tr>
<tr>
<td>7 – Rim only</td>
<td>8%</td>
<td>6%</td>
<td>54%</td>
<td>17.5%</td>
</tr>
<tr>
<td>8 – Depressions</td>
<td>10%</td>
<td>19%</td>
<td>4%</td>
<td>13%</td>
</tr>
<tr>
<td>9 – No apparent signature</td>
<td>3%</td>
<td>(&lt;0.5%)</td>
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ON CONDENSATION RESERVOIR OF CAI OF CARBONACEOUS CHONDRITES. Galina K. Ustinova, Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow V-334, 119991 Russia; E-mail: ustinova@dubna.net.ru

Introduction: The aggregates of CAIs (calcium-aluminum-rich inclusions) of carbonaceous chondrites consist of some high-temperature minerals, such as perovskite, mellilit, spinel, pyroxene et al., which are considered to be the first condensates of the primordial matter. It is natural that their origin is in the focus of the most intensive investigation and heated controversies.

According to the physical-chemical parameters, the refractory minerals of CAI could be condensed in the reservoir of the disrupted supernova (case 1, e.g. [1-3] et al.), as well as in the region of reconnection of the magnetic fields between the protosun and the inner embedded edge of the forming protoplanetary disk (case 2) [4]. In the first case CAIs were formed in the periphery of the collapsing protosolar nebula surrounded by the expanded supernova shell [5], whereas in the second case they originated at the distance of about 0.06 AU and they were thrown out to planetary distances by some bipolar outflows (the x-winds of the disk inflows). In both the cases CAI could be enriched with the extinct radionuclides and pure oxygen isotope $^{16}$O. In the supernova reservoir, they are the products of the nucleosynthesis and, besides, the extinct radionuclides could be produced by spallation of some target nuclei with high energy nuclear-active particles (e.g. [3, 6-12] et al.). In the second scenario the extinct radionuclides, in principle, also could be the spallation products of high energy nuclear-active particles at the stage of T-Tauri, and the reservoir of the pure $^{16}$O could be formed as a result of the mass-independent fractionation effects during formation of solids and the photochemical self-shielding effects in CO (e.g. [4, 13-15] et al).

However, the reservoir of the disrupted supernova, as well as the range of interaction of the strong protosolar winds with the matter of the embedded edge of the forming protoplanetary disc, are the regions of matter highly reprocessed by the shock waves. That implies some rigid radiation conditions onto the formation of the extinct radionuclides, which is considered below.

High radiation conditions during CAI formation: The supernova explosion established peculiar radiation conditions in the early solar system [16]. The tremendous explosive shock wave and supersonic turbulence resulted in acceleration of particles in the cosmic plasma with forming a power-law energy spectrum $F(\geq E_0) \sim E^{-\gamma}$ of very high rigidity ($\gamma \rightarrow 1$) [17]. Shock waves pick up new particles from the background plasma and pump over the particles from the low energy range of the spectrum to the high energy one. That leads to the enhancement of fluxes of nuclear-active particles (and, therefore, of spallation production rates of isotopes) above the energy $E_0$ (e.g. above the threshold energy of nuclear reaction) by one-two orders of magnitude [18]. That strongly increases the share of the spallation processes in the last nucleosynthesis event of the primordial matter of the solar system. For instance, the consideration of the problem of Li, Be and B generation in the high radiation conditions of the supernova explosion not only ensures the observed abundances of the light elements, but it also allows us to understand why $^7$Li survived better than other isotopes [19].

One may see the growth of the $^{26}$Al/$^{27}$Al ratio with decreasing $\gamma$ in Fig.1 [20]. The canonical value of the ratio and its higher observed values in CAI of carbonaceous chondrites [7] point out to very rigid radiation conditions with $\gamma \sim 1$. A similar growth is exhibited by the production rates of Ne isotopes in the gas-dust nebula and SiC [21]; the measured levels of Ne-E(H) and $^{21}$Ne production rates in SiC of the Murchison chondrite [22] testify to the high radiation too. Similar regularities are observed for other extinct radionuclides and noble gases [21]. Therefore, in the absence of the additional nucleogeneic processes of the isotopes (case 2) the observed contents of cosmogenic nuclides in CAI could be produced only under the most rigid radiation ($\gamma \sim 1$). However, such a rigid irradiative criterion is just the distinctive criterion of the supersonic turbulence in the supernova neighborhood [23], which allows us to give preference to the supernova (case 1).

The absence of the products of r-process among the extinct radionuclides in CAI with the intervals of formation $\leq$ 1 Myr indicates that the last supernova was the Type Ia supernova (Sn Ia), which could not survive the carbon explosive burning and was fully disrupted [20,24]. The Sn Ia explosion injected 0.4-0.6 $M_\odot$ of iron nuclei ($^{56}$Fe) into the protosolar nebula [25]. Their shock wave acceleration was especially strong [26]. Indeed, the pre-accretion tracks of VH-group nuclei in silicate minerals of some low-metamorphosed ordinary chondrites demonstrate the much flatter gradients ($g \sim 0.7-2.0$) than those for the solar cosmic rays ($g \sim 3.0$) [16].

Discussion: Aggregates of CAI occur mainly in carbonaceous chondrites (6-12%, 10-15% and 5% of volume in CV, CO and CM chondrites, respectively), whereas in ordinary and enstatite chondrites they occupy only <1% and <1% of volume, respectively [27]. One may suppose that the matter of the carbonaceous chondrites was formed in the peripheral regions of the protosolar nebula, enriched with the fresh synthesized matter of the supernova (in particular, with the unburned C and O nuclei of Sn Ia),
CONDENSATION RESERVOIR OF CAI OF CARBONACEOUS CHONDRITES: Galina K. Ustinova

The arguments for the early condensation of the CAI minerals and the late accretion of the carbonaceous chondrites are discussed, for instance, by R. T. Dodd [28]. The radial gradient of decrease of the extinct $^{53}$Mn with decreasing heliocentric distance, which has been derived in [10, 29], testifies to the entrance of the supernova matter into the protosolar nebular just from its periphery. The similar gradient must be operative for CAI distribution, and so there were few CAI inside the regions of the ordinary chondrite matter formation and, especially, inside the region of the enstatite chondrite matter formation, being the nearest to the protosun. Besides, of course, while being drawn to the protosun, a part of CAI could vaporize and recondense.

