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THERMODINAMIC CALCULATIONS OF VENERA/VEGA MAGMA CRYSTALLIZATION TRENDS USING COMAGMAT SOFTWARE: PRELIMINARY RESULTS. A. M. Abdrakhimov, Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Russia, Kosygina St. 19, 119991, Moscow, albert@geokhi.ru.

Introduction: Soviet landers Venera 8, 9, 10, 13, 14, and Vega 1 and 2 made geochemical analyses of Venus' surface materials [1] and this is the only direct geochemical information about Venus' surface. For Venera 13, Venera 14 and Vega 2 surface materials major petrochemical elements were measured by XRF (without Na$_2$O). All studied materials are close in their chemistry to terrestrial basalts. The goal of this work is to trace the trend of equilibrium crystallization of Venus magma and compare with those for terrestrial oceanic from three tectonic magma generation environments.

Data and results: The Venus material analyses were recalculated to obtain sulfur-free compositions. The thermodynamic software COMAGMAT was used for simulation of equilibrium crystallization [2]. The following parameters have been assumed: QFM oxygen buffer, 0% H$_2$O, isobaric crystallization. The simulations run for different sodium abundance.

The results of modeling are shown in Fig.1. 

Venera 13 magma crystallization is characterized by high-potassium trend, which strongly differs from the trends of the mid-ocean ridge magmatic series and from the ocean island arc magmatic series trends, and close to the ocean hot-spot trends.

Vega 2 magma evolution trends are characterized by very low enrichment in titanium, which differs from the ocean hot-spot magmatic series trends. Vega 2 trends differ from those of MORB in high Al$_2$O$_3$ and low CaO abundance. They are more close to ocean-island arc tholeitic trends, especially for the magnesium varieties, such as boninites.

Implication: The Venera 14 magma melt is close in its composition to MORB tholeites, typical series for the spreading tectonic environments. Venera 13 magma is close in its composition to the alkaline magmatic series of the terrestrial hot-spot tectonic situation.

Unexpected result of this simulation is that Vega 2 magma is rather close to the tholeitic series of en- simatic Ocean Island Arcs, the typical terrestrial recycling environment. But this may also be due to potentially large errors of the Vega 2 measurements: sum of components in the published analysis is significantly smaller than 100%. The Magellan mission data showed no evidence for global plate tectonics on Venus, including absence of the island arc-like geologic structures. It is possible that Vega 2 magma generation could be due to recycling mechanics related to corona formation [6].

Introduction: Processes of frost cracking on the Earth are widely distributed both in a zone of distribution permafrost rocks, and in an active layer. As result of the process the net of the polygonal relief is widespread on the different geomorphologic surface of the permafrost areas. The size of the features varies from first meters to 70-100 m in cross depending of many climatic and lithological factors [1,3]. In the terrestrial arctic zone the features represent the main morphological indicators of water ice presence in the rocks and sedimentary deposits [2]. Recently wide spreading of the polygonal landscapes have been found in the high-latitude regions of Mars due to the high-resolution imaging of Mars (1.4+10.0 meter/pixel) by the Mars Observer Camera (MOC) boarded on the spacecraft “Mars Global Surveyor”[4]. Studying of the Martian polygonal forms of a relief in comparison with morphologically similar forms on the Earth can help more precisely to understand the frost processes occurring on Mars.

Observation: It was shown [3] that in many case the polygonal relief on Mars is located in the interior parts of the impact craters. As object of our research one of such craters (MOC image R1104544) has been chosen. The diameter of the crater is about 2800 meters and it is located in near-polar area in the southern hemisphere (fig. 1).

The crater is characterized by presence of two-generation polygonal net (the first - large, the second – fine within the first one). Three zones have been allocated within the crater floor based on the morphological distinction in a polygonal net of the first generation. The first zone located in a peripheral part of the crater’s bottom is characterized by distinct orthogonal crossing of cracks resulted to rectangular shape of polygons were generated. The zone represents a ring belt with a width from 400 up to 600 meters and length about 8000 meters. The second zone settles down closer to the center of a crater and differs from first one by more irregular shape of the polygons, and also prevalence of three-beam crossings of cracks. It, as well as the first zone represents a ring in width from 750 up to 1250 meters and length about 4000 meters. The third zone is a center of a crater floor with radius 150-200 meters. Polygons here are fragmentary and less distinct due to overlapping by their young sediments, or their erosion. Their exact sizes cannot be established because of the limited resolution of the picture (1.5 meter/pixel).

For detailed morphometric and the statistical analysis of the polygons the quarter sector of the crater (see fig. 1) is chosen. During researches the net of a polygonal relief has been outlined and the diagrams of the polygons size distribution were constructed (fig. 2).
The analysis has shown that for the first zone the disorder of polygons makes from 45 up to 172 meters at mean value in 93 meters.

For the second zone the disorder of the sizes of polygons is characterized by wider range and has made from 35 up to 289 meters at mean value in 116 meters. As seen from fig.2, both zones are characterized by similar character of the polygons distribution.

For definition of the sizes of smaller polygons the analysis of larger ones has been made and some well defined large polygons were chosen in ring zones (fig. 3).

![Image](image-url)

**Figure 3.** Images and sketches show the net of the second generation polygons. a) – from the first zone, b) – from the second zone.

It was found that the size ranges of the polygons of second generation in both ring zones are very similar: from 3 up to 14 meters in the first zone (in the size of 7 meters prevail) and from 2 up to 13 meters in the second zone with prevailing size 6 meters (fig.4).

![Image](image-url)

**Figure 4.** Diagram of size distribution for the polygons of the second generation.

Also average sizes of disclosing of cracks have been determined. For first, larger generation, the sizes make from 3 up to 8 meters. For the second generation the sizes of disclosing of cracks are close to the resolution of a picture and matter up to 1.5 – 2 meters.

**Discussion:** Conducted analysis shown that the net of the polygonal relief within the crater floor have distinct circular-radial shape. At that, in the peripheral belt of the net the sharp rectangular polygons are dominated while from periphery to the center of the crater floor the polygons characterized by more irregular shape and here three-beam crossing of cracks starts to prevail above four-beam.

Formation of the similar concentric-radial zoning of the polygonal net is very typical in terrestrial conditions for the surfaces of the meanders and closed thermokarst depressions-alases. However it is not yet clearly whether studded two generations of cracks on Mars represent consecutive cracking polygons with formation of smaller forms (i.e. influence of process of cracking in time) or fine ranges are the reason of presence of a upper layer of younger sediments which material lithologically differs from a underlying material.

For understanding of leading factors of formation of two-generation polygons on Mars we plan to increase number of researched craters for carrying out of the similar analysis in various geomorphologic conditions.

**Reference:** 1- Kuzmin R.O. et al., (2002), 33rd LPSC . #2030; 2-Kudryavcev V.A. and Dostovalov B.N. (1978), The general permafrost. Moscow State University Press; 3-Kuzmin R.O., Zabalueva E.V., (2003), 34th LPSC. #1912; 4-Malin M.C., Edgett K.S., (2001), J.G.R., 106(E10): 23429-23570; ACKNOWLEDGMENTS This study was supported by the Russian Foundation for Basic Research (project no. 03-02-16644 and 04-05-65110).
THE PATERAE OF HELLAS VIEWED BY THE MEX HRSC: AMPHITRITES PATERA, PRELIMINARY RESULTS. M. Aittola¹, V.-P. Kostama¹, J. Korteniemi¹, J. Raitala¹, G. Neukum² and the HRSC Co-Investigator Science Team. ¹Planetology group, Univ. of Oulu, P.O. Box 3000, Oulu, Finland, <marko.aittola@oulu.fi>, ²Inst. of Geosciences, Freie Universitaet Berlin, Germany.

Introduction: Amphitrites Patera is located on the ridged plains of Malea Planum, to the south of Hellas basin [1,2] (Fig. 1). It was interpreted to be of volcanic origin on the basis of the Viking and Mariner 9 data [3,4], and that has been established by the later data sets [5]. The Patera is 121 km across and the elevations around its margin are ~1.7 km [5]. The nearby Peneus Patera is locating just 250 km west from the center of Amphitrites Patera. These two features combine a volcanic complex, which is surrounded by ridged plains and the channels of the Axius valles in the north. The ridges have several hundreds of meters relief and they appear to be oriented radially to the Amphitrites, in general [5]. To the south of Amphitrites there is a complex impact crater, Bernard, which is 120 km in diameter. It has been suggested that the shield of Amphitrites Patera is mainly made up of pyroclastic rock due to the broad form, low relief, and heavily dissected and degraded morphology [1].

Observations from the HRSC data. Different channels of the HRSC camera offer us a possibility to define the surface materials with different properties. When using red- and infra-red channels it is easy to recognize two different material units within the caldera of Amphitrites Patera (Fig. 2). The darker material is located in the western part of the caldera. The MOLA as well as HRSC stereo data shows that the area of this dark deposit is the deepest part of the caldera. Thus, the dark unit may reflect to the deposit accumulated to the caldera floor. The material could hence be for example of aeolian origin. The other caldera unit is seen as little brighter in the red channel and it covers the eastern part of the caldera floor (Fig. 2.). This unit clearly differs from the material of the caldera rim and the flank. Furthermore, the material of the unit seem to intrude to the fractures and the canyons of the rim thus superposing the rim of the patera. Also the northern and southern flanks of the patera show very different characteristics in the red channel, the northern flank being bright and southern dark. On the other hand this characteristic may have been affected by the illumination geometry. In green channel the area to the south of caldera seem to differ notably from the materials of the caldera and of the northern caldera flank. Same area is seen dark also in red channel, but not so well defined. The blue channel image is mostly featureless, showing a white NW-SE trending stripe on the top of caldera indicating most probably the existence of cloud layer.
The higher resolution nadir image of MEX-HRSC reveals a few examples of small interesting features—possibly evidence of later volcanic activity. On the south-eastern rim of the caldera there is a feature which basically looks an impact crater (Fig. 3 right). However, the western side of its floor displays a distinct depression. Furthermore, the flow-kind of feature is originating from the crater from the side of collapse. This could suggest that the flow has a volcanic origin, perhaps driven by the impact event. Another possible mark of minor volcanic activity is seen in Figure 3 (left), which presents an example of possible mesa inside the Amphitritae caldera. The third example is presented in Figs. 4-6, which display a possible example of a small volcanic edifice, with lava flows very well identifiable in the HRSC green channel image. They differ from impact crater ejecta blankets on the area. The same edifice can be seen also in Themis IR image.

Figure 3. The image on the left shows the mesa structure inside the caldera and the image on the right shows the crater with flow-like feature on the south-eastern rim of the patera. Both images are HRSC nadir images.

Figure 4. The HRSC green channel image of the possibly volcanic edifice on the north-western rim of the patera.

Figure 5. The HRCS nadir image of the same structure than in Fig. 4.

Figure 6. The Themis IR image (I07682003) of the possible volcanic edifice presented in Figs. 4 and 5.

Conclusions: The new HRSC data opens advanced new possibilities to study the volcanoes on the Martian surface. In the case of Amphitritae Patera, already a very preliminary study of the multi-channel data reveals different surface properties within the caldera as well on the flanks of the caldera. In addition, the nadir image displays evidences of small scale volcanic activity postdating the formation of the patera. The further studies of the MEX HRSC data together with the other new data sets will give us great opportunity to study the relatively unknown old Martian volcanic features in the southern highland.

Acknowledgements. We gratefully acknowledge the efforts made by the MEX-HRSC Photogrammetry Team in processing the digital image data.

COSMOGENIC RADIONUCLIDES IN METEORITES, GALACTIC COSMIC RAY MODULATION, SOLAR ACTIVITY AND CLIMATE OF THE EARTH.
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Introduction: The processes on the Sun, e.g., the solar activity, have an impact on the processes in the heliosphere due to formation and disturbance of the interplanetary magnetic fields (IMF). The most striking example is the solar modulation of galactic cosmic rays (GCR), i.e., the GCR intensity changes at the different heliocentric distances and at the different heliographic latitudes, in accordance with the variations of the solar activity. Thus, it is clear that the GCR intensity may be considered as a subtle tool for the study of electromagnetic structure of the heliosphere and its changes conditioned by the solar activity. However, for such investigations the long series of uniform data (over a number of 11-year solar cycles) on the GCR intensity in the interplanetary space are required. At present only the natural detectors of GCRs – cosmogenic radionuclides with different half lives $T_{1/2}$ in the chondrites fallen to Earth in 1959-2000 – provide such data on the GCR intensity at ~2-4 AU over ≥4 solar cycles [1]. The most valuable detectors are $^{54}$Mn ($T_{1/2}$=300 days), $^{22}$Na ($T_{1/2}$=2.6 years) and $^{26}$Al ($T_{1/2}$=7.4-$10^5$ years), which provide information on the average GCR intensity along the chondrite orbits for ~450 days, ~4 years and ~1 million years, respectively, before the fall of the chondrites to the Earth. Such an averaging smooths essentially the temporal, as well as the spatial variations of GCRs along the meteorite orbits, deriving the most important regularities [2].

Correlation analysis of the solar activity and GCR variations in the heliosphere: Virtually, the structure and dynamics of the IMFs are determined by the structure and dynamics of the magnetic fields on the Sun (SMFs). The well-known 11-year solar cycle is only a half of the 22-year magnetic solar cycle, when after two successive reversals the polarities of the north and south hemispheres of the Sun return to their initial states. However, there are, apparently, solar cycles of longer duration: 80-year, or secular cycles; 600-year cycles; etc. [3]. It is clear that their presence must be expressed as some violations of the ordinarily observed picture of the solar activity influence on the processes in the heliosphere. In this connection, the rigorous correlation analysis of distribution and variation of GCRs in the heliosphere, depending on different parameters of the solar activity, is of paramount importance. We have carried out such an analysis of the correlations of GCR gradients (meteorite data [1], GCR rigidity $R>0.5$ GV) with the different indexes of solar activity (the sun spot numbers [4] and the intensity of the green coronal line (GCL, $\lambda$=5303 Å) [5]), as well as with the inclination of the heliospheric current sheet [6] and with the IMF strength [7]. In Fig.1 the results of the correlation analysis of the GCR gradients and the GCL intensity at the polar angles $\theta=60^\circ$ and $\theta=120^\circ$ are presented. The curve for the GCR

![Fig.1](attachment:fig1.png)

Fig.1- (a) GCR gradient variations, according to the meteorite data (1) and variations of GCL intensity at $\theta=60^\circ$ (2) and $120^\circ$ (3); (b) time delays $\Delta t_1$ ($\theta=60^\circ$, solid line) and $\Delta t_2$ ($\theta=120^\circ$, dashed line) of the GCR gradient change from the GCL intensity change at the maximum correlation coefficients; (c) N-S asymmetry of the delays at $\theta=60^\circ$and $\theta=120^\circ$: $A_{NS}=|\Delta t_1-\Delta t_2|/ [\Delta t_1+\Delta t_2] \cdot 100\%$ (solid line); dotted line is $\pm 1\sigma$
gradients is doubly smoothed over experimental points for each 5 points by the polynomial of the first degree [8]. One can see that the GCR gradient variations are successively behind the GCL intensity changes. The average delay amounts to ~1 year, and it is about 3 times higher in 1970-1980 (especially in N-latitudes) at the correlation coefficients as high as 0.9. The average N-S asymmetry totals 17±10%, decreasing below ~10% in 1971-1982 (the positive phase of the magnetic cycle) and rising above ~20% in 1977-1986 (the negative phase of the magnetic cycle). Such an effect is conditioned by the different moments and durations of the solar polar magnetic field inversions in N and S hemispheres.