ORBITS AND PROBABLE PARENT BODY OF THE KILABO AND BENSOUR LL6-CHONDRITES.

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Introduction: Striking resemblance of two LL6-chondrites fallen within five months of each other in Africa in 2002 attracts general attention [1,2]: the chondrite Bensour of total mass of ~45 kg has fallen on February 11th along the Algerian/Moroccan border, and the chondrite Kilabo of total mass of ~19 kg has fallen on July 21st in northern Nigeria. In particular, their similar petrography and fayalite composition: (F_{25-30}) and (F_{26-30}), respectively, could indicate a common source of origin of both the chondrites. Moreover, the parent body of these chondrites is supposed to be the main belt asteroid 3628 Bozemcov: it was shown earlier in [3] that the reflectance spectrum of that asteroid is practically coincident with the average reflectance spectrum of LL6-chondrites and, in particular with that of the Manihoom LL6-chondrite.

The problem of origin and evolution of meteorites cannot be solved without the knowledge of their orbits. Just the orbits can identify the belonging of meteorites to some family of the celestial bodies, among which the sources of meteorites, their parent bodies, should primarily be looked for. Nowadays, the orbits of only four chondrites are known exactly: Pribram (H5, fall on 7.04.1959, q'=4.05 AU), Lost City (H5, 3.01.1970, q'=2.35 AU), Innisfree (LL5, 5.02.1977, q'=2.76 AU) and Peeskill (H6, 9.10.1992, q'=2.10 AU), which, at least, testify to the belonging of the chondrites to the bodies of the Solar system, most likely, to the asteroids. In the report we give the estimates of sizes of the Kilabo and Bensour orbits, which really allow us to consider the asteroid 3628 Bozemcovova as the parent body of the LL6-chondrites.

Isotopic criterion of size of orbits of ordinary chondrites: The isotopic approach elaborated formerly [4-7], based on the content of cosmogenic radionuclide $^{26}$Al, to estimate the position of aphelion q’ of orbits of the chondrites under consideration is used. Indeed, according to radioactivity of $^{26}$Al in the chondrites with known orbits (Pribram, Lost City and Innisfree), the galactic cosmic ray intensity gradient of about ~20-30%/AU along the meteorite orbits, average over ~1 My, exists, so that the $^{26}$Al content in the chondrites of more prolonged orbits is essentially higher. It is possible, at least within the chondrite orbits (~5 AU from the sun), to approximate, with regard to experimental errors, the derived growth of $^{26}$Al content in the chondrites (corrected, certainly, due to the depth effects), by a broken line corresponding to the $^{26}$Al minimal production rate ($H_{min}$) at the average galactic cosmic ray intensity for the solar cycle in the chondrites with q’ ≤ 2 AU and to the $^{26}$Al maximum production rate ($H_{max}$) at the unmodulated galactic cosmic ray intensity in the chondrites of large orbits. The $^{26}$Al production rates $H_{min}$ and $H_{max}$ in L(LL)-chondrites of different sizes, which are calculated with the analytical method [4], are presented in Fig.1.

The $^{26}$Al measured content, or the $^{26}$Al experimental production rate $H_{exp}$ (in dpm/kg) in the chondrites can be expressed in the following form:

$$H_{exp} = H_{min} Z + H_{max} (1 - Z), \quad (1)$$

where $H_{min}$ is the $^{26}$Al production rate during the time Z (in parts of the orbital period P) when the chondrite flew outside the range of ≤ 2 AU, and $H_{max}$ is the $^{26}$Al production rate when the chondrite flew outside the range of 2 AU from the sun during the rest time (1-Z). Therefore, in the frame of the adopted approximation, one may estimate the orbit size of chondrites if their $^{26}$Al content is measured. This regularity can be successfully expressed in the phenomenological form in terms of aphelion q’ [4-6]:

$$q’(Z) = 1.25 + 0.13Z + 0.53Z^2, \quad (2)$$

where q’ is the aphelion in AU. After determination of Z from (1) (according to the data on $^{26}$Al content in the chondrites), one may immediately estimate the aphelion of their orbits from (2). In the case of cosmic bodies, which can fall to the Earth, i.e. for meteorites, the most probable values of the semimajor axis a and the eccentricity e can be pointed out too [4]:

$$a = (q’^2 + 1)/2; \quad e = (q’^2 - 1)/(q’^2 + 1). \quad (3)$$
Orbits of the Kilabo and Bensour chondrites:
According to the $^{60}$Co evidence and the evidence of tracks of VH-nuclei, the pre-atmospheric size of the Kilabo chondrite is ~ 34 cm, and the shielding depth of the investigated sample is 6 ± 3 cm [8,9]. The measured content of $^{26}$Al in this sample is 68±7 dpm/kg, which corresponds to the orbit with aphelion $q' = 3.6±3.1$ AU and to the most probable parameters of the orbit: semimajor axis $a = 2.3$ AU; eccentricity $e = 0.565$; orbital period $P = 1273.2$ days. This orbit as the regularity $r(t)$ (where $t$ is time and $r$ is a heliocentric distance), calculated with the Kepler formulae [10], is shown in Fig.2 together with the orbit of the asteroid 3628 Boznemcova ($q' = 3.2994$ AU, $a = 2.538$ AU, $e = 0.3$, $P = 1475.81$ days [11]). Both the orbits cross at the points at 3.20 AU and at 2.15 AU. The range between 3.1-3.4 AU is characterized with the absence of the chondrite aphelia [7], as well as, apparently, the other cosmic bodies due to some selection of orbits by the dynamical processes in the interplanetary space, by the existence of the Kirkwood ports and secular resonances [12,13], so that any catastrophic collision of the 3628 Boznemcova asteroid in that range was hardly probable. It is believed that just at 2.15 AU, i.e. in such a densely populated range of the interplanetary space, near the inner boundary of the asteroid belt the chondrite Kilabo was excavated from the asteroid 3628 Boznemcova to the more eccentric orbit due to a catastrophic collision with some cosmic body.

![Orbits of the Bensour (Be) and Kilabo (K) chondrites together with the orbit of the 3628 Boznemcova (Bo) asteroid (r is heliocentric distance; t is time on the orbit before the perihelion transit; dashed lines mark the average heliocentric distance of the Kilabo orbit: $r_e = 2.91$AU; asterisk marks intersection of the Kilabo and Bensour orbits with that of 3628 Boznemcova)](image)

Under the reasonable assumption of similar ablation of the Kilabo and Bensour LL6-chondrites (96.4%), the preatmospheric radius of the Bensour chondrite was $R \approx 45$ cm [8], and the shielding depth of the sample investigated in [2], according to the measured ratio $^{22}$Ne/$^{21}$Ne = 1.123, was $d \sim 4$-$7$ cm [8]. The measured content of cosmogenic $^{26}$Al in that sample is $62\pm1.2$ dpm/kg [2], which corresponds to the following orbit: $q' = 3.51\pm3.48$ AU; $a = 2.255$ AU; $e = 0.557$; $P = 1236$ days (see Fig.2). The orbit of the Bensour chondrite is smaller than that of the Kilabo chondrite: seeing the Bensour chondrite had greater mass, it had received a smaller velocity at the explosive pulse. The orbits of the Bensour chondrite and the 3628 Boznemcova asteroid intersect at 2.16 AU and at 3.25 AU, i.e. practically just at the same point near the inner boundary of the asteroid belt, as in the case of the Kilabo chondrite.

Summary: The cosmic ray exposure ages of the Bensour and Kilabo chondrites are different: their exposure ages $T_{21}$ from the content of cosmogenic $^{21}$Ne equal 19 and 33 My, respectively [2]. Basing only on these data, one may suppose that the Kilabo chondrite was ejected due to a catastrophic collision of the 3628 Boznemcova asteroid with a cosmic object near the inner boundary of the asteroid belt 33 My ago, and after 14 My the Bensour chondrite was ejected in the recurrent catastrophic collision of the 3628 Boznemcova in that region. However, taking into account the similarity of structure and content of both the chondrites, the vicinity of their orbits, the intersection of the orbits with that of the asteroid practically at the same point, as well as the complex exposure history of the Kilabo chondrite [9], much more preferable is the scenario suggested in [2]: ejection of both the chondrites from the 3628 Boznemcova near ~2 AU from the Sun 19 My ago, the chondrite Bensour being ejected from the deep layers of the asteroid, which were completely screened from the galactic cosmic rays, and the Kilabo chondrite being, probably, exposed to the galactic cosmic rays on the surface of the asteroid during 14 My. The contemporary size of the 3628 Boznemcova asteroid amounts to ~ 7 km in diameter [11].

References:
THE GEOLOGY OF SATURN’S SATELLITE RHEA SEEN BY THE CASSINI ISS CAMERA: CRATERED PLAINS, IMPACT BASINS, AND TECTONIC STRUCTURES. R. J. Wagner1, G. Neu-kum2, B. Giese1, T. Roatsch1, and U. Wolf2, 1Inst. of Planetetary Research, German Aerospace Center (DLR), Rutherfordstrasse 2, D-12489 Berlin, Germany, e-mail: roland.wagner@dlr.de; 2Inst. of Geosciences, Freie Universitaet Berlin (FUB), D-12249 Berlin, Germany (gneukum@zedat.fu-berlin.de).

Introduction: Rhea has a mean diameter of 1528 km and is the second-largest satellite of Saturn. Rhea has a low average density of 1.233 g cm⁻³ [1] which implies a more or less icy body. Water ice dominates the surface of Rhea, inferred from a high geometric albedo of 0.65 and from the presence of water absorption bands [e.g. 2]. During a close flyby of Cassini at Rhea, the axial moment of inertia could be determined which showed that this moon is more or less a homogeneous, not differentiated body composed of approximately 75% ice and 25% rock and metal [3].

Geology of Rhea prior to Cassini: At a maximum spatial resolution of 500 m/pxl, Rhea was the best one imaged of the saturnian satellites. Voyager-1 images showed a densely cratered leading hemisphere, while the trailing hemisphere, seen only at very low resolution, was characterized by bright, filament-like wispy markings, similar to those on Rhea’s inner neighbor Dione [4]. The densely cratered terrain was subdivided into three to six units by different investigators, based on crater abundance, texture and the presence of lineations [5][6]. Up to three large multi-ring structures were identified [5]. Tectonic features observed are troughs, scarps, ridges (of minor abundance), and lineaments [5][6][7].

Image data base. Rhea was imaged by the Cassini Narrow Angle (NAC) and Wide Angle Camera (WAC) during several non-targeted flybys at resolutions better than 1 km/pxl (in flybys Rhea 00C, 005, 016, 018, 019, 020, 022, 027), with the highest resolutions (better than 10 m/pxl) achieved so far in flyby 018 (Nov. 2005). A targeted flyby (Rhea 049) providing extensive areal coverage and high-resolution imaging is planned for Aug. 30, 2007.