The strong growth of time delay of the GCR gradient variations is observed in the correlation analysis with all the other parameters: solar spot numbers, inclination of the heliospheric neutral current sheet and the IMF strength. Everywhere the 20th solar cycle (1965-1976) and the first part of the 21st solar cycle stand out sharply against the regularities in the other time intervals. That testifies to a deep disturbance of the IMFs during those periods, which could be conditioned by a peculiar transformation of the SMF structures, as compared with the commonly observed events.

Discussion: The cause of such strong violations, apparently, lies in replacing the solar cycle of much longer duration than 11-year cycle. First of all, it may be a secular (80±50 y) solar cycle. Indeed, in Fig2a the secular cycles of the solar activity in 1700-2001 are presented. The secular curve is obtained due to smoothing the maximum average annual numbers of the solar spots [9]. One may see that the last change of the secular cycle happens just in the 20th solar cycle. The secular curve has a character of free harmonic oscillations, conditioned by cyclic variations of the depth of the convective zone of the Sun and reflected in the strong SMF transformations. The increasing character of the secular regression line is interesting. It may imply that an ascent of a still longer solar cycle (~600 y?) takes place. It is tempting to connect the observed global rise in temperature on the Earth [10] (see Fig2b) with such a growth of the solar activity. Indeed, the analysis of the regression lines in Fig.2a,b displays that the solar activity gradient is 0.22%/year, whereas the temperature gradient of the Earth low atmosphere is 0.0054%/year. The low temperature gradient is natural, for instance, due to many inertial processes on the Earth, firstly, due to the inertial processes in the world ocean. Therefore, the solar activity may be considered as one of the main factors exerting the influence upon the climate of the Earth.

![Image of Fig. 2(a) showing secular cycles of the solar activity in 1700-2001 (solid curve); Ri are maximum values of the Wölf annual average numbers smoothed by the Gleisberg method [9]; (b) temperature variations in the Earth low atmosphere over the last 150 years (solid curve); dotted regression lines are y=-203+0.17x (a) and y=265+0.0042x (b)]

ABOUT THE CAUSE OF ASYMMETRY OF PLANETS SURFACE ON MARS

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Many planets of Solar system, among them Earth, Mars, and Moon have obvious morphological asymmetry of a surface. Most brightly this asymmetry is expressed on Mars, surface of which is dismembered by means tectonic border at two hemispheres. There are the tectonics passive – southeast “continental” hemisphere, and the tectonics active northwest – “sea” hemisphere. Surface of a passive hemisphere is raised above a level of active hemisphere on 4-5 km. She is richly covered by large craters which have diameter of tens - hundreds km. Surface of a sea hemisphere is mainly represented by smooth plains. Quantity of her craters is not large [1]. Craters of continental and sea hemisphere differ not only by quantity, but by a different structure and formation time [2] and also by function of sizes distribution [3, 4].

The specified features of distribution of Mars craters may be explained by fall on planet of two different types of cosmic bodies [3]. The first are asteroids and comets of Solar system, and the second – comets of Galaxy. First bodies type is falling quite regularly all time and forming craters mainly in an equatorial zone of Mars. Second bodies type bombards a planet through each 20÷37 million years in rather short epochs (~1-5 million years) of Sun stay in jet streams and spiral branches of Galaxy [4].

Owing to inclination of the ecliptic plane to the galactic plane on corner 60°, the area bombarded by comets moves over planet surface with the period of orbital movement of Sun in Galaxy ~223 million years [4]. In consequence of modern position of Sun on galactic orbit three last bombardments of Solar system by galactic comets took place basically in southern hemisphere of planets. In particular, on Mars, as we suppose, it has resulted in abrupt asymmetry of his south-east and northwest hemisphere.

Galactic comets basically are consisted of water ice and other frozen gases. Comets nucleuses have density ~1 g/sm³, and their diameter, weight and energy accordingly are 0.1÷3.5 km, 10⁴÷10¹⁴ kg and 10²⁰÷10²⁵ J [7]. During one bombardment ~10⁴÷10⁶ galactic comets may fall on Mars and Earth. In result on planets without a gas environment or with very rarefied atmosphere as at Mars, a surface sites bombarded by galactic comets completely are sated with large craters. The craters density here reaches theoretical limit is 100 craters in diameter of ≥10 km on area 1 million km² [3].

In gas environments of planets a nucleuses of galactic comets intensively evaporate. They lose of mass and are decreasing in size [7]. As a result of this process a comets craters even on Mars with his strongly rarefied atmosphere have diameters systematic less, than on Moon completely deprived of gas atmosphere (fig.).

Fig. Comparison of integrated craters distributions for continental hemispheres of Mars and Moon completely sated by comets craters. Distributions of craters are constructed on the data [6].

The differences in craters distributions of Mars and Moon in the field of the big sizes can be explained by reduction of a comets
nucleuses diameter approximately on 300 m owing to their evaporation in a Martian atmosphere. Surplus on Mars of small craters in comparison with Moon is produced by destruction of a comets nucleus in an planet atmosphere [7].

Much more dense, than at Mars, gas environments of Earth and Venus are practically impenetrable for galactic comets [3]. In dense atmosphere of these planets galactic comets inevitably perish with formation of a hypersound shock wave to which the basic energy of galactic comets is given. At this case after of achievement by a wave of solid planetary surface a huge kinetic comet energy goes not on creation of a crater, and may be spent mainly for warming up and fusion of big local zones of rocks deeply under impact place [7]. On the Earth these zones directly display themselves as magmatic cameras. Under thin ocean plates such cameras may exist during of hundreds millions years. This cameras deliver big volumes of a lava on surface of modern ocean bottom and participate in construction of submarine mountains [8].

Mars in this respect, obviously, differs from Earth and Venus. The big part of comets energy here is spent for craters formation, and her essentially smaller part is transferring to shock wave. Besides in conditions of Mars a influence of shock waves on a lithosphere, apparently, is not so concentrated as on Earth and on Venus and a shock waves energy basically goes into heating Martian astenosphere. Thus laves outpouring on a Martian surface, as supervision testify, has the subordinated character and obviously concedes to scales of this phenomenon on Earth and Venus.

Whereas a falls of comets are very intensive, and a conduction cooling of an astenosphere has very small speed, this should create strong swell of Martian surface which has undergone comets bombardment.

According calculations which are similar executed in [8], for an explanation by this mechanism of an difference in heights of continental and sea hemisphere of Mars, it is necessary to assume that astenospheric layer of significant thickness should be under continental hemisphere. Thickness of this layer may be ∼100-250 km in dependence on a degree of his substance smelting. This estimation coincides with size of a layer astenosphere under continents of the Earth [9]. Occurrence on Mars such astenosphere may be connected to his bombardments by galactic comets in cenozoic.

So at moving the Sun in Galaxy comets bombard serially [10] as southern as well as northern hemisphere of Mars, the astenospheric layer may migrate in a body of a planet, both its continental and sea hemisphere through half of galactic year may exchange place.

This opportunity pulls together Mars and Earth especially, because on our planet continental plates gather periodically together, forming supercontinents near South Pole [11], exceptionally into epochs when comets bombard her southern hemisphere.

Tidal and shell-dynamics interactions of the given celestial body with external celestial bodies lead to variations of their tensional state and to variations of different planetary processes including variations of seismic activity. It is clearly observed that variations of lunar seismicity have the celestial mechanical nature and depend from the Moon perturbed orbital motion. Using dynamical analogy in translatory-rotary motions of synchronous satellites and Mercury we have obtained evaluations of periods of variations of seismic activity of the Titan and Mercury. High level of endogenous activity of Titan was predicted earlier [1]. The full elastic energy of luni-solar tides superposition is not additive sum of elastic energies of separated tides and contains additional terms of mutual character, which play significant role in geodynamical life of the Earth [2]. Tensional state of the Earth is characterized by the elastic energy stored in superposition of tides. Part of elastic energy dissipates and goes to warm energy and to an energization of different geodynamical processes in definite rhythms. Correlation of the extreme variations of the elastic tidal energy of the Earth with earthquakes and moonquakes events (in period 1971-1976 years) was established [3]. This regularity of seismic process have used for prediction of the dates of some large earthquakes in 2003 and 2004 years. In particular the date of phenomenal Hokkaido quake of 25 September (M=8.3) was predicted with high accuracy in [3]. Mechanism of differential and cyclic action of the Moon and Sun on the interacting plates has been suggested (as trigger mechanism) for explanation of discovered regularity.

Cyclicities of the Moon seismicity. The relative orbital motions of the Moon and the Earth are identical. Periodicities in their orbital motions in same style influence on the tidal processes on both celestial bodies and, consequently, rhythms at identical periods can be expected in seismic processes for the Moon and for the Earth. First confirmation of mentioned correlations has been obtained for shallow earthquakes (with magnitude >7.3) and shallow moonquakes in period 1971-1976 [4]. Tidal nature of moonquakes has been discussed by an interpretation of results of their spectral analysis [5], [6]. In mentioned papers the periods of the Moon orbital perturbations in 27.4, 13.6, 206 days and some others have been determined. Our spectral-temporal analysis of the full series of the deep moonquakes from catalogue which has been kindly presented us by Y. Nakamura (7344 events, [7]) have been let us to confirm mentioned periods and to establish some new cyclicities in quake activity of the Moon and to describe a fine structure of some from orbital periodicities. So were determined variations of Moon seismicity with another periods multiple to orbital draconic (T_{drac}), anomalistic (T_{anom}) and synodic (T_{synod}) periods. From another the more remarkable and observed periodicities of the Moon activity are (in days):

<table>
<thead>
<tr>
<th>Periods</th>
<th>Periods (in days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>T_{syn}/1=6.898</td>
<td>6.871 (289); 3.9864 21.992</td>
</tr>
<tr>
<td>T_{drac}/5=3.442</td>
<td>3.470 (249); 1.9932 10.996</td>
</tr>
<tr>
<td>T_{anom}/5=5.508</td>
<td>5.504 (261); 3.1891 17.594</td>
</tr>
<tr>
<td>T_{syn}/8=3.692</td>
<td>3.735 (236); 1.9961</td>
</tr>
<tr>
<td>T_{synod}/7=4.219</td>
<td>4.205 (234); 2.2813</td>
</tr>
</tbody>
</table>

From another the more remarkable and observed periodicities of the Moon activity are (in days):

441.800 [436.9 ± 8.7], (819);
366.9 [347.5 ± 11.3], (661);
206.6 (847);
441.8 [436.9 ± 8.7], (819);
227.9 (564);
299.5 (515);
258.5 [250.2 ± 3.6], (393);
148.9 (484);
81.73 [81.8 ± 0.4], (360);
61.13 (324);
106.8 [107.5 ± 0.9], (305);
101.3 (295);
68.13 (291);
138.1 (280);
178.3 [178.1 ± 2.3] (274);
161.1 (271);
117.3 (256);
58.78 (242);
34.300 [34.5 ± 0.1], (311);
33.575, [33.6 ± 0.1] (268);
32.75 (500);
31.950 [32.0 ± 0.1], (424);
14.05 (462);
30.024 [30.6 ± 0.1], (344);
25.20 (440);
25.91 (420);
16.495 [17.9 ± 0.1], (240);
13.13 (348);
6.561, [6.5 ± 0.1] (263),
± 5.710 [5.7 ± 0.1] (250);
± 3.395 [ 3.3 ± 0.1],
± 3.505 [3.5 ± 0.1], (259) and others.

Here in square brackets are given values of periods obtained independently by Dr. Kaftan V.I. and in parentheses are given conditional amplitudes of corresponding seismic variations. Obtained results were analyzed and compared with similar results obtained for a random temporal distribution of quakes. The main conclusion is: the seismic rhythms on the Moon have the celestial-mechanical nature and are dictated by gravitational influence of the external celestial bodies. Although, the spectral-temporal analysis has revealed some temporal instability of some rhythms observed [6]. As known the many from the Earth processes are characterized by a similar behavior. To explain observed cyclicities in the moon seismicity we study a possible role of the orbital and rotational tides and a role of mechanical interaction between non-spherical mantle and core of the Moon induced by gravitational action of the Earth and the Sun.

Variations of the Earth seismic activity. We have been fulfilled statistical analysis of differences of dates of big earthquakes (in last 30 years, [9]) and close dates of extremes of tidal elastic energy. Obtained results in general confirm correlation of these dates and a new phenomenon of displacements of dates of the big quakes on 1.5-2.0 days with respect to the dates of extremes of elastic energy has been observed.

We can dear to suggest that variations of elastic tidal energy exert some control on seismic processes. It is important to note that the crossed term of the elastic energy plays a relevant role in the observed correlation with seismic events. It seems natural: part of the elastic energy accumulated with every orbital cycle of Moon (and Sun) dissipates to inner geodynamical processes that its variation drives.

For explanation of observed regularity of a seismic process we have been suggested a mechanism of differential and cyclic action of the Moon and Sun on the interacting plates. This mechanism is a trigger mechanism which control big seismic events in different time-scales.

Cyclicities of Titan seismicity. Titan is synchronous satellite of Saturn with parameters of orbital motion similar to Moon motion. In assumption that the mechanisms of endogeneous activity of Titan and their displays are identical with the Moon we can give some first evaluations of periods of variations of Titan processes including its seismic activity. These periods are given in analogy with observed lunar variations of seismicity (in Table 1). Here we have used known model of the Titan orbital motion [8].