Procedure: The work presented in this paper continuous our investigations of Rhea’s inner neighbor Dione [8] whose geologic features are comparable to those on Rhea. (1) Geologic units are mapped and compared to the units mapped in Voyager data. (2) Ages of these units are obtained from crater size-frequency measurements and from application of impact chronology models. Ages are assigned by means of impact cratering chronology models: (a) Model I with a lunar-like (exponential) decay in cratering rate with time and with a more or less constant cratering rate since about 3 b.y. (billion years) [9][10][11], and (b) Model II with a constant cratering rate throughout most of solar system history [12].

Results: At regional scale (500 -1000 m/pxl), the densely cratered plains on Rhea show little variation in terms of albedo and morphology (Fig. 1). Lineations observed in one unit by [5] appear to be characteristic features of the cratered plains. Despite the high density of craters, measured crater frequencies are production distributions but close to equilibrium, especially at smaller craters. Average cratering model ages in the cratered plains are on the order of 4.2 Gyr (Model I, [11]) or 3.6 Gyr (Model II, [12]). Variations in large-crater abundances discussed by [5][6][13] can be confirmed. Smoother areas with a paucity of smaller craters are also observed but lack clear evidence for cryovolcanic resurfacing, or mantling by airfall deposits discussed by [6]. Based on the measurement of Rhea’s moment of inertia and its more or less undifferentiated state [3] it seems unlikely that cryovolcanic activity has ever occurred, however.

Basins and large craters. Large craters and ring structures (basins) are abundant, but are heavily degraded and are more easily detected and mapped using a digital terrain model (Fig. 2). The anti-Saturnian hemisphere is characterized by two well-preserved basins, 400 to 500 km in diameter. One of these two basins, Tirawa has a slightly elliptic outline. The elliptic shape and the elongated central peak complex infer an oblique impact. Tirawa overlaps another (so far unnamed), larger and more degraded basin to its southwest (Fig. 1). While a lower crater density and hence a younger age was reported for Tirawa [5], our crater counts show that crater frequencies inside Tirawa and the adjacent basin are comparable to the frequencies outside hence both basins are old features on the order of 4 Gyr.

Ray craters. While stratigraphically young, bright ray craters characterize the surfaces of the icy satellites of Jupiter, they are not common on the saturnian satellites. Only one prominent ray crater is found on Rhea, located at lat. 12.5° S, long. 112° W. The bright rays of the 48-km crater show a butterfly wing pattern implying an oblique impact from the east. Using the frequency of superimposed small craters in the continuous ejecta measured on a high-resolution WAC image (34 m/pxl resolution), model ages for this ray crater are either 2.5 Gyr (Model I, [10]), or only 70 Myr (Model II, [11]). This crater has been observed again by ISS during the Aug. 30, 2007, flyby in much more detail and with extended areal coverage to show the context.
**Tectonic features.** Like Dione, Rhea was known for its bright filament-like so-called *wispy streaks* which were observed only at low Voyager resolution on both satellites [4]. On Dione, the tectonic nature of these streaks and various tectonic episodes that created these structures could by verified by Cassini ISS data [8]. On Rhea, Cassini ISS data also revealed the tectonic nature of these structures. While troughs and ridges found in various locations on Rhea infer preferentially extensional and (minor) compressional tectonism [5][7], the en-echelon pattern of scarps and troughs on the trailing hemisphere indicates shear stress.

**Work in progress and outlook:** Large craters and basins as well as major tectonic features will be assigned names on near-term [14] in order to facilitate stratigraphic work (e.g. establishing rock-stratigraphic group and formation names). Some of the new ISS camera data from the recent successful Aug. 30, 2007, flyby (Rhea 049) which have not yet been interpreted when this abstract was submitted will be shown and discussed.

CHEMICAL WEATHERING IN THE AMAZONIAN. M. B. Wyatt, Department of Geological Sciences, Brown University, Providence, RI 02912 (Michael_Wyatt@Brown.Edu).

Introduction: Recent results from the Mars Express (MEX) Observatoire pour la Minéralogie, l’Eau, les Glaces, et l’Activité (OMEGA) have been used to propose three new sequential eras of martian geologic history based on the identification of surface alteration products: (i) a nonacidic aqueous alteration, traced by phyllosilicates (the “phylllosian” era); (ii) an acidic aqueous alteration, traced by sulfates (the “theiikian” era); and (iii) an atmospheric aqueous-free alteration, traced by ferric oxides (the “siderikian” era) [1]. This study addresses the most-recent of these three eras, the siderikian, and demonstrates that the identification of amorphous high-silica alteration products with the Mars Global Surveyor (MGS) Thermal Emission Spectrometer (TES) supports near-surface aqueous chemical alteration during the Amazonian. The alteration is immature, spatially extensive, limited by low water to rock ratios, and likely resulted from surface materials at high-latitudes interacting with near-surface volatiles deposited as ice (snow) during phases of high obliquity.

Background: Figure 1 shows a sketch of the alteration history of Mars from [1] compared to the traditional nomenclature (Noachian, Hesperian, Amazonia) derived from crater-counting statistics. The transition between the phyllosian and theiikian occurred early in martian geologic history and is interpreted to reflect an environmental change from alkaline to acidic conditions. It is proposed by [1] that this change was driven by extensive outgassing of volatiles, including sulfur, linked to widespread volcanic activity. If phyllosilicates formed at or near the surface during a wet early Noachian, [1] also proposes that the change from an alkaline to acidic global environment would have been coupled to a rapid drop in atmospheric pressure in the late Noachian to early Hesperian.