Cyclicities of Mercury seismicity. The translatory – rotary motion of Mercury is also resonant. In assumption that the mechanisms of endogeneous activity of Mercury and their displays are identical with the Moon we can remark on the possible variations of Mercury processes including. So must be observed variations of the seismic activity of Mercury with periods multiple to period of its orbital motion 87.969 days (some from them are given in Table 1).

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References


NEW VIEW OF OLYMPUS MONS BY THE MARS EXPRESS HRSC CAMERA. A. T. Basilevsky1,2, G. Neukum2, B. A. Ivanov2,3, S. C. Werner2, S. van Gasselt2, J. W. Head4, R. Jaumann5, H. Hoffmann5, E. Hauber5, and the HRSC Co-Investigator Team. 1 Vernadsky Institute of Geochemistry and Analytical Chemistry, RAS, 119991 Moscow, Russia; 2 Institut fuer Geologische Wissenschaften, Freie Universitaet Berlin, 12249 Berlin, Germany; 3 Institute of Dynamics of Geospheres, RAS, 119334 Moscow, Russia. 4 Dept. Geol. Sci., Brown University, Providence, R.I. 02912 USA; 5 DLR-Institut fuer Planetenforschung, Berlin, Germany.

Introduction. The High Resolution Stereo Camera (HRSC) images [1] taken on orbit 143 yield a measurable high-resolution view of the western part of Olympus Mons providing a new insight into the geology of this structure. Since the early 70’s, it has been considered as a volcanic construct composed of accumulations of mafic lavas [2,3]. The HRSC-based analysis, suggests a more complicated geologic history with the interplay of volcanism and several other processes [4,5].

Morphology. There are three major morphological elements of the region (Figure 1): 1) the western part of the Olympus Mons summit, hereafter summit plateau; 2) its western slope subdivided into three types; and 3) lowland plains with pieces of the Olympus Mons aureole [6,7].

Summit plateau. The HRSC-based DTM shows altitudes within the summit plateau from +7.2 to +9.7 km. The plateau is bordered on the west by steep (20-30°) to gentle (4-10°) slopes 6-8 km high. In the southern and central parts of the summit plateau, lava flows are numerous (Figure 2) while in the northern part they are rarely observed. The impact crater frequencies show that the lava-covered areas vary in age from 20 to 450 m.y. with one subarea being as young as 2-3 m.y [4]. The plateau surface also shows numerous brighter and darker wind streaks (Figure 1).

Close to the plateau edge there are several mesas composed of layered deposits, at least in their upper parts. Figure 3 shows a mesa having at its north the 1.4 by 1.7 km oval pit. The pit is steep-sloped but without an elevated rim and ejecta blanket suggesting a non-impact, probably collapse-type origin. The topographically depressed part of the mesa (image center) is embayed by lava flows from the NE. Embaying the mesa, they become surrounded by local troughs as if the mesa material collapsed [8,9]. Crater frequency age of mesas is close to 3 b.y. [4].

The northern part of the summit plateau edge is rimmed by a ridge (hereafter ridge) standing up to 0.3-0.5 km above the adjacent plateau. Its western slope is the upper part of the plateau scarp dissected by ravines (Figure 4). The ravines look similar to those cut by water on steep slopes of Earth (e.g., in Death Valley) and to ravines on steep tectonic slopes on Venus where running water is ex-
The eastern slope of the ridge has an unravined surface, sometimes showing the layered structure.

In one of the gaps of the edge ridge at an altitude above +7 km, a flow-like feature is observed resembling a glacier and suggesting glaciation on the volcano top. It flowed ~200 m.y. ago [4] to the S and SW, i.e., not down the Olympus scarp, but inward toward the volcano (Figure 5).

The plateau western slopes can be subdivided into three morphological types (Figure 1): 1) Type 1 slopes (Figure 4) have a steep (20-30°) layered upper part dissected by ravines, described above, while the lower part of the slope is gentle and smooth, looking similar to talus observed in the lower parts of terrestrial and Venussian ravined slopes.

Type 2 slopes are less steep: 8-12° in the upper and 3-5° in the lower parts. They often start with the areas of etched terrain, where sub-horizontal layers are locally seen (Figure 6).

Downslope or in direct contact with the summit plateau are observed irregular depressions several kilometers across (Figure 7) morphologically similar to “chaos” of the source areas of the Martian outflow channels [10]. Downslope-trending sinuous channels and chains of rimless craters are seen in association with them. The presence of all these features suggests ground ice melting, surface collapse and water run off, perhaps due to lava invasions. Locally at the lower parts of Type 2 slopes, tongue-like flows of the rock-glacier type are seen [4,5].

Type 3 slopes are generally the least steep (3-5° to 7-9°) and have no ravines, chaos-like depressions, channels, chains of rimless craters and tongue rock-glacier-like flows. It is covered with obvious lava flows, which continue for some distance into the lowland plains.

The lowland plains are generally below the reference level and display three varieties: 1) plains with practically no visible landforms at the very west; 2) plains with large glacier type flows below Type 1 and 2 slopes [14] (Figure 8); the glacier type flows show ages from 4-6 to 180 m.y. with one MOC-based value ~400 m.y.; and 3) plains with lava flows below Type 3 slopes.

Discussion and conclusion. The morphologies described imply several geologic processes. Mafic volcanism is responsible for formation of lava flows until recently (2-3 m.y. ago). Mesas with layers imply sedimentation of dust (maybe ash [11]) and ice. The presence of ice is implied from the deep rimless pits and from specific interaction of the mesa material with lava flows. The presence of chaos, channels, and the tongue flows on Type 2 slopes and the presence of glacier type flows below Type 1 & 2 slopes implies the presence of ground ice and episodes of glaciation [4,5,14] probably controlled by the planet spin-axis obliquity changes [13]. Eolian processes partly modified all of these landforms. So, the geologic history of Olympus Mons involved much more than just the accumulation of lavas.

THICKNESS OF THE OLYMPUS MONS LAVA FLOWS AS MEASURED FROM THE MGS MOC AND MOLA DATA. E. A. Bazilevskaya. Geological Department, Moscow State University, Vorobievy Gory 119992, Moscow, Russia.

Introduction. Here we report on the study of thicknesses of lava flows observed on the surface of Olympus Mons volcano (Figure 1).

Figure 1. Olympus Mons volcano as viewed from the north. Source: http://esamultimedia.esa.int/marsexpress.

This 600 km in diameter feature was discovered in the images taken in 1972 by Mariner 9 and soon classified as a shield volcano composed of accumulations of basaltic lavas [1]. Since then, this classification has been confirmed by many new missions to Mars [see, e.g., 2]. Lava flows which are the subject of this study are typically 10 to 20 km long and 200 m to 2 km wide streaming downslope in directions generally radial to the volcano center. We plan to make extensive study of these lava flows based on analysis of the data taken by the Mars Express High Resolution Stereo Camera as well as Mars Global Surveyor camera MOC and laser altimeter MOLA. Here we report on the first stage of this study, that is, the measurements of lava flow thicknesses.

Measurement technique. The lava flows studied are of rather small width so data with high spatial resolution are needed for their analysis. Unfortunately the available high-resolution images taken by MOC and recently by HRSC have been taken at high Solar elevation above the horizon (typically >60°), so the shadow technique often used in planetary science is practically not applicable in this case. Measurements using MOLA profiles crossing individual lava flows are possible but because the MOLA footprint is about 200 m across, this is comparable to the width of some flows and to significant elements of others; therefore, we decided not to use this approach. The HRSC stereo images provide reliable digital terrain models but in the routine technique its spatial resolution is also close to 200 m.

Therefore, we used a different approach: We have been searching for MOC images on which steep scarps of the volcano are seen. We used for measurements only those images where on the top of the scarp typical lava flows were observed and on the scarp itself layers obviously consisting of these and similar lava flows were visible (Figure 2).

Figure 2. Lava flows forming layers seen in the scarp. Fragment of MOC image R04-01474.

On these images we have measured the apparent thicknesses of the most prominent layers. For part of these images MOLA profiles are available so we could measure the slope steepness aiming to transform apparent thicknesses to true ones. However, as a rule, MOLA profiles were not oriented directly downslope but under some angle to it that demanded additional corrections. Moreover because MOLA profiling was being done only until the MGS MOC working stage E6, only part of scarps with the outcropped layers are characterized by MOLA profiles. So, after selecting images with visibly steep scarps with prominent lava flows on them we made a search for MOLA data for any visibly steep scarps within the Olympus Mons construct. Then we determined steepness of these slopes, including corrections when the MOLA profile was oblique to the slope (Figure 3), calculated mean value of the slope steepness and applied it to calculate true thickness for all our measurements.
Results. We have found on the available MOC images of Olympus volcano only 13 sites where prominent lava flows were seen on steep scarps and their thickness could be measured (Table 1):

Table 1. Results of measurements of lava flow apparent thickness (AT) and true thicknesses (TT).

<table>
<thead>
<tr>
<th>Situation</th>
<th>No. meas</th>
<th>AT meters</th>
<th>St. dev</th>
<th>TT meters</th>
<th>St. dev</th>
</tr>
</thead>
<tbody>
<tr>
<td>NE scarp</td>
<td>19</td>
<td>9.7</td>
<td>2.4</td>
<td>6.9</td>
<td>1.7</td>
</tr>
<tr>
<td>N scarp</td>
<td>10</td>
<td>7.2</td>
<td>2.5</td>
<td>5.1</td>
<td>1.8</td>
</tr>
<tr>
<td>N scarp</td>
<td>19</td>
<td>6.6</td>
<td>1.6</td>
<td>4.7</td>
<td>1.1</td>
</tr>
<tr>
<td>N scarp</td>
<td>8</td>
<td>5.6</td>
<td>1.1</td>
<td>4.0</td>
<td>0.8</td>
</tr>
<tr>
<td>N scarp</td>
<td>14</td>
<td>8.4</td>
<td>1.6</td>
<td>6.0</td>
<td>1.1</td>
</tr>
<tr>
<td>N scarp</td>
<td>5</td>
<td>4.2</td>
<td>1.6</td>
<td>3.0</td>
<td>1.1</td>
</tr>
<tr>
<td>N scarp</td>
<td>11</td>
<td>6.5</td>
<td>1.2</td>
<td>4.6</td>
<td>0.9</td>
</tr>
<tr>
<td>N scarp</td>
<td>14</td>
<td>6.6</td>
<td>1.3</td>
<td>4.7</td>
<td>0.9</td>
</tr>
<tr>
<td>N scarp</td>
<td>19</td>
<td>10.4</td>
<td>2</td>
<td>7.4</td>
<td>1.4</td>
</tr>
<tr>
<td>S scarp</td>
<td>11</td>
<td>7.4</td>
<td>1.6</td>
<td>5.3</td>
<td>1.1</td>
</tr>
<tr>
<td>S scarp</td>
<td>22</td>
<td>6.8</td>
<td>1.9</td>
<td>4.9</td>
<td>1.4</td>
</tr>
<tr>
<td>N scarp</td>
<td>12</td>
<td>10.2</td>
<td>1.8</td>
<td>7.3</td>
<td>1.3</td>
</tr>
<tr>
<td>SE scarp</td>
<td>12</td>
<td>11</td>
<td>2.3</td>
<td>7.8</td>
<td>1.6</td>
</tr>
<tr>
<td>Mean of means</td>
<td>7.7</td>
<td>5.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>St. dev of means</td>
<td>2.1</td>
<td>1.5</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The results of the slope steepness measurements are given in Table 2:

Table 2. Results of measurements of steepness of visually steep scarps using combination of MOC and MOLA data.

<table>
<thead>
<tr>
<th>MOC number</th>
<th>Situation</th>
<th>Steepness, degree</th>
</tr>
</thead>
<tbody>
<tr>
<td>M19_01926</td>
<td>Olymp SE scarp</td>
<td>38.7</td>
</tr>
<tr>
<td>M11_02262</td>
<td>Olymp N scarp</td>
<td>35.7</td>
</tr>
<tr>
<td>E04_00550</td>
<td>Olymp N scarp</td>
<td>25.76</td>
</tr>
<tr>
<td>M08_05053</td>
<td>Imp crat NW scarp</td>
<td>35.2</td>
</tr>
<tr>
<td>M08_05053</td>
<td>Imp crat SE scarp</td>
<td>29.7</td>
</tr>
<tr>
<td>M19_00402</td>
<td>Caldera SW scarp</td>
<td>49.94</td>
</tr>
<tr>
<td>M08_05053</td>
<td>Caldera SE scarp</td>
<td>32.76</td>
</tr>
<tr>
<td>M08_05053</td>
<td>Caldera NW scarp</td>
<td>34.8</td>
</tr>
<tr>
<td>E05_01829</td>
<td>Caldera SW scarp</td>
<td>33.8</td>
</tr>
<tr>
<td>E01_02040</td>
<td>Caldera NW scarp</td>
<td>38.6</td>
</tr>
</tbody>
</table>

Mean value 35.5
Standard deviation 6.4

It is seen from the Table that the scarp steepness does not depend noticeably on the geologic situation (collapse scarps of the volcano external slopes, summit caldera scarps, uppermost scarps of large impact crater) being probably just a repose angle. So we probably can use the mean (for the studied situations) value of the scarp steepness (α) to apply for recalculation of apparent thickness (AT) to true thickness (TT) by multiplying AT by tg α. The mean values for the sites (25 to 50°) and the mean of mean values (35.5°) are in good agreement with measurements for the Western scarp of Olympus Mons based on HRSC based digital terrain model [3].

Summary. Our work showed that mean values of the true thickness of the Olympus Mons lava flows, measured in the MOC images of steep scarps, vary from site to site from 3 to 8 m with the mean of means being 5.5 m. This is in a good agreement with the mean thickness of basaltic lava flows of Earth: 3-20 m [4].


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IDENTIFICATION OF LUNAR ROCK TYPES. A. A. Berezhnoy¹,², N. Hasebe¹, M. Kobayashi¹, G. Michael³ and N. Yamashita¹ ¹Advanced Research Institute for Science and Engineering, Waseda University, Tokyo, Japan ²Sternberg Astronomical Institute, Moscow, Russia ³German Aerospace Center, Institute for planetary research, Berlin, Germany

Abstract: The quality of Lunar Prospector measurements of elemental composition of the lunar surface is checked by comparison between remote sensing and returned sample collection data sets. In western maria Si content is underestimated, but Mg content is overestimated by Lunar Prospector. Petrologic mapping of the Moon with usage of Lunar Prospector Mg, Al, Fe abundances is performed. Relative content of end-members as mare basalts, ferroan anorthosites, Mg-rich rocks is estimated. Special technique for identification of unusual rock types is developed by analysis of distances of Lunar Prospector pixels from end-member plane.

Introduction: Elemental mapping of the lunar surface is very useful technique for study of petrologic provinces on the Moon, for search for chemical anomalies sites and ancient cryptomaria, and for identification of lunar rock types. Global mapping of Th and Fe content on the Moon was conducted using low resolution Lunar Prospector gamma ray spectra [1]. Preliminary data about abundances of other elements as O, Si, Mg, Ca, Al, K, U, and Ti are presented also [2].

In this work we analyze the quality of Lunar Prospector gamma ray spectrometer data comparing Lunar Prospector measurements with the results of investigations of elemental composition of returned samples. Other our goal is petrologic mapping of the Moon.

Quality of Lunar Prospector elemental data: Let us estimate the quality of Lunar Prospector gamma ray spectrometer data. One of the possible ways is comparison of elemental composition of Apollo and Luna landing sites measured by Lunar Prospector and by analysis of returned samples. Bulk composition of returned samples sites is taken from [3]. Correlation coefficients between both data sets are maximal for Ti and Fe data. For other elements with weaker gamma ray lines agreement between both data sets is not so good. Correlation coefficients are positive for Mg, Ca, Al and negative for Si. Si content in Th-rich western maria is lower on 5-10 wt% according to Lunar Prospector results than that measured in returned samples. Underestimation of Si content leads to overestimation of Mg content on some weight percents in west maria. This fact can be explained by incorrect calculations of partial intensity of 2754 keV Si and Mg gamma ray lines due to interference with 2615 keV Th line. Let us note that Al has gamma ray line at 2754 keV also. However, Lunar Prospector measured Al content in Th-rich regions correctly, because Al content can be determined with use of other Al gamma ray lines.

Petrology of the Moon: Using Apollo gamma ray and X-ray spectrometers data such as Fe and Th content and Al/Si, Mg/Si ratios, it was proposed that all observed elemental abundances on the Moon can be explained by presence of three end member rock types (ferroan anorthosite, mare basalts, KREEP basalts and Mg-rich rocks) [4]. In our work we choose Mg-Al and Mg-Fe diagrams and mare basalts, ferroan anorthosites, and Mg-rich rocks as end members. Let us assume that Mg-rich rocks have the same composition as troctolites. This assumption leads to underestimation of Mg-rich rocks content, because the difference between troctolites and other end-members composition is significantly larger than that for other Mg-rich rocks as norites and gabbro-norites. The elemental composition of end-members is taken from [5]. The relative abundances of end members are plotted in ternary space for each pixel on the lunar surface. Primary colors red, blue, and green are assigned for mare basalts, ferroan anorthosites, and Mg-suite rocks, respectively (see Fig. 1). The ternary space defined by these points is represented by the mixture of these primary colors.

![Fig. 1. Scattergram shows Lunar Prospector gamma-ray spectrometer data for 5 degree squares in Mg-Fe compositional space.](image1)

![Fig. 2. Petrologic map of the Moon based on Fe and Mg Lunar Prospector data.](image2)
ected in the majority of known cryptomaria: the Lomonosov-Fleming basin, the Schiller-Schickard and Mendel-Ryder regions, and in Mare Orientale. Our results agree with petrologic mapping of the Moon conducted by [6]. This fact demonstrates the suitability of our approach for representing of lunar petrologic provinces.

![Fig. 3. Petrologic map of the Moon based on Al and Mg Lunar Prospector data.](image)

It is possible to search for rare rock types, based on Lunar Prospector data. The biggest pyroclastic deposit region at the Aristarchus plateau is distinguished from surrounding places by unusual high Mg/Al ratio, and high Th and Ti content. Other pyroclastic deposits are too small for detection at 150 km spatial resolution. While we choose troctolites as a third end-member, there are no regions on the Moon with elemental composition typical for troctolites. But other Mg-rich rock types were detected. Gabbronorites are located at the edges of eastern maria. It is difficult to distinguish norites, because norites have the same elemental composition as the mixture of ferroan anorthosites and troctolites. For better identification of lunar rock types near infrared spectra of the lunar surface must be analyzed together with gamma-ray data.

If the three end-member hypothesis is correct, the colors of all pixels on both the Mg-Fe (see Fig. 2) and Mg-Al (see Fig. 3) maps should be the same. However, there are more red and green pixels on Mg-Al map than on Mg-Fe map. This fact can be explained by existence of errors in elemental data and by the presence of rocks with different elemental composition from the end-member rocks. The degree of the difference between end-members and measured elemental composition is proportional to the distance of pixels from three end member plane. Negative ρ values mean that Fe, Mg, or Al content in Lunar Prospector pixel is higher than that in a mixture of end-member rocks, while positive ρ values mean underestimation of Fe, Mg, or Al content in comparison with end-members.

Map of ρ values on the Moon is useful for improvement of three end-member model. Addition of fourth end-member is required, because Lunar Prospector data do not lie on the three end-member plane in Mg-Al-Fe compositional space. The biggest region with negative ρ values is located in Th-rich western maria. Addition of fourth end-member with elemental composition typical for western maria is desirable for better representation of Th-rich region. Lunar Prospector measured the following elemental composition of western maria: 15 wt% Fe, 7 wt% Mg, 8 wt% Al. The best candidates for fourth end-member are KREEP basalts, which are abundant in Th-rich western maria. However, Lunar Prospector elemental composition of western maria is different from composition of KREEP basalts (9 wt% Fe, 4 wt% Mg, 8 wt% Al), norites and mare basalts. This means that the appearance of this region on ρ values map can be explained by incorrect estimation of Mg content. Thus, the quality of Lunar Prospector data is not suitable for addition of fourth end-member with use of Mg-Al-Fe petrologic technique.

Pixels with maximal positive ρ values showed as violet pixels on Fig. 3 are located in far side highlands. These pixels have lower Al content (9-11 wt%) and higher Ca content (12-16 wt%) in comparison with surrounding places. Such Ca-rich, Al-low rocks are not presented in current lunar rock collection.

When more accurate elemental data will be available it will be possible to use Mg-Al-Fe and Mg-Th-Fe petrologic techniques for determination of content of four end-member rock types (ferroan anorthosites, mare basalts, KREEP basalts, Mg-rich rocks) on the lunar surface.

**Conclusions:** Comparison between elemental composition of landing sites measured by Lunar Prospector and composition of returned samples is a good technique for estimation of the quality of Lunar Prospector gamma-ray spectrometer data. Si data are not suitable for data analysis. Mg content is overestimated by Lunar Prospector, especially in western maria. Petrologic mapping of the Moon, using Fe, Mg, and Al content, is powerful method for estimation of abundances of ferroan anorthosites, mare basalts, and Mg-rich rocks on the lunar surface. Ca-rich, Al-low small-area anomalies are detected in far side highlands. This petrologic technique can be used for analysis of future SELENE gamma ray spectrometer data.

**References:**

Venusian canali: searching for ‘starts’ and ‘stops’.
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1 - Department of Earth Sciences, Carleton University, Ottawa, ON, Canada
2 - Geological Survey of Canada, Ottawa, ON, Canada

Since the Magellan Mission in the early 90s showed the presence of canali on Venus, they have remained a geological mystery. Morphologically Venusian canali resemble terrestrial fluvial systems [4], retaining a constant width along unusually long distances [1] and tend to be concentrated on some of the plain regions of Venus [6]. To explain these unique features hypotheses for canali origin include carbonatite and sulphur-rich flows, in addition to komatiite and anhydrous lunar-type basalt [1, 3, 5].

The only known active carbonatite volcano on Earth is Oldoinyo Lengai in northern Tanzania. The carbonatite lavas are rich in Na and have the lowest measured eruptive temperatures of any terrestrial magmas (500-600°C). They also have viscosities, thermal diffusivities, specific heat capacities and latent heats of fusion considerably lower than those of basaltic lavas [2]. The viscosity and temperature of sulphur flows under Venus conditions would resemble terrestrial water and hence could be another possible means of producing canali. Other problems that involve canali are the nature of the flows, i.e. the number of pulses (one or many), the degree of erosion of the substrate, and the present location of lava that formed the canali (spread out at the end, and/or breached from the sides, and/or remained in the channel). To help answer the above questions we examined canali terminations (‘starts’ and ‘stops’), their morphology, and their relationships to nearby lava fields. Using Magellan Mission radar images, we mapped nine canali of Rusalka and Llorona Planitia including the longest, Baltis Vallis, identifying two definite and one possible ‘start’, two definite and two possible ‘stops’ and we also examined two ‘middles’.

The first definite ‘stop’ we examined is located at the northern end of Baltis Vallis. Based on a topographic study along the full length of the canali, this must be the distal (from source) end [1]. This canali can now be tracked for an additional 300 km further than previously mapped, and it stops against the topographic uplift associated with Nijole Mons (Fig.1.). A wrinkle ridge obscures the connection between the previously mapped termination and the newly discovered ‘stop’. The continuation of the canali discovered by our team has also been recognized by E.B. Grosfils during his current mapping of V14. The previously proposed end of Baltis Vallis [6], we consider to be a breached lava flow. It extends for 150 km to the south (covering an area of 1,313 km²) and is broader than the normal width of about 3 km for Baltis Vallis.

Another ‘stop’ was determined in a canali from the northwest part of Llorona Planitia, and a ‘stop’ is proposed for the Martuv Vallis; both exhibit a braided pattern which is similar to a terrestrial river delta.

Jutrzenka Vallis, located at the northeast end of Oya Dorsa, meanders through the high topography. A graben-like feature in the southwestern part of this canali could be interpreted as either the ‘start’ or the ‘stop’. Near the northern end of Ikhwezi Vallis [8] we found a previously unknown canali with
a less distinctive trough. Ikhwezi Vallis appears to terminate at the younger lava field that could represent ‘start’ or ‘stop’. However, a previously unknown canali could be a continuation of Ikhwezi Vallis.

A branch of a canali, located in the southwestern part of Rusalka Planitia, was discovered during our mapping. This branch begins from the northwest at the volcano-dominated topographic high and joins a known canali approximately 24 km from its start. In addition, we discovered a continuation of the northern end of Taipe Valles located in the western part of Rusalka Planitia [8]. A graben-like feature is proposed to be a new ‘start’ for the canali as it merges into the canali which trends to the southeast. A continuation of the canali appears to be filled by a later stage lava flow.

Two ‘middles’ of Ikhwezi Vallis were examined. The canali runs through Rusalka Planitia from northwest to southeast, exhibiting meandering, river-like morphology. The first examined ‘middle’ portion of the canali is located near its southern end, showing bifurcation and merging. The width of the canali changes from 1.5-3 km. The second examined ‘middle’ is located near the northern part of the Ikwezi Vallis. This remarkable canali meanders through high topography similar to rivers on Earth. The morphology of the canali’s trough has a constant width, 0.9 km to 1.2 km.

Our findings confirm that the canali follow topographic features, can bifurcate, can be braided, and generally have a constant width along their length. Sometimes canali are filled with lava and in some cases, such as Baltis Vallis, the canali-producing lava escaped by breaching a wall of the channel. We conclude that graben-like features are related to both ‘starts’ and ‘stops’ as previously hypothesized [7]. In addition to graben, ‘starts’ can be associated with volcano-dominated regions and lava fields, and ‘stops’ are related to topographic highs (tesserae). Some ‘stops’ exhibit a delta-like feature, explained by decrease in topographic gradient [4] or possibly by multiple pulses. It has been observed that wrinkle ridges, in all cases, are younger geological events.

**Summary:** Laboratory measurements of low-albedo powdery samples making to interpret observational data for F-type asteroids show that the optical homogeneity of microstructure of their regoliths on the scales of the order of the wavelength may be responsible for the relatively small inversion angles observed for the F-type asteroids.

**Introduction:** Polarimetry is a powerful tool to investigate of physical properties of atmosphereless bodies, in particular, asteroids. An important output of the asteroid polarimetry is in formation as for the asteroid surface texture. Dollfus et al. [1] considered the relationship between the depths of negative polarization and the inversion angles as diagnostic for the surface texture. Progress in theoretical and laboratory modeling of backscattering phenomena [see, e.g., 2,3] returns an interest to asteroid polarimetric observations. Here we present results of an interpretation of new and previous observations of several F-type asteroids, which is based on laboratory simulations of lightscattering by particulate surfaces.

**Why F-type asteroids are a type:** The F-type was first introduced by Gradie and Tedesco [4] to distinguish low-albedo asteroids with a flat (that is why F) spectrum in the wavelength range 0.3 – 1.1 μm. The typical range of albedo of F-type asteroids is 0.03-0.07. Their flat spectra were characterized with little or no absorption features. In Tholen’s classification there are 27 asteroids of the F-type that includes only about 3% of all classified objects [5]. The F-type asteroids are assumed to contain free organic compounds and resemble organic rich CI1-CM2 meteoritic assemblages [6]. We note a distinctive feature in the distribution of the F-class in the asteroid belt. Unlike other primitive classes dominated in the outer asteroid belt, there is an abundance of the F-type in the Polana family at about 2.44 AU [7]. Two near-Earth objects (3200 Phaeton and Wilson-Harrington) are also classified as F-type and both of them assumed to have the cometary origin [8].

The F-type asteroid has shown unique polarimetric characteristics. Objects of the F-type are characterized with the negative polarization branch atypical for low-albedo asteroids. The depth of its negative polarization is about 1% and the inversion angle is close to 14° (see Fig. 1), that is the smallest value ever observed for asteroids. We would remind that the typical values of these parameters for C-type asteroids, which are also very dark, is approximately 1.7 % and 18°, respectively [9].

**Laboratory simulations:** The suggested here explanation of the polarimetric peculiarity of F-type asteroids is that their surfaces have more uniform optical microstructure than other low albedo bodies. In fact, laboratory studies of mixtures of powders that are very contrast in albedo have shown a sharp amplification of the negative polarization as compared to that for the mixture components [10]. Thus, one can anticipate that adding small amount of a bright powder (e.g., MgO) to a dark powder (e.g., pure carbon soot) noticeably changes the negative polarization, producing a small albedo effect.