In the siderikian, liquid water did not play an important role based on the lack of hydration of ferric oxides in OMEGA data compared to the detection of hydrated phyllosilicates and sulfates [1]. However, [1] does note that liquid water may have been present during transient and local events (volatile release by impact or melting ice), but these episodes were too short to leave a significant mark on surface compositions. Specifically, transient water events are not responsible for recent global surface alteration, which has mostly been caused by surface oxidation and the production of nanophase ferric oxide, without hydration [1].

In summary, the model by [1] can be seen in two ways: First, as describing a transition from an alkaline to acidic weathering regime (phyllosian to theiikian). Second, as a transition from aqueous chemical alteration to anhydrous alteration (phylllosian and theiikian to siderikian). This study addresses the second part of this model and shows that transient water events have indeed been responsible for widespread, near-surface chemical alteration at high-latitudes on Mars during the Amazonian (siderikian).

Figure 1 Sketch of the alteration history of Mars from [1] compared to the traditional nomenclature (Noachian, Hesperian, Amazonia) derived from crater-counting statistics.

TES High-Silica Alteration Products: The abundances and compositions of TES mineral-phases at high-latitudes on Mars have been a focus of considerable study and debate in recent years. Large expanses of Acidalia Planitia surface materials were originally characterized by the TES Surface Type 2 (ST2) spectral endmember [2]. A central question with this endmember is whether it represents the spectral signature of a high-silica primary volcanic lithology (andesite) [e.g. 2] or the effects of chemical alteration on basaltic surface materials [e.g. 3-8].

Geologic Context: The competing spectral interpretations for ST2 were addressed by [9] in examining the geologic context of the Surface Type 1 (ST1) and ST2 global distribution pattern. It was demonstrated that ST1 materials dominate equatorial and mid-latitude regions and ST2 materials dominate the high-latitude northern lowlands and southern highlands [9]. This spatial distribution is further related to near-surface ice and ice-rich mantle deposits and [9] proposed both a latitude and topographic influence on the global surface alteration of Mars. A gradual transition from ST1 to ST2 in the Southern Hemisphere correlates well with the transition from a lack of ice-rich material (0°–25°S), to a maximum percentage of dissection (25°S–60°S), to uniform mantles of ice-rich deposits (60°S–90°S). This trend is interpreted to re-
fleeting increased amounts of chemical weathering from basaltic interactions with icy mantles. In the Northern Hemisphere, an abrupt transition from ST1 to ST2 occurs at $\sim$20° and is correlated both with ice-rich mantle deposits and the Vastitas Borealis Formation (VBF) boundary. VBF materials have been interpreted as sediments formed by the reworking of near-surface, in situ volatile-driven processes [10] and as a sublimation residue from frozen bodies of water [11]. Thus, alteration of sediments in the northern lowlands may have been enhanced by temporary standing bodies of water and ice.

**TES-GRS Comparisons:** Additional support for high-silica alteration products in the northern lowlands comes from comparisons between MGS-TES and Mars Odyssey (ODY) Gamma-Ray Spectrometer (GRS) observations. The TES and GRS datasets provide unique and complementary insights into martian surface compositions as TES measures the composition of the upper hundred microns of the surface while GRS measures the composition of the upper few tens of centimeters. The TES instrument has mainly been utilized as a mineralogical tool, but thermal emission spectroscopy also provides a means for deriving chemical oxide abundances. MGS-TES chemical compositions are calculated from deconvolved modal mineralologies (vol. %) by combining the compositions of the spectral endmembers (wt. % oxides) in proportion to their relative modeled abundances. Comparisons between TES and GRS chemistries can thus provide insight to the vertical chemical stratigraphy of near-surface materials.

**Results:** The most significant chemical differences between TES ST1 and ST2 are higher abundances of FeO for ST1 (ST1 15.2 % vs. ST2 12.4 %) and higher abundances of SiO$_2$ for ST2 (ST2 57.9 % vs. ST1 53.9 %) [12]. GRS RT2 chemistries however have higher abundances of FeO (RT2 20.1 % vs. RT1 17.6 %) while abundances of SiO$_2$ (RT1 44.7 % and RT2 45.8 %) are similar to the bulk planet and do not show significant spatial variations [12].

**Discussion.** The major chemical trends from TES and GRS appear to be in disagreement. TES ST1 is enriched in FeO while GRS RT1 is depleted in FeO. TES ST2 is enriched in SiO$_2$ while GRS RT2 shows no relative enrichment in SiO$_2$. One can account for these apparent discrepancies, and constrain igneous and alteration processes, by considering the dramatic sampling depth differences between TES and GRS.

Thin coatings or rinds of secondary amorphous high-silica phases (tens of microns) significantly affect the overall shape and position of absorptions in thermal emission spectra of basalt. Such coatings on Mars may form from near-surface ice interactions with little to no water penetrating or cycling into the surface. The probability of sedimentary silica forming on Mars under a wide range of temperature, pressure, and pH conditions was summarized by [13]. Limited degrees of alteration in only the upper few tens of microns of the surface would significantly affect TES derived chemistries and be undetectable to GRS due to its deep sampling depth. Laboratory studies of thermal emission spectra of altered basalts have constrained the secondary phases. The secondary phases likely consist of some combination of Al- or Fe-rich opaline silica, silica-rich allophane-like mineraloids, palagonites, and/or the most silica-rich zeolites [3-8]. The lack of specific hydrated spectral bands in OMEGA near-infrared observations related to such alteration products is not inconsistent, as a combination of compositional, textural, and particle size effects can mask features [16].