**Table 1. Parameters of the photopolarimeters.**

<table>
<thead>
<tr>
<th></th>
<th>Small phase angles</th>
<th>Large phase angles</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ranges of phase angles</td>
<td>0.2 – 17°</td>
<td>2 – 160°</td>
</tr>
<tr>
<td>Steps of phase angles</td>
<td>0.024°</td>
<td>1°</td>
</tr>
<tr>
<td>Diameters of samples</td>
<td>60 mm</td>
<td>10 mm</td>
</tr>
<tr>
<td>Distance from sample to source and receiver</td>
<td>1200 mm</td>
<td>350 mm</td>
</tr>
<tr>
<td>Angular diameters of source and receiver apertures</td>
<td>0.05°</td>
<td>0.8°</td>
</tr>
<tr>
<td>Whole spectral range</td>
<td>0.4 – 0.8 μm</td>
<td>0.4 – 0.8 μm</td>
</tr>
</tbody>
</table>

To illustrate this we measured low-albedo samples with laboratory photopolarimeters that are constructed and located at Astronomical Institute of Kharkov National University. One of the photopolarimeters allows photometric and polarimetric measurements at small phase angles, 0.2° - 17°. The other one permits us to study a wide range of phase angles.
angles, from 2° to 150°. In Table 1 we show several characteristics of the instruments. As an example, Fig. 2 shows an image and scheme of the instrument for measurements in the wide range of phase angles.

Fig. 2. Image and scheme of the laboratory photopolarimeter for measurements in the range of phase angles 2° - 150°.

Fig. 3 shows results of our laboratory measurements of pure carbon soot (albedo 3.6 % at phase angle 2°) and a mixture (albedo 4.3 % at phase angle 2°) of the carbon soot (95 wt. %) and very fine powder of MgO (5 wt. %). The mixture was dry and produced with shaking the powders in a test-tube. Points and crosses in Fig. 2 correspond respectively to measurements with the small-phase-angle and large-phase-angle photopolarimeters. After adding the high-albedo component, the sample albedo remains in the low-albedo domain, but for polarization the changes are significant. For instance, $\alpha_{inv}$ goes from 13.0° (for carbon soot) to 15.2° (for the mixture); this is considerable. We note also the large variation of $P_{min}$, approximately from 0.5 % to 0.8 %.

**Conclusion.** Thus, the relatively small values of the parameters $P_{min}$ and $\alpha_{inv}$ for F-type asteroids can be treated as a revealing of the optical homogeneity of regolith microstructure on the scales of the order of the wavelength. This consists with the fact that the asteroids have flat spectra, as any spectral feature (bands or UV absorption) is largely produced with relatively bright regolith components that should also strengthen the negative polarization [10].

![Polarization phase dependences of modeling samples.](image)

NORTH-SOUTH ROUGHNESS ANISOTROPY ON VENUS: SIGNATURE OF CRATER-RELATED PARABOLAS. N. V. Bondarenko1,3, M. A. Kreslavsky3-2 and J. W. Head1, 1Institute of Radiophysics and Electronics, National Academy of Science of the Ukraine, 12 Ak.Proskury, Kharkov, 61085, Ukraine, natasha@mare.geo.brown.edu; 2Kharkov Astronomical Observatory, Kharkov, Ukraine; 3Dept. Geol. Sci., Brown University, Providence RI, USA.

Introduction. Raw data of the Magellan radar altimeter (RA) experiment were processed independently by two teams. Processing carried out at MIT resulted in widely used maps of topography (GTDR data set from the Planetary Data System PDS), as well as maps of Fresnel reflectivity (GReDR) and Hagfors' roughness parameter (loosely named RMS slope) (GSDR data set). Processing carried out at Stanford University was focused on surface roughness properties extraction rather than on ranging results. Initial results from these works were reported in [1]. Later these efforts led to archiving of the results in the PDS as the so-called SCVDR data set (Surface Characteristics Vector Data Record), where data are stored as points along orbits in the sequence the observations were done, and the GVDR data set (Global Vector Data Record), where data are presented in a map-projected gridded format. We are now applying these data to geology-related studies.

Here we report on our preliminary results on the study of the Doppler centroid shift \( f_0 \), one among a number of parameters from the SCVDR and GVDR. This parameter characterizes along-track anisotropy of the backscattering function of the surface, it is most probably related to N-S asymmetry of the decameter-scale surface topography.

Doppler centroid maps: The RA echo has been sampled according to the Doppler shift into 17 bins, 935 Hz per bin (details are in [1]). This sampling is equivalent to subdivision of the RA footprint into 17 stripes normal to the orbit track. The single-burst footprint size is ~25 km near the Magellan orbit periapsis (~10°N) and up to 220 km in the polar regions.

For a globally horizontal surface with an isotropic backscattering function, the echo spectrum is symmetric with respect to the Doppler frequency corresponding to the nadir, and the maximum echo returns from the nadir stripe. For some areas on Venus, however, the observed Doppler spectrum is biased toward either positive or negative frequencies [1]. This means that the strongest echo in the along-track direction is coming from either ahead of or behind the nadir, respectively. This effect is correlated among different orbits over hundreds of kilometers on Venus, and is considered to be real [1]. Uncertainties in the Doppler shift due to inaccuracy of orbit and gravity field knowledge is at least 2 orders of magnitude smaller than the observed effect.

The SCVDR data set includes estimates of the centroid of the Doppler echo sample relative to the nadir from data averaged over each five RA bursts along orbits. The GVDR data set includes a map of the Doppler centroid; the gridding algorithm accounting for the effective footprint size, which provides theoretically optimal noise suppression with maximal preservation of information. We, however, generated another version of a gridded map of \( f_0 \) using a simpler algorithm, which ignores the footprint size: for each map cell (in the simple cylindrical 4-cells-per-degree projection) we took a weighted average of all data points whose footprint centers are within the cell; we used the inverse formal error in the \( f_0 \) determination as the weight. Typically, there are from 1 to 4 data points per map cell. Our map provides better visual sharpness, which is useful for morphological comparison with radar images. A small part of the map is shown in the inset in Fig. 1.

Physical meaning of Doppler centroid: The reason for the \( f_0 \) shift from the nadir can be large-scale surface tilt [1]. One bin (935 Hz) corresponds to the surface tilt of 0.4° over the footprint in the periapsis regions and up to 0.8° in the polar regions. In rare cases large-scale slopes of comparable steepness exist on Venus, and they are clearly seen in the \( f_0 \) maps. The region adjacent to Artemis Chasma was mentioned in [1]; we found a number of other examples.

In addition, somewhat shorter slopes of degree-scale steepness are more frequent on Venus, and we observe a characteristic signature of such slopes in the map at several sites, where they are located amid plains (e.g., wide arrow in Fig. 1).

In large tessera areas, relatively steep large-scale slopes are ubiquitous, and \( f_0 \) for each RA burst is defined by the actual balance of ten-km-scale slopes of different orientation within the footprint. This leads to an extremely noisy, spotty appearance of tessera in our map (e.g., T in Fig. 1). Sharp boundaries between units with extremely different normal reflectance can cause a shift of \( f_0 \); we encountered examples of this kind. These places in most cases should have high formal standard deviation of \( f_0 \) over the five RA bursts forming one data point, and weights in our gridding procedure diminish the role of such footprints.

In the plains, there are usually no large-scale slopes that can affect \( f_0 \), while its systematic shift is observed in many locations. In [1] saw-like small-scale topography and anisotropy of the backscattering function were considered as possible causes of this effect. Since large-scale strong anisotropy of electromagnetic properties of surface material is highly improbable, the only cause for the backscattering function anisotropy is anisotropy of subsresolution-scale surface topography and structure, and there is no reason to distinguish between a specific surface topography effect and backscattering function anisotropy. Thus, in flat areas, the Doppler centroid is a measure of the north-south asymmetry of surface topography (roughness) at scales from centimeters to hundreds of meters. Since the slopes - Doppler shift relationship and the footprint size strongly change with spacecraft altitude, \( f_0 \) is much more sensitive to the asymmetry near periapsis (low latitudes) than at high latitudes.
ROUGHNESS ANISOTROPY ON VENUS: N. V. Bondarenko et al.

Anisotropy of small-scale topography on Venus has been also observed at larger radar incidence angles with Magellan SAR data. In [2], three cases of very strong east-west asymmetry of the radar cross-section were interpreted to be due to the presence of microdunes at the surface. Systematic study [3] has revealed that much weaker east-west asymmetry is ubiquitous. It has been attributed to small-scale dunes and ripples, and some indications have been found that clear lava flow surfaces also have anisotropic topography.

**Analysis of Doppler centroid maps:** The plains areas of pronounced non-zero \( f_D \) are concentrated in an equatorial belt, which is an obvious result of the latitude trend of \( f_D \) sensitivity to roughness anisotropy. Areas of pronounced non-zero \( f_D \) on both sides of Aphrodite Terra tend to have the opposite signs of non-zero \( f_D \) (although some exceptions exist). This can be related either to a systematic pattern of microdune-forming winds related to global atmospheric circulation, or to a systematic pattern of plains-forming lava flow directions related to global topographic trends.

We searched for correlations of contrasts in non-zero \( f_D \) with geologic units boundaries, using Magellan SAR mosaics. Such correlations exist in some places, which provides an independent proof that the \( f_D \) variation is a real surface signal rather than an observational artifact. The most well-expressed correlation found so far is related to dark diffuse crater-associated parabolas. One of the best examples is shown in Fig. 1. In this example the parabola related to crater Bassi has \( f_D = 0 \), while the volcanic plains unit over which the parabola is superposed, has a pronounced N-S slope asymmetry. A similar situation is observed for craters Ban Zhao, Boleyn, Faustina, Carson, Himiko and others. In many cases the background has no pronounced asymmetry, and parabolas do not appear in the \( f_D \) map. Nevertheless, absence of roughness asymmetry is typical for all dark parabolas.

**Doppler centroid signature of dark parabolas.** Parabolas are thought to be formed by a meter-scale-thick flat-surface mantle of loose material deposited after the impact [4, 5]. Recently multipolarization Arecibo radar observations gave an independent support for this model [6]. In the frame of this model, the RA echo from the parabola is composed of two components. The first component comes from the mantle surface and has very narrow unbiased Doppler spectrum. The second component comes from the mantle/substrate interface, which we suppose to be asymmetric (because the surface adjacent to the parabola has \( f_D \neq 0 \)). The Doppler spectrum of this component is narrower than that of clean substrate surface due to the deflection of electromagnetic waves at the mantle surface, hence, its \( f_D \) is smaller. This component is much weaker than the echo from a clean surface due to absorption within the mantle. Thus, the composite echo would have small \( f_D \) of the same sign as the clear substrate. Our estimates show the \( f_D \) of the mantled surface would be indistinguishable from 0, taking into account low S/N ratio of \( f_D \) measurements.

**Discussion and future work:** The north-south anisotropy of small-scale topography adds a new dimension to understanding Venus surface properties and surficial deposits. Further work on the correlation of areas of well-expressed asymmetry with geological features seen in the SAR mosaics is underway in order to understand the role of wind-related processes in asymmetry formation. The \( f_D \) map is a potential source of information on the presence and distribution of surficial deposits. This information is critical for planning future landing missions to Venus.

A few independent lines of evidence show that dark crater-related parabolas are made of loose material and have flat surface. It remains flat for geological long enough time to form a few more large craters. Further study of parabola surface properties and the sequence of their degradation can help in reconstruction of Venus’ geological history [e.g., 7].


![Fig. 1. Magellan SAR mosaic of crater Bassi (35 km diameter, 19°S 64.7°E, thin arrow) and vicinity. Inset shows the Doppler centroid map of the same areas, gray shades denote \( f_D \approx 0 \), dark and bright shades mean deflection of the Doppler centroid from nadir. Wide arrow shows a ridge with clear expression in the Doppler centroid map. T. Ovda Tessera.](image-url)
TWIN RING YANA/OYMYAKON CORONA-LIKE FEATURE IN NORTHERN SIBERIA: YOUNG STRUCTURE WITH THE PRESENT-DAY SEISMIC ACTIVITY ALONG THE NE RIM SEGMENTS

G. A. Burba, Vernadsky Institute of Geochemistry and Analytical Chemistry, Russia’s Academy of Sciences, 19 Kosygin St., Moscow 119991, Russia; e-mail: burba@online.ru

Introduction: This abstract is aimed to describe the main points of geologic history and the geophysical environment of the mountain country adjacent to Cherskiy Range in NE Siberia, Russia. This area have been revealed to consist of the mountain ridges arranged as the two adjacent ring structures [1, 2]. Such evidence is a new look on the gross topographic structure of the area and could be of importance for its tectonic and geologic studies. On the other hand, the general appearance of these ring structures resembles the large circular features on Venus termed Corona. Such similarity could be of importance for comparative planetology studies.

Geographic setting: The detailed description of topography and location have been made earlier [1, 2], so just the main features are given here. The twin ring structure is located between Lena River mouth and Magadan coast of the Sea of Okhotsk (Fig. 1).

![Fig. 1. General view of NE Asia on satellite photo in visible spectral band with Yana (YRS) and Oymyakon (ORS) Ring Structures. Photo: NASA/Goddard.](image)

The outer diameter of each structure is ~700 km. The rim crest diameters are ~500 km for Yana Ring Structure (YRS), NW in the couple, and ~400 km for Oymyakon Ring Structure (ORS), SE one (Fig. 2). They are named after Yana river and Oymyakon settlement. YRS is located between 63 and 70°N, 125 and 140°E, and ORS – between 61 and 67°N, 136 and 151°E. The general topographic shape (Fig. 3) of the ring structures – a higher mountain ring (altitudes up to 3000 m) with a lower, but still topographically high (1000-1200 m) plateau inside, and lowland plains (50-200 m) outside – resembles typical topography of the large circular features on Venus termed Corona.

Recent geologic history: The analysis of facies and thicknesses of Neogene deposits in Cherskiy Range and the adjacent Moma and Upper-Nera depressions have shown that those segment of Cherskiy Range, which is part of the rim of ORS, survived orogenic cycle in Pliocene: intensive upward movements accompanied by gravitational nappes and overthrusts, as well as folded deformations of more ancient deposits. The cycle terminated with relief lowering and formation of a regional planation surface, which tooks place mainly due to isostatic subsidence [4].

Our analysis of topo maps revealed that in general YRS is lower in topography than ORS, especially with its inner area. It could points that YRS might undergo another evolution than the ORS, either evolution was the same, but each of the structures is currently within the different stage of its geologic life.

Geophysical environment: There are certain features, which indicate the current interior activity in the area of Yana and Oymyakon ring structures.

Seismicity. The modern seismic belt of Moma-Selennyakh rift, a part of the global rift system, follows along the NE halves of the mountain rims of the two large ring structures. The belt involves Cherskiy Range, which forms the NE halves of the two rings, being curved, convex, northeastward as two arcs – one along Selennyakh River rift valley, and another along Moma River rift valley. So the NE sides of the two ring structures, especially of the ORS, are active seismically. More than 10,000 earthquakes took place here during the recent half of century. The earthquakes occurred mainly within the rift’s mountain frame – Cherskiy Range; they are 1000 times more common here, than within the rift bottom [5].

Crust thickness. The thickness of the Earth crust is smaller within just the YRS + ORS area – there are values 30-35 km along NW-SE elongated “tongue”, located SW of Cherskiy Range – from the middle part of Yana River to the upper part of Kolyma River. The crust thickness values in the adjacent areas are considerably higher, 40-45 km [5]. The thinner crust area is considered as the area over the mantle plume [5].
**Geothermal data.** The area of YRS + ORS is also known for its thermal features. There are hot springs at the inner and outer foothills of the ring structures’ rims. The geothermal heat flux in the area is about twice higher than the mean planetary value. In the drill holes within Cherskiy Range the flux values are up to 88 mW/m² [5].

**Discussion:** All these data could be considered as the indication that Yana and Oymyakon ring structures are located over the currently active, ascending mantle plumes (hot spots). And Oymyakon plume could be supposed as more active, as it provides ORS with more considerable dynamic support in topography, then Yana plume provides to YRS. It could be supposed also that the main period of the hot spot’s activity under the YRS took place some time earlier, but now it is less active, so YRS lost its dynamic support and due to it YRS have lower position in topography if compare with ORS.

**Conclusions:** (1) The highest area of the vast mountain country in the North-East Asia consists of mountain ridges arranged as the two adjacent ring structures. (2) Cherskiy Range forms the NE halves of the two rings, being curved, convex, northeastward as two arcs: along Selennyakh and Moma river valleys. (3) Seismic belt of Moma-Selennyakh rift, part of the global rift system, follows along the NE halves of the rims of YRS and ORS. (4) The high geothermal heat flux and thinner Earth crust within YRS and ORS indicates that the both structures are located over the ascending mantle plumes. (5) Yana RS is somewhat lower than Oymyakon RS. It could points that YRS might undergo another evolution than the ORS, either the structures are currently within the different stage of the same evolutional trend. (6) Yana and Oymyakon ring structures are the surface manifestations of the currently active mantle plumes.

GEOLOGIC HISTORY OF VENUS IN SIMPLIFIED PRESENTATION: EDUCATIONAL MAP FOR THE ELECTRONIC ATLAS OF TERRESTRIAL PLANETS AND SATELLITES
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Introduction: The electronic version of the Atlas of terrestrial planets and satellites [1] is under development at the Moscow State University of Geodesy and Cartography (MIIGAiK). The geologic map of Venus is one of the new maps to update the earlier published paper version of the Atlas [1]. Such map was absent in the paper version. The main outlines of the map are reviewed in this abstract.

Map Programme: The geologic map for the electronic atlas is intended to present the structure of the surface of Venus in simplified version for the educational purposes. This map is supposed to be used for the first acquaintance with the Venus geologic history through the stratigraphic/time consequence of the geologic units on Venus as presented on the map.

Format: The map is at 1:75,000,000 scale. It is designed as the two hemispheres – “Alphan” with central longitude 0°E and “Aphroditan” with central longitude 180°E. The map projection is modified Lambert transversal equal-area azimuthal projection. The modification of the projection includes overlapping zones around each hemisphere [1]. The hemisphere have an additional 10°-wide strip along its circumference. Due to it the hemisphere diameter is not the usual 180°, but 200°. It is resulted in 20° overlapping zone along the boundary of each hemisphere.

Legend: The geologic units are the main content of the map. The number of units is minimized as the map have an elementary education purpose. The sequence of five units will be presented. They are (from older to younger):
Tesserae (T),
Ancient plains (Pa),
Intermediate-aged plains (Pi),
Young plains (Py), and
Recent rift zones (R).
These units are to be shown as colored areas. Beside them some important structures will be presented. The structures are: Ridge belts, Sinuous channels, and Craters. Such legend is based on the more detailed stratigraphic scheme of Basilevsky and Head [2], but some units of their scheme are united into one unit to form the simplified legend.

Color Design: Colors of the five stratigraphic units are correlated with those used in Venus Global Geologic Map project (VENGLOBGEC). Tessera unit is given in pink-lilac with light gray, ancient plains – in greenish-brown with light grey, intermediate-aged plains – in very light green, young plains – in bright yellow, and recent rifts – in bright red. The structures are: ridge belts – dark blue lines with bars on both sides, sinuous channels – dark green lines, and craters – black dots. Such color decision would provide visual impact of three groups of units: ancient (T and Pa) as more dark and partly achromatic, intermediate (Pi) as light and pure in color, and young (Py and R) as bright pure colors. It could create the imagination of the consequence from dark and achromatic to bright and pure, which should be associated with the consequence from ancient to young age of the units.

Feature selection: The scale of the map is of synoptic class to present the whole hemisphere on one sheet of the usual atlas format. There are 750 km of the real planetary surface in 1 cm on the map. So, only the largest craters could be shown in the scale (crater with 75 km in diameter is presented as the dot 1 mm in diameter). Due to it only the larger craters will be portrayed on the map. The diameter of 50 km has been choosen as the lower one for the map presentation (it corresponds to 2/3 mm on the map). On the same reason only the longest ridge belts and sinuous channels are to be presented. The minimal length is choosen to be 500 km, i.e. 7 mm on the map. The usual cartographic generalization take place for the areal unis. In some cases the smaller areas are united into one with the same size of the area, which is occupied by the unit on the map. In other cases very small areas are omitted and their territory is involved into the adjacent unit.

Acknowledgements: Gratitude is to K. B. Shingareva (MIIGAiK) for the general formulation and support of this work.

POSSIBLE SPECTRAL SIGNS OF SERPENTINES AND CHLORITES IN REFLECTANCE SPECTRA OF CELESTIAL SOLID BODIES

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2 Department of Spectroscopic Methods, Institute of Geochemistry, Mineralogy and Ore Formation, Academy of Sciences of Ukraine, 03142 Kiev, Palladina pr., 34, Ukraine;
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Introduction: Serpentines and chlorites are known as usual hydrothermal products of mafic minerals (olivines, pyroxenes, etc.) in terrestrial conditions [e. g., 10]. They are also major constituents of matrix minerals in the most primitive carbonaceous chondrites (CI, CM) [e. g., 5, 11, 12]. According to our laboratory spectral investigations, serpentines and chlorites have prominent visible to near-IR spectral characteristics allowing to distinguish them from those of other mineral specimens in carbonaceous chondrites. It is probably the main reason of why reflectance spectra of a large number of celestial solid bodies, from primitive C-, P-, F-, G-type asteroids [8, 9] and hydrated M-, S-, E-type asteroids [2] to Kuiper belt objects [1, 6], are similar to spectral characteristics of serpentines and chlorites. Therefore we may suppose that the phyllosilicates are widespread on silicate or silicate-icy solid bodies of the Solar System.

Laboratory Measurements and Results: Visible and near-IR reflectance spectra of serpentines and chlorites as powders of <0.20-0.30 mm grain size were measured. Previously we have obtained reflectance spectra of some carbonaceous chondrites [3] (Fig. 1). All the spectra were scanned in the 400-1000-nm range with a single-beam spectrophotometer based on a SpectraPro-275 triple-grating monochromator and controlled by an IBM 486 PC. The incident and reflected beam angles and the light beam diameter were 45°, 0° and 5 mm, respectively. Compressed powder of MgO was used as the reflectance standard. The root mean square relative error (RMSRE) for the reflectance spectra does not exceed 3% in the visible region and increases gradually to 1–2% at the red end of the operational spectral region. The reflectance spectra of serpentines and chlorites are shown in Fig. 2 and 3. A brief description of the samples is given in the Table 1.

A wide absorption band or a pair of more narrow ones (up to 30% in relative intensity) (see Fig. 2 and 3) presents in the reflectance spectra of serpentines and chlorites in the range of 500-1000 nm. There is also a specific absorption band (up to 25% in relative intensity) in the reflectance spectra of serpentines at 440 nm.

We have performed also measurements of the main oxide content in the samples by a scanning electron microscope (CamScan-4DV). Briefly, the samples do not include some anomalous quantities of oxides of transitional metals (with the exception of ferrous and ferric iron) which could influence considerably on the reflectance spectra.

Mossbauer measurements (with a molsbauer spectrometer MC1101E) were performed additionally to check contents of Fe(3+) in the serpentine samples investigated. The obtained relative quantities of Fe(3+) (in reference to values of total iron) for the serpentine samples are: 77.3% (1a), 48.4% (4b), 68.8% (10a), 88.5% (28) and 91.9% (2540).

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**Table 1. A brief description of the samples.**

<table>
<thead>
<tr>
<th>The sample names</th>
<th>Numbers</th>
<th>Physical state of the samples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orgueil (CI)</td>
<td>2476</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Mighei (CM2)</td>
<td>1856</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Murchison (CM2)</td>
<td>15044</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Old Boriskino (CM2)</td>
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<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial quartz-chlorite schist</td>
<td>3</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial clinochlore</td>
<td>22</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial chloritoid</td>
<td>32</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial Fe-clinochlore</td>
<td>35</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial serpentine (~85% lizardite-ophiite)</td>
<td>1a</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial serpentine (~70%) developed on olivine</td>
<td>4b</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial serpentine (~95% f-lizardite)</td>
<td>10a</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial serpentine (~95% f-lizardite)</td>
<td>28</td>
<td>powdered &lt;0.25 mm</td>
</tr>
<tr>
<td>Terrestrial serpentine (~85% a-lizardite) developed on pyroxene</td>
<td>2540</td>
<td>powdered &lt;0.25 mm</td>
</tr>
</tbody>
</table>
Discussion: As seen from figures 1, 2 and 3, absorption bands at 600-900 nm in reflectance spectra of the CI-CM-carbonaceous chondrites are similar to those of serpentines and chlorites. Therefore, the spectral features of carbonaceous chondrites may originate mainly in serpentines and chlorites. As authors of previous works [4, 7], we assume that the spectral features are caused by intervalence charge-transfer electronic transitions between iron ions (Fe$^{2+}$→Fe$^{3+}$).

As for nature of an absorption band at 440 nm in the reflectance spectra of investigated samples, we have no a definite opinion yet. The obtained intensity of the band (in particular for 1a, 28 and 2540 samples) is considerably more than that for similar serpentine samples in earlier investigations [e.g., 4, 7] where it was interpreted as a spectral feature produced by crystal-field spin-forbidden electronic transitions in Fe(2+) ions. As follows from our spectral and mssbauer measurements, at a qualitative level intensity of the absorption band depends on Fe(3+) contents in the serpentine samples. To solve the problem accurately, we plan to perform additional investigations of more pure serpentine samples (petrographic, mssbauer, X-ray and so on).

Thus, we suppose that the above-mentioned spectral features of serpentines and chlorites may be used as indicators of silicate matter in oxidized and/or hydrated states on different solid bodies of the Solar System.

The authors thank T. A. Smirnova (VIMS, Moscow) for help in mineralogical-petrographic investigations of the serpentine samples.

The ESA Mars Express mission was successfully launched on 02 June 2003 from Baikonur, Kazakhstan, onboard a Russian Soyuz rocket with a Fregat upper stage. The mission comprises an orbiter spacecraft, which has been placed in a polar martian orbit, and the small Beagle-2 lander, due to land in Isidis Planitia but whose fate remains unknown. 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.

Following the Mars Express spacecraft commissioning in January 2004, most experiments onboard began their own calibration and testing phase already acquiring scientific data. This phase will last until June when all the instruments will start their routine operations. The MARSIS radar antennas, however, will only be deployed in early May in order to maximise daylight operations of the other instruments before the pericentre natural drift to the Southern latitudes, which coincides with nighttime conditions required for subsurface sounding. Initial science results are summarised below.

The High-Resolution Stereo Colour Imager (HRSC) has shown breathtaking views of the planet, in particular of karstic regions near the Valles Marineris canyon (pointing to liquid water as the erosional agent responsible for modifying tectonic and impact features in the area) and of several large volcanoes (Olympus Mons caldera and glaciation features surrounding Hecates Tholus). The IR Mineralogical Mapping Spectrometer (OMEGA) has provided unprecedented maps of water ice and CO\textsubscript{2} ice occurrence in the South pole, showing where the two ices mix and where they do not. The Planetary Fourier Spectrometer (PFS) has measured atmospheric CO variations in each hemisphere and 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 CO\textsubscript{2} density and temperature, and has simultaneously measured the distribution of water vapour and ozone. The Energetic Neutral Atoms Analyser (ASPERA) has identified the solar wind interaction with the upper atmosphere and has measured the properties of the planetary wind in the Mars tail. Finally, the Radio Science Experiment (MaRS) has studied for the first time the surface roughness by pointing the spacecraft high-gain antenna to the Martian surface, which reflects the signal before sending it to Earth. Also, the martian interior has been probed by studying the gravity anomalies affecting the orbit due to mass variations of the crust.