**Implications for Water.** The Dry Valleys of Antarctica may be the best terrestrial analogue for weathering on Mars because of the cold, hyper-arid environment, stable permafrost, and ground ice. Basalts in these environments are dominated by plagioclase and pyroxene, with limited abundances of alteration phases similar to those proposed for ST2 materials. Near-surface ice and liquid water, although limited, drives near-surface alteration in this environment, and thus significantly affects thermal infrared spectra. Recent general circulation models have shown transient liquid water as being stable on the martian surface over a range of obliquities and surface pressures [14]. The time periods during the Amazonian when transient liquid water is stable may be the only conditions suitable for chemical alteration. If martian geologic history may be classified based on the presence of surface alteration products, consideration should be given to renaming the Amazonian period to the “silexian”, based on the Latin word silex (flint) from which the word silicon is derived.
CORRELATIONS BETWEEN IRON ABUNDANCES AND LUNAR SURFACE FEATURES: CRATER KEPLER AREA. Lu Yangxiayi, V.V. Shevchenko. Sternberg State Astronomical Institute, Moscow University, Moscow. marsplus@gmail.com

Introduction. Distribution of the iron abundances on the Moon’s surface is important for addressing many lunar science problems. Since iron is one of the major mineral forming elements on the Moon, iron abundance can show composition and the stratigraphy of the lunar crust details, and it can help us to understand the formation, distribution and variety of lunar mare basalts. Beginning with the Apollo missions, a number of regional iron abundance measurements have been made using a variety of remote sensing techniques including gamma-ray spectroscopy, UV-VIS multispectral imaging, and neutron spectroscopy.

Remote sensing data of iron abundances. For chemical analysis of the crater Kepler area we used Lunar Prospector results. Global measurements of iron abundances on the lunar surface were made using the Lunar Prospector Gamma-Ray Spectrometer and Neutron Spectrometer. In this study, we used data derived relative iron abundances from the low altitude, high spatial resolution (~ 45 km² per pixel) Lunar Prospector data using the 7.6 MeV neutron capture gamma-ray doublet [1]. It was found from global Lunar Prospector data there are large expanses of mare basalt in the western mare regions that have very high iron abundances (22–23 wt. % FeO). These features are unusual for mare soils, which typically contain a significant amount of highlands contamination. According to a previous analysis of the authors using thermal and epithermal neutrons the lunar highlands have the low iron abundances (~ 5 FeO wt. %) [2]. It may be demonstrated that the lunar crust formed by a relatively simple magma ocean process.

Crater Kepler area. Relatively fresh impact crater Kepler exposed in the center of the investigated area (Fig. 1). Crater rays are relatively bright albedo features and extend radially away from the center crater (Fig. 2). Usually rays appear to be formed by the excavation and deposition of material from both the main crater and secondary craters. Crater Kepler (32 km in diameter) has intermediate age [3]. It is younger than Copernicus (inferred age of ~ 810 Myr), but older than Tycho (inferred age of ~ 109 Myr). It’s needed to note that crater Kepler placed on the background formation that is possibly part of highland near crater Copernicus. Low albedo basalts occur preferentially along the northwestern areas of Oceanus Procellarum and in the southeastern regions of the studied area, i.e., in Mare Insularum.

Distribution of the iron in the surface layer. To study iron distribution we prepared the local iron map on the base of the Lunar Prospector catalog data [4]. Surface resolution of the data is 0.5°x0.5° (per pixel). The iron values are shown in the Fe weight percent (wt. %). The map is presented in the Fig. 3. The map square is about 600x600 km². Counter interval is 1 wt. %.

Interpretation. The center formation of the investigated area is highland premare fragment around crater Kepler. According map shown in Fig. 3 iron content in this surface material is about 15 – 16 wt. %. Using 1 km/pixel FeO abundances from Clementine and Lunar Prospector GRS spatial footprint information, authors of [5] have been able to obtain plausible thorium distributions around Kepler crater at a resolution of 1 km/pixel. The materials around Kepler crater appear to be a relatively simple mixing of thorium-rich mafic impact-melt breccias compositions and high-thorium mare basalts. Crater Kepler has depth a few kilometers (not more than 5 or 6 km). The material inside this depression has iron content about 13 – 14 wt. %, that is typical for a number of basalts (Fig. 3). According to these data it can suggest that formation around crater Kepler may be fragment of an old premare basaltic structure. It’s needed to note that we can’t identify the crater ray structure around Kepler in Fig. 3. It’s known that crater rays are bright because they excavate immature soils. The bright, optically immature materials gradually darken as a function of exposure time on the lunar surface. Another factor is the amount of mixing of ejecta with local, more mature material when the ejecta is deposited. The ejecta at a crater rim is thicker and less mixed with local materials than distal ejecta or rays, which are thinner and more mixed with local mature soils [6]. In Fig. 3 we can see that ray material has iron abundant about 15 – 16 wt. % near crater rim and about 20 wt. % at great distances from it. As note authors of [6] one is certainly the thickness of the ejecta deposit, which decreases with increasing distance from the crater rim. In the case the process of the local material and ejecta material mixing will be very intensive.

In northwestern and southwestern regions of the studied area we can see two anomalies of iron contents (A, A1, B) where iron abundances are more than 25 wt. %. We suggest these anomalies are places of the volcanic centers – sources of young basalts. Using craters that excavated highland material from beneath the mare basalts in Oceanus Procellarum, authors of [7] estimates that the basalt are 160–625 m thick, with thicknesses ranging from tens to hundreds of meters near the mare/highland boundaries and several hundreds of meters closer to the center of the mare. Data of [3] show that volcanism in the investigated region was active over a long
period of time from 3.93 to 1.2 b.y., a total of 2.7 b.y.

possible the small areas \(A, A1, B\) contain youngest basalts from most depth of basaltic lava layers.