Water is the unifying theme of the mission to be studied by all instruments using different techniques. Geological evidence, such as dry riverbeds, sediments and eroded features, indicates that water has played a major role in the early history of the planet. It is assumed that liquid water was present on the surface of Mars up to about 3.8 billion years (from crater counting relative ages), when the planet had a thicker atmosphere and a warmer climate. Afterwards, the atmosphere became much thinner and the climate much colder, the planet loosing much of its water in the process as liquid water cannot be sustained on the surface under present conditions. Mars Express aims to know why this drastic change occurred and where the water went. A precise inventory of existing water on the planet (in ice or liquid form mostly below ground) is important given its implications on the potential evolution of life on Mars, as the 3.8 b.y. age is precisely when life appeared on our own planet, which harbored similar conditions to Mars at that time. Thus, it is not unreasonable to imagine that life may also have emerged on Mars and possibly survive the intense UV solar radiation by remaining underground. The discovery of methane in the atmosphere could indicate just that or the presence of active volcanism. From previous orbital imagery, volcanoes on Mars were assumed to have been dormant for hundreds of millions of years. This idea needs a fresh look as the implications of currently active

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volcanism are profound in terms of thermal vents providing niches for potential ecosystems, as well as for the thermal history of the planet with the largest volcanoes in the Solar System. 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 lifetime of the orbiter spacecraft is of one Martian year (687 days), potentially to be extended by another Martian year to complete global coverage and observe all seasons twice. *Mars Express* is the first European mission to another planet. For details: [http://sci.esa.int/marsexpress/](http://sci.esa.int/marsexpress/)
GENERALIZED TOPOGRAPHY OF THE LUNAR SOUTH POLE – AITKEN BASIN. V.I.Chikmachev, S.G.Pugacheva and V.V.Shevchenko, Sternberg State Astronomical Institute, Moscow University, Moscow, chik@sai.msu.ru

**Introduction.** The nature and origin of the enormous pre-Nectarian South Pole – Aitken (SPA) basin remain the most important problems in the current lunar studies. The basin represents a geophysically, compositionally, and topographically unique lunar formation. The studying SPA is significant for principal reasons concerning earliest epoch of the bombardment rate of the Moon and possible investigations of the very deep lunar interior. Recently the SPA basin has been proposed as a high-priority target for future robotic sample return mission [1].

We revised topographical data obtained for the region to construct a generalized structure of the SPA basin as ringed formation. We have used a cartographic method to analyze properties of the system of concentric depressions inside the ring structure.

**Hypsographical map of the basin area.** We have compiled a variety altitude data sets to construct general hypsographic map of the lunar far side included the ring basin structure: Clementine laser altimetry results [2]; Zond catalog of the absolute heights [3]; catalog of the lunar limb heights [4, 5]. Perspective azimuthal orthographic projection was used as cartographic basis of the general hypsographic map. Because of our previous analysis of the ring structure [5, 6] we assumed the projection center coordinates as 180°W and 40°S. The hypsographic map shown in Fig. 1 demonstrates a hemisphere of the Moon which contains the SPA basin structure and its environs.

**Size of the SPA basin.** General size of the SPA basin is defined by outermost ring of the structure that has been extensively modified by post-SPA impact events. According to [7] multiple craters are superposed on the SPA basin. This process brings forward extensive destruction of the initial form of the outmost basin ring. So, we can observe a relic relief in this area now. It’s reason of the large range of estimations of initial SPA basin size.

We tried to trace relic features of the possible initial outermost basin rim. Fig. 2 shows 3-D model of relief corresponding outmost basin ring. This model shares out relief, which has height more than 0 km. As usually, this level is in agreement with lunar sphere of radius equal 1738.0 km. However, the model shown in Fig. 2 is “flat”, i.e. it is constructed on the plain without considering spherical effect.

We interpret that largest segment of the ancient outermost rim is placed in azimuth range from 335° to 150°. (The azimuth A is measured clockwise from central meridian 180°). This segment looks like an extent ridge of mountain in form of huge arch with height from 2 km to 8 km (from zero level). The most width of this formation exceeded 600 km in area of the post-SPA impact Korolev basin (A = 35°). The east part of the ridge is placed near environs of Mare Orientale basin (A = 110°). The southern part of this arch segment is modified by the younger depression placed between craters Mendel and Rydberg (A = 135°). This unnamed depression was shown in the sketch map published by Hiesinger and Head [7] too. Further we can see a few separate tops with heights of ~ 2 km (A = 150° – 165°), which are interpreted by us as relic details of the ancient outermost rim. South part of the supposed rim (A = 180° – 200°) encloses ridge of mountain with elevation ~ 4 km placed between craters Demonax and Boguslavsky.

In range of azimuth from 200° to 265° we observe an extend drop in elevation connected with Mare Australe. The new segment of the probable outermost ring extends in north - west part of the rim (A = 265° – 330°). This ridge of mountain with elevation ~ 2 ÷ 4 km includes crater Tsiolkovskij.

On the basis of our ring reconstruction we concluded that original size of the SPA basin (outermost ring diameter) is approximately 3300 km. Then we find the basin outermost ring center at roughly 180° and 40°S.

**Inner structure of the SPA basin.** As following from Fig. 1 and Fig. 2 the border of the depression with values of heights H < 0 km has form of the near right ellipse. Dimensions of the oval formation are approximately 2200 km x 1800 km. Center of the ellipse displaces from center of the outermost rim to the south on ~ 300 km. The next depression has oval form too. The elevation level of the its border is H < - 4 km. The transversal is about 1400 km, and ratio of the axes is equal to 1.2. Center of the depression displaces from center of the outermost ring to the south-east on ~ 500 km. The deepest inner depression has nearly circular form with diameter of 600 km. Its height level is H < - 6 km. Its center displaces from the basin center to the south-east on more than 700 km.

**Conclusions.** On the basis of our study of the generalized structure of the SPA basin we conclude that giant impact formed this basin unit was oblique or trajectory of the impactor was tangent to the surface of the lunar sphere. Because of very small value of ratio “deep – diameter” (~ 0.004) and small possibility of the long-term viscous relaxation [7] we propose that impactor had a small density of its matter, i.e. it was a giant comet body.

We propose a radio-science experiment in the frame of the European Space Agency Bepi Colombo mission to study the deep interior and subsurface of Mercury. The proposed experiment HeRS (HeRmean Radio Science experiment) will only use the nominal TT&C equipment onboard the Planetary Orbiter (MPO) and the ground segment without modifications and will be able to address fundamental questions regarding the physical characteristics of the innermost planet of our solar system, Mercury, by measuring its gravity field and the physical librations. The general objectives of the HeRS experiment are:

1. To study the orbital motion of the spacecraft to obtain the global gravity field of Mercury and the tidally driven gravity variations, to an accuracy required to constrain the structure of the mantle and crust/mantle interface, as well as the deep internal structure of the planet;
2. To measure the degree-2 gravity coefficients and the rotation state of Mercury, so as to constrain the size and physical state of the core of the planet.

We plan to establish a global gravity map with a resolution of about 300 km (harmonic degree 25) by means of a dynamical modeling of the MPO orbit. This gravity map will be supplemented with local maps in areas under the pericenter at a resolution of 200 km with line-of-sight inversions of the Doppler data. The geometry of the orbit, such as the eccentricity and the inclination, is fundamental in the determination of the gravity field precision that can be achieved by using HeRS data. As determined from simulations, the MPO orbit is appropriate to reach the ESA required precision on the gravity field by using the nominal radio-science configuration.

We propose the to use the gravity anomalies to investigate the crust and lithosphere of the planet. Gravity anomalies, in conjunction with the altimeter data, will allow us to estimate the thickness of the crust as well as its spatial variability. The lithospheric thickness can also be deduced from a combined analysis of gravity anomalies and topography, and will be used to determine the heat flow at the time of loading.

The low degree gravity coefficients $J_2$ and $C_{22}$ will be used, in conjunction with libration and obliquity observations to determine the principal moments of inertia of the core, the mantle, and the whole planet. The amplitude of libration and the obliquity angle will be determined jointly with other BepiColombo instruments (the camera and the star-tracker). These measurements will benefit from Earth-based radar data. We expect a precision of 0.003 on the moment of inertia factor $C/Mr^2$ and a precision of 0.05 on the ratio between the moment of inertia of the part of the planet participating in the libration to the moment of inertia of the whole planet ($C_m/C$). In order to interpret libration in terms of the interior of the planet, a precise theory of the spin and orbital motions will be used. This theory, in addition to the classical 88 days libration, will allow accounting for the proper frequencies generated by the 3:2 spin-orbit resonance of Mercury.

The tidal variations of the gravity field will provide the value of the Love number $k_2$, which is very sensitive to the state and dimension of the liquid outer core and the solid inner core. The Love number, together with the moment of inertia values, can be interpreted in terms of physics of the interior of the planet. This study will therefore provide the best values for core parameters such as radius, density and composition. We will for instance be able to constrain the amount of sulfur in the iron alloy of the core and set limits on the temperature. This is also important for modeling a possible hydromagnetic dynamo.

Introduction and Background: Since the detection in MOC images of geologically recent gullies on Mars [1], several hypotheses have been proposed for their origin, broadly partitioned into models of groundwater seepage [1-3] and melting/runoff of isolated snowpacks [4]. While these models have attempted to account for the concentration of gullies in the mid-latitudes, few studies have examined their immediate geological context. Christensen [4] noted that MOC data have revealed gullies on isolated surfaces that would be unlikely locations for groundwater seepage. In this work, we present the preliminary results of our survey of geological environments of gullies revealed in MOC images. We show examples of gullies found on central peaks and on the outside of crater rims, where there is insufficient volume for a contained aquifer, a critical component of all seepage models. These examples also illustrate families of gullies with alcoves that emanate from different elevations, not distinct layers. This leads us to favor a surface melting/runoff origin for these examples.

Central Peak Gullies: Our survey has revealed several examples of gullies that are found on the walls of central peaks within craters in the southern mid-latitudes, in addition to gullies found on isolated mesas. An example is provided in Figure 1a (subframe of MOC E15/00539), which shows the alcoves and channels of several gullies incised into the central peak of Lohse Crater, at 17.21°W, 43.72°S. Alcoves range in greatest width from ~100 m to ~700 m, and channels extend for several hundred meters to the east, off of the MOC frame. While these gullies exhibit a west-east trend, gullies found in the same image to the north show both north-south and south-north trends. The alcoves are found at variable elevations along the slope of the central peak, and no evidence for layering has been observed. Both the alcoves and channels exhibit lower albedos than the terrain into which they are incised. The gullies appear fresh and no superposed impact craters are observed.

Gullies on the outside of crater rims: We have observed multiple examples in MOC data of gullies that have formed on the outside of craters with raised rims. An example is shown in Figure 2b (subframe of MOC E11/03663), which shows the southern rim of a ~5km diameter crater on the floor of the larger Newton Crater, at 157.62°W, 39.77°S. These are found within a region of high gully concentration (Figure 2a). These gullies show a poleward orientation, consistent with the well-formed gullies on the inside of the crater's northern rim. The dimensions of these gullies are smaller than those to the north, with channels that extend only a few hundred meters, and depositional fans that are ~100m long and ~100m wide at their terminus. Alcoves are not resolved at MOC resolution, given the small size of these gullies. It is unclear whether they emanate from a continuous layer on the outside of the crater rim or from a consistent elevation. Consistent with previous studies of gully landforms, these gullies appear to be the youngest features within their proximity, showing depositional aprons that superpose the terrain immediately surrounding the crater.

Well-formed gullies on the outside of crater rims have also been observed along the northeast rim of Hale Crater, a location well studied due to its high-concentration of gullies. Figure 3a (subframe of MOC image R07/02777) shows a traverse of the northeast rim of Hale Crater, at 35.60°W, 35.43°S. Gullies are observed on both the inside and outside of the rim. Alcoves for all gullies are close (within tens of meters) to the crest of the rim, but the rim has remained intact. Well-defined fans are observed for the gullies within Hale Crater, but channels extend off of the MOC frame for the gullies on the outside of the rim. All gullies observed appear youthful and are stratigraphically the youngest features in the region of study.

General Observations and Discussion: We have begun a survey of the geologic setting of MOC images that exhibit gully landforms in the middle/high latitudes of Mars in an effort to decipher patterns with regard to their morphology, distribution, and their immediate geological context. Our initial findings of gullies on central peaks, mesas, and the outside of crater rims confirm the assertion of Christensen [4] that specific families of gullies on Mars occur in locations unlikely to produce groundwater seepage.

Models that invoke groundwater seepage for the formation of martian gullies [1-3] all depend upon either sufficient subsurface volume behind the gully for a confined aquifer that undergoes fluctuation in pressure [1,2] or sufficient geothermal activity that would transport volatiles from greater depth to the surface [3], which would then be released through seeps. Our observations of the geological context of gullies suggest that these conditions are not met across the surface of Mars. Well-defined gully alcoves in regions of small subsurface volume have been found (Figures 1 and 2), and no relationship between gully distribution and centers of geothermal activity has been observed.

In their updated survey of gullies on Mars, Edgett et al. [5] restated their observation that gully alcoves emanate from specific layers exposed on given slopes. While this is common, we have documented examples of gully alcoves that do not exhibit this relationship (e.g. Figure 1a and Figure 2a), and we suggest that this will be an effective preliminary test to differentiate gullies formed by groundwater seepage and those formed by runoff of melted snow. Edgett et al. [5] also noted that there does not appear to be a global preference for poleward-facing slopes, as was initially thought [1]. They did observe that gullies in specific areas face the same direction, and our findings are generally consistent with this observation. No craters have been observed in our survey that exhibit gullies facing each other. We have, however, found occurrences of gullies on the inside and outside of the same crater rim, as in Figure 3.

All proposed models acknowledge that gully formation is likely tied to orbital cycles in the Late Amazonian. Recent work has attempted to model the orbital cycles of Mars over the last 10 Myr. [6], and to tie those cycles to the geological record observed on the surface [7,8]. Our future work will further this effort and attempt to determine the orbital, climatic and geologic conditions necessary for the formation of the gullies we have documented.

Figure 1. (a) Subframe of MOC E15/00539, showing gullies incised into the central peak of Lohse crater. (b) MOC E15/00540, context for 1a.

Figure 2. (a) MOC E11/03663. Box is context for 2b. (b) Subframe of MOC E11/03663, showing gullies on the outside of a crater rim. (c) Sketch map for 2b.

Figure 3. (a) Subframe of MOC R07/02277, showing gullies on opposite sides of the rim of Hale Crater. (b) Sketch map for 3a.
NEW TASKS ON MARS INVESTIGATION IN INTEREST OF COSMOGONY, PLANETOLOGY, AND VERIFICATION OF JORDANO BRUNO’S HYPOTHESIS ABOUT GREAT NUMBER OF INHABITED WORLDLS. E.V.Dmitriev, Euro-Asian Astronomical Society (EAAS), Moscow, e-mail: deva1001@mtu-net.ru

Basis. The hypothesis about Megatungus explosion in the primary earthlike atmosphere (PMA) of the Mars due to which the planet dichotomy was created [1,2] requires to be verified. The hypothesis supposes that the temperature in the Mars nourishment zone was low thanks to this the planet embryo had time to absorb significant portion of pre-planet cloud gas and further increment of the planet mass had taken place already in presence of PMA. This circumstance became the determining factor during the planet forming and its early history. Due to PMA the Mars obtained torque, its rock body wasn’t exposed to high warming-up and was protected against right catastrophic impacts of large objects within the period of giant meteorite bombardment of the Solar system planets. This scenario of the planet forming and its relative small mass – ten times as lighter than Earth – didn’t favor to evolution of endogene processes characteristic of Venus and Earth. Mars kept its original view that is why its superficial layers would be satiated with fragments of comets and asteroids, and other space objects.