**Conclusions.** In this paper we can try to better understand the current spatial resolution of iron surface distribution in an effort to recognize how well we resolve small area features. For example, the presented map of Fig. 3 shows many small areas in the highland fragment around crater Kepler and western part of Oceanus Procellarum that may or may not be real. By means of the data presented here we try to analyse what we currently understand about the Moon. For example, do lunar mare basalts indeed occur in areas with Fe abundances greater than 25 wt. % in the regolith? What does this apparent limit mean both about how mare basalt is formed in the source regions, where these basalts are, and how basalts with different age were erupted to the surface?

INTERNAL STRUCTURE OF ICY SATELLITES OF JUPITER AND SATURN.
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Models of the internal structure of completely differentiated Europa and Ganymede, and partially differentiated Callisto have been constructed and a comparison with the internal structure of Titan has been made.

Approach. The purpose of this study is to reproduce characteristic features of the internal structure of icy satellites of Jupiter and Titan. The mass and moment of inertia are used as input data for determination of the thickness and phase state of an outer water-ice shell, and the core sizes and masses. Various compositional models are considered for a satellite core: γ-iron core, Fe-10 wt%S core, eutectic Fe-FeS core, and troilitic FeS core. The phase compositions and mantle densities are modelled within the system Na2O-TiO2-CaO-FeO-MgO-Al2O3-SiO2-H2O-Fe-FeS including the solid solutions. The equilibrium phase assemblages were calculated using the technique of free energy minimization. The density variations in the mantle and core radii are found by the Monte-Carlo method [1].

Europa. The results show that Europa is differentiated into a water-ice shell, anhydrous mantle and iron-sulfide core. Both L/LL and CM chondrite compositions match the total mass and moment of inertia value of Europa and can be regarded either as the primary material of Europa (carbonaceous chondrites) or as a reasonable analogue for its anhydrous rock-iron core (ordinary chondrites). The amounts of iron and iron sulfide, and the (FeO/Si) ratio of Europa's anhydrous rock-iron core are not consistent with the bulk compositions of the most oxidized CI chondrites and the most reduced H chondrites. It is likely that Europa inherited a significantly higher proportion of material close to the moderately oxidized L/LL type chondrites rather than to the carbonaceous chondrites. The allowed thickness of Europa's H2O layer (whether liquid or ice) ranges from 115±10 km (6.8±0.6% of total mass) for a differentiated L/LL-type chondritic mantle with a crust to 135±10 km (7.9±0.5%) for an undifferentiated L/LL chondritic mantle (Fig. 1) [2].

Ganymede. Two alternative density models of an outer shell are considered [1]. Model (A) - an outer shell is completely composed of the high-pressure ice phases (no water is present), resulting in a maximum in the density of an outer shell. Model (B) - in the three-layer model of an outer water-ice shell, we assume that below a shell of ice I (30-120 km thick), a liquid layer of 230-140 km thick may exist, resulting in a minimum density of an outer shell. We adopted a “conductive” model where a mixed layer of water and high-pressure polymorphs of ice may coexist at depths between 260 km and an ice-rock interface. Our calculations show that the ice thickness of the outer shell in model (A) is about 890-920 km and in model (B) is 780-850 km (Fig. 2).

Callisto. We show that Callisto must only be partially differentiated into an outer ice-I layer, a water ocean, a rock-ice mantle, and a rock-iron core free of ice (mixture of anhydrous silicates and/or hydrous silicates + Fe-FeS alloy). Assuming conductive heat transfer through the ice-I crust, heat flows were estimated and the possibility of the existence of a water ocean in Callisto was evaluated. The liquid phase is stable (not freezing) beneath the ice crust, if the heat flow is between 3.3 and 3.7 mW m², which corresponds to the heat flow form radiogenic sources. The thickness of the ice-I crust is 135-150 km, and that of the underlying water layer, 120-180 km. The allowed total (maximum) thickness of the outer water-ice shell is up to 270-315 km. The results of modeling support the hypothesis that Callisto may have an internal liquid-water ocean (Fig. 3) [2].

Schematic FeO/Si atomic ratios for the terrestrial planets, meteorites and Galilean satellites are shown in Fig. 4. A comparison of the internal structure of the Galilean satellites and Titan has been made.


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Fig. 1. Thickness of the water-ice shell of Europa.

Fig. 2. Thickness of the water-ice shell of Ganymede.

Fig. 3. Internal structure of Callisto with a subsurface ocean. A layer of a mixture of high-pressure ices and rock-iron material (rock-ice mantle) must exist between the outer ice-water shell and the rock-iron core. The vertical line corresponds to the thickness of ice-I crust of 150 km. The permissible thickness of the underlying internal ocean is about 120-180 km. The maximum radius of the rock-iron core is 950 km.

Fig. 4. Schematic Fe/Si atomic ratios for the terrestrial planets, meteorites and Galilean satellites. Geophysical and geochemical constraints show that the bulk compositions of the rock-iron cores of the Galilean satellites are similar and may be described by the composition close to the L/LL chondrites.
SURFACE VARIATIONS OF PHASE FUNCTION STEEPNESS FOR TWO LUNAR SITES FROM SMART-1 AMIE DATA. V. Kaydash1, M. Kreslavsky2, Yu. Shkuratov1, S. Gerasimenko1, P. Pinet3, S. Chevrel4, J.-L. Josset5, S. Beauvivre6, M. Almeida7, B. Foing8, 1 Astronomical Institute of Kharkov University, Sumskaya 35, Kharkov, 61022 Ukraine. kvg@vk.kh.ua, 2University of California, Santa Cruz, CA, USA, 3UMR 5562/CNRS/ Toulose III University, Midi-Pyrenees Observatory, 14 Av. E. Belin, 31400 Toulouse, France, 4Space Exploration Institute (CH-2002 Neuchâtel, Switzerland) 5ESA/ESTEC (Keplerlaan 1, 2201 Noordwijk, The Netherlands)

Introduction: We use images obtained in 2006 by Advanced Moon Micro-Imager Experiment (AMIE) camera onboard SMART-1 spacecraft to access photometric properties of two lunar areas in the context of the regolith microstructure. We map the steepness of phase function for these sites and find local anomalies of the parameter. These anomalies are discussed in terms of lunar surface geology.

Model of lunar photometric function: The photometric function of the Moon describes the dependence of lunar surface brightness on the incidence \(i\), emergence \(e\) and phase \(\alpha\) angles. We use an approximation of this function proposed by Akimov [2], which was successfully applied to Clementine data [1]. This approximation expresses the photometric function through the phase angle \(\alpha\), photometric latitude \(\beta\) and longitude \(\lambda\). The first multiplier is the phase function dependent solely on \(\alpha\); the second is the disk function describing the orientation of the scattering surface to the Sun and observer for given \(\alpha\):

\[
F(\alpha, \beta, \lambda) = \exp(-\eta\alpha) \cos(\alpha/2) \times \\
\frac{1}{1-\sin(\alpha/2)} - \frac{\sin(\alpha/2)}{\sin(\alpha/2)^{\nu+1}} (\cos \beta)^{\nu}
\]

This formula contains only two adjustable parameters, the parameter of disk function \(\nu\) and the steepness of phase function \(\eta\). Thus we apply this description to SMART-1 data in order to map and analyze the photometric function parameters.

AMIE photometric data processing: Several lunar sites were selected for target-tracking campaigns during the SMART-1 mission. These targets were imaged in wideband spectral area of the AMIE micro-imager with 100-200 m / pix resolution. We converted raw counts of images to the values proportional to the bidirectional reflectance under given observational geometry with the preliminary pipeline calibration [3] which applies Master Flat fields and Dark frames. This procedure is sufficient to make a normalized phase dependence of brightness using multiangular AMIE data. Calculation of photometric angles (\(\alpha, \beta, \lambda\)) for each image and spatial transformations of images obtained at different geometries to the common projection were performed using the latest versions of AMIE SPICE kernels [4]. Residual misregistrations of images were eliminated with an autocorrelation procedure for subpixel transformation of images. The local surface tilts (surface topography) disturb the values of \(\beta\) and \(\lambda\); we neglect this effect applying our method to flat areas only (mare, crater floors etc.). Knowing the reflectance and photometric angles, we apply the least squares fit procedure to find the parameter \(\eta\) in Eq. (1). We run this algorithm several times for different fixed values of the parameter \(\nu\) and found that the disk function varies slightly for a wide range of \(\nu\), therefore, it does not affect the spatial pattern of \(\eta\) affecting only the absolute values of \(\eta\). Finally, we adopt \(\nu = 0.3\) in accordance with previous studies [5].

Gruithuisen domes: The first area (centered at 39.5 W, 35.8 N) we studied covers Gruithuisen domes and surrounding mare in the western part of Mare Imbrium. Five successive images were obtained in the SMART-1 orbit 2236 with photometric angles \(i = 45^\circ\), \(e = 4-30^\circ\), \(\alpha = 30-70^\circ\) and 110-160 m / pix spatial resolution. Fig. 1a presents a part of an image of this area (north is up), a map of the parameter \(\eta\) is shown in Fig. 1b. Major apparent variations of the steepness parameter are associated with surface tilts and look like illuminated topography in Fig. 1b. The map also reveals detectable true photometric anomalies associated with small mare craters where previous spectral analysis indicates the presence of MS2-like regolith [6]. Arrows \(i\) in Fig. 1 mark two craters with high \(\eta\) values (“positive” photometric anomaly) and arrows 2 point to three craters with diffuse extended halos of low \(\eta\) (“negative” anomaly); these halos are not seen in albedo. Similar crater-related negative anomalies were observed in other sites with Clementine data [5]. The negative anomaly for distal ejecta areas may be explained by disturbing the “fairy-castle” microstructure of the regolith by the impact event. The local regolith modification can produce a less porous layer with suppressed shadow-hiding effect [5]. We interpret the positive anomaly for craters \(i\) as an increase of mesoscale roughness in the proximal ejecta zone, making the phase function steeper. This roughness can be due to the presence of an anomalously large number of boulders and blocks. Such photometric anomalies were observed in areas studied in [5]. Large-scale subtle variations of \(\eta\) over the mare surface (outlined circle in Fig.1b) may be explained by the presence of more fine-grained pyroclastic material in the vicinity of volcanic domes.
Crater Lavoisier: Another spot-pointing AMIE campaign was carried out for the orbit 2251; 33 shots of the cracked-floor crater Lavoisier (80.8 W, 38.2 N; Fig. 2a) were made. Photometric angles are $e = 0$-$45^\circ$, $\alpha = 26$-$80^\circ$; while $i = 45^\circ$, the resolution varies from 110 to 200 m/pix. The $\eta$ parameter map in Fig. 2b highlights the absence of low albedo areas at the periphery of the crater floor (cf. Fig. 2a). These dark lava flows, compositionally different from the crater floor material, have however the same regolithic microstructure as a result of space-weathering processes, and thus do not show up in the $\eta$ map (Fig. 2b). Two more examples of extended negatively anomalous halos around young small craters are shown with arrows in Fig. 2. The most puzzling feature is the large negative anomaly outlined by the circle around the large craters (4.6 and 5.6 km in size). It might be caused by pyroclastic deposits [7] associated with tectonic fractures across the floor.

Conclusions: AMIE preliminary studies reveal new photometric anomalies of the lunar surface detected at a typical 100 m/pix resolution. These examples demonstrate the interest of orbital lunar photometry for characterizing the regolith microstructure; it illustrates both the usefulness of spot-pointing imaging observations and the scientific value of the AMIE data set in documenting geological processes associated with lunar regolith reworking.

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