The incomplete scenario of the planet forming [3] showed that the determinant stipulating the Solar system forming became the moment when intensive star wind of the proto-Sun appeared. Namely this circumstance divided the Solar system into two groups differs sharply from one another as to their properties – the planets of the Earth group and the gaseous planets-giants. There are many considerations allowing to suppose that this scenario of the planet forming is most likely ordinary phenomenon. All the above raises definite hope for enough high percent of probability that the Earth-type planets could be appeared near several star types on which the life could be conceived and transformed in highly developed civilizations in time. Most probably that the evolution of such civilizations was accompanied with exhaust of large amount of their activity products into space. The Solar system crossed many times the Galaxy branches characterized with heightened star population during its motion about the Galaxy center. A very long period of time (~4.5 billion years) and good keeping of substances disposed onto the Mars surface thanks to its neutral atmosphere give reason to raise a question about purposeful search of artifacts or signs of non-Earth’s civilization representatives staying on the Earth. It is necessary to note that the Mars surface areas almost equaling to the Earth’s land, which is being permanently renewed. To verify the suggested ideas it is proposed to perform a series of purposeful right investigations of Mars with the help of automated probes.

Tasks. The main tasks of the suggested investigations are to collect material evidences of: the planet PMA availability in past (task No.1); the Farcide hill forming due to joining of large planetzimal to Mars (task No.2); influence of shock waves created during the Megatungus explosion upon the planet solid surface (task No.3); availability of non-Earth civilization activity products or signs of staying on the planet. (task No.4).

Investigations. According to the suggested hypothesis [3], the Mars lithosphere created as a result of the subsequent deposition of high-melting component of PMA was not melted and was not significantly changed later on. During the final phase of the Mars accretion, PMA dust deposition should be accompanied with gravitation-aerodynamic separation of substance that should cause poorness of atmosphere upper strata as to heavy elements and enrichment with light elements. That is why Task No.1 can be solved by investigating the lithosphere upper layers up to depth of several kilometers beyond the boundaries of the north planet depression. Scientists are very lucky in this case. It was noted many times that Mars is unique planet as to science. Many billions years ago the Mariner canyon had “ripped up” lithosphere up to depth of 7 km thus to solve Task No.1 it is required only to investigate the observed layers of the canyon walls up to the maximum possible depth. For this purpose it is necessary to deliver to the planet surface a winch (main module) and robot-cliffer. The winch should be secured at brink of the most deep place of the canyon to be able to lower or lift the robot-cliffer which would be investigate subsequently the canyon wall layers. The main module would provide power supply and control of the robot. Meanwhile it should be able to move in different directions, be protected against overturning or be able to recover its operational position after overturning. Investigation of soil layers of the canyon walls of the Mariner valley would allow the estimate of the saturation degree of the Mars surface
with meteorite substance to be made. It may appear that the significant part of the rocks observed on the planet surface are impact ones as well as fragments of comets and asteroids fallen out onto its surface during the billion years.

The second task is solved significantly easier: it will be sufficient to find the signs of general impact-metamorphosis transformation of rock in the most lowered parts of the north depression allocated between 70° and 50° NL, i.e. within the zones of maximum shock waves affecting upon the planet surface created due to the Megatungus explosion. Besides these rocks shall be depleted with volatile components in comparison with the lithosphere upper layers of the southern hemisphere. The most suitable objects to be investigated might be breccia released from craters. To perform these investigations the Mars Exploration Rovers of “Spirit” and “Opportunity” types could be used well.

The same way could be used for solving the third task. It is necessary to determine what rocks (volcanic or sedimentary rocks) make the Farsid hill spur jutting into the northern depression.

To solve the fourth task - searching of extraterrestrial civilization signs – it is necessary to perform the detailed photographing of the planet with resolution of one decimeter or better prior to start up the intensive exploration of Mars. This photographing could be realized with the help of low-orbit satellites or onboard equipment of aircraft. When “suspicious” objects are found the Mars rovers equipped with manipulators should be sent to them. This will allow to perform three-dimensional photographing and to extend significantly the possibility of their investigation.

**Scientific significance of the proposed investigations.** In contrast to many Mars programs having cognitive character and are designed partially to be luck the suggested investigation plan has sharply defined goals to be sought and directed to solve fundamental problems of planet creation, early history of Mars and other planets as well as to find another civilization signs.

**References:**
THE NEW DATA ON THE EARLY STAGE OF DEVELOPMENT OF THE EARTH, MARS, THE MOON AND MERCURY. A.V.Dolitsky¹, R.M.Kochehtov², E.A. Kozlova³, J.F.Rodionova¹, 1 - United Institute of Physics of the Earth RAS, Moscow, av1386@comtv.ru, 2 - Moscow Technical University of communication and information, Moscow, krmkm@rol.ru, 3 – Sternberg State Astronomical Institute, Moscow, jeanna@sai.msu.ru

The graphic method of the analysis of an arrangement of linear structural elements on geographical and geological maps of continents (A.V.Dolitsky, [1]) has allowed to find out global stress fields of the past and among them one which directions (the main normal and maximal tangents) describe more than 80 % of all linear elements of a relief. It has given the basis to the author to accept it for an initial global stress field of the Earth. Association in system of poles of a global stress field has allowed to find its trajectory and to establish laws of rotation of the mantle over the core [2]. In 2003, A.V.Dolitsky and R.M.Kochehtov have developed a computer method of the similar analysis of an arrangement of faults. Z.F.Rodionova and F.F.Ajnetdinova have created a databank on faults of Mars, and it has been successfully used for a finding of a trajectory of movement of its geographical pole (rotation of the mantle over the core). Results of the analysis have been reported on the previous Microsymposium [3]. Later the computer program has been advanced. It allows to connect now databanks on faults of the planets, created on the certain rules, and to find coordinates of poles of a global stress field, as the points of crossing of arches of the big circle describing these faults. The found poles share program way on density of an arrangement on 3-4 groups. Databanks on faults (set by the coordinates of their final points) of the Earth and the Moon (A.V.Dolitsky, N.N.Semenova), Mars (Z.F.Rodionova) and Mercury (E.A.Kozlova) are created. As against rotational stress fields, symmetric to geographical poles, the found out global stress fields (and the faults) are symmetric to points of the most powerful explosions happened at early stages of development of the Earth and planets of terrestrial group. The arrangement of the areas of development of such explosions allows to do new conclusions on features of their development.

The computer analysis of an arrangement of faults on a surface of the Earth has opened completely unexpected laws. Poles of global stress fields are found within the limits of the areas of the different form. The greatest number of poles and their greatest concentration are found out within the limits of circles in diameter in 3000 km in the Central Europe (the center - pole L1, Fig.1) and in antipodal areas of the Southern hemisphere (the center - pole L2). These poles were earlier established by the author graphic methods [1]. Appeared, that there is more than 60 % from the common number of such poles. Probably it is polar areas of the Earth where from the lower mantle, at non-uniform rotation of the Earth and change of its axial compression, the fused magma acted, causing the crushing of the mantle accompanied with formation of global stress fields. Consequence of such magmatic outpourings in polar areas could become formation in them of superfluous weights and their displacement (in structure of the mantle) to equator under action of centrifugal forces of rotation of the Earth. Zones of high concentration of the poles, containing about 20 % from their common number, are found out along four axial zones of meridian directions the prodeletings, uniting circular (polar) areas (Fig.2). Probably, these zones of meridian directions have arisen as a result of fast reduction of volume of the Earth at formation of the core (Fig.3). That fact, that more than 80 % of poles have strictly ordered arrangement on a surface of the Earth, testifies to an invariance of an arrangement of faults answering to them, hence, about an invariance of an arrangement of the blocks divided by them. It means, that the displacement of blocks during a geological history and shifts, at the summation do not collect, and are mutually compensated, specifying elasticity of lithosphere. The marked magmatic processes proceeded, apparently, in the lower mantle under action of heat just the arisen the core when the upper mantle remained still cold. Result of these processes became extensive outpourings of granitoids, containing radioactive elements. By the data of the computer analysis, contours of storage sites of these breeds are close to contours of modern continents Eurasia (Fig.4) and America N. and S. (Fig.5). It means an invariance of their arrangement and their origination from granitoids of a lower mantle at the early stages of development of the Earth. The nature of a continental crust is those. Process of warming up of the mantle, connected with disintegration of radioactive elements later began to develop and prove. It has caused carrying out from it on a surface of gases and the liquids which have formed air and water environments. The upper mantle, having lost a significant part of the liquid and gaseous components (existing in it originally as ice) has been sated with the new mineralization which has come from depths and metals. The nature of an oceanic crust is those. The computer analysis of an arrangement of faults of Mars (Fig.6), the Moon (Fig.7) and Mercury has found out existence at Mars four depressions of meridian directions, at the Moon - three and at Mercury - two. On an arrangement of these depressions in an equatorial belts of planets and on meridian directions of their axes as at the Earth, their formation can be counted simultaneous to occurrence of cores at these planets. Conditions of formation of a continental and oceanic crust of these planets are same with the Earth conditions, probably.
References:
Radiating graben-fissure systems on Venus: A tool for characterizing magmatic events and their age relationships

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Magmatism on Venus: There are several scales of magmatic events on Venus (Ernst and Desnoyers 2004): 1) isolated coronae, volcanoes, flow fields, and radiating graben systems; these range in scale up to 1000 km or more in diameter; 2a) individual and small clusters of volcanoes and coronae associated with topographic swells, geoid highs, and triple junction rifting; these are most clearly indicative of terrestrial-type plumes originating from the deep mantle; 2b) coronae distributed along rifts (chasmata); these are the clearest examples of melt generation associated with rifting; 3) regional concentration of activity in the Beta-Atlas-Themis (BAT) region; this is the closest example of a terrestrial plume cluster event, sometimes termed a ‘superplume event’; and 4) global volcanic resurfacing of the volcanic plains; no terrestrial analogue is confirmed, although the global burst of terrestrial plume activity in the Neoarchean is a possible analogue. New insights into this record are available from detailed mapping of graben-fissure systems.

Importance of Graben-Fissure Systems:
Reconnaissance-scale (225 m/pixel C1-MIDR) Magellan images revealed 163 radiating graben-fissure systems of which 118 were interpreted to be underlain by dykes (Grosfils and Head 1994) as well as linear and circumferential systems. More detailed mapping using full resolution (75 m / pixel) Magellan images is revealing a significantly larger population. A 14 Mkm$^2$ area (264$^\circ$-312$^\circ$ E, 24$^\circ$-60$^\circ$ N) near Guinevere Planitia (Ernst et al. 2003) contains thirty-four radiating systems, of which 16 have radii greater than 300 km and eight have radii greater than 1000 km. (In this area only 5 radiating systems were identified in the earlier reconnaissance mapping). Furthermore, twenty-six linear (straight) systems with a length greater than 300 km have been distinguished of which six have a length greater than 1000 km, and 19 circumferential systems are mapped and identify coronae. In addition, preliminary detailed mapping in the adjacent area (216$^\circ$-264$^\circ$ E, 24$^\circ$-60$^\circ$ N) has revealed 30 radiating systems of which 9 have radii greater than 300 km and two have radii greater than 1000 km. (In this area only 4 radiating systems were identified in the earlier reconnaissance mapping).

The radiating graben-fissure systems that we catalogue represent a database of tectonomagmatic centres that complements the centres defined using other criteria, e.g. large volcanoes, coronae, and shield fields. Using radiating graben-fissure systems in this way has some distinct advantages (e.g. Ernst et al. 2003), summarized below:

1) The areal extent of most systems greatly exceeds that of the volcano or corona on which they are centred. Thus there is the potential for accessing the relative age of widely separated volcanoes or coronae by the crosscutting relationship between their associated radiating graben-fissure systems.

2) Some radiating systems lack any volcano or corona at their centre, indicating radiating graben systems can identify cryptic mantle plumes / diapirs.
3) Faint closely-spaced lineaments, often in crosscutting sets (“gridded terrain”), can often be traced into unambiguous graben-fissure systems, suggesting that the faint lineament sets can be used to extend the distribution of graben-fissure systems.

4) Graben-fissure systems are especially prominent in fracture belts and densely fractured plains, but some can be traced as more subdued features into adjacent plains. This indicates that fracture belts and densely fractured plains represent ‘basement’ units, which are overlain by plains units. The local thickness of the plains units will determine whether the graben-fissure systems in the ‘basement’ are partially or completely obscured. Therefore, regional variation in the visibility of partially flooded graben systems can potentially be used to determine the variation in thickness of plains lavas.

5) Some radiating graben systems extend a few hundred km before swinging into a common direction that presumably reflects the influence of a regional stress field. Others continue radiating for their full extents. The pattern and age distribution of fully radiating vs. swinging swarms should reflect the regional stress pattern and its variation through time (Grosfils and Head 1994). These results can be compared with those determined from wrinkle ridges which appear to correlated with long-wavelength topography.

6) Since Venus is a one-plate planet with a stagnant lithosphere, individual large volcanoes represent a combination of plume head and subsequent plume tail activity. It is possible that a giant radiating swarm is associated with only the plume head stage and its distribution could help define plume-head vs. plume tail components of the volcanism. Additional graben-fissure system geometries are important. Circumferential systems are used to map coronae, and it should be possible to establish age relationships with respect to associated radiating systems. Linear systems may be linked with rift zones but may also be the distal portions of larger radiating systems.

**Lessons for Earth:** On Venus, graben-fissure systems preserve their primary geometry and distribution because of the absence of plate tectonics. This characteristic provides insights into the primary distribution of intraplate magmatism on Earth (including that associated with mantle plumes) where regional relationships are obscured by the uncertainty of plate reconstructions, especially in the Precambrian.

**References**


LACUSTRINE DELTAS IN A CRATER LAKE IN THE NILI FOSSAE REGION: NEW EVIDENCE FOR PERSISTENT FLOW OF WATER ON THE SURFACE OF MARS.  C. I. Fassett and J. W. Head III, Dept. of Geological Sci., Brown Univ., Providence, RI 02912. (e-mail: Caleb_Fassett@brown.edu, James_Head@brown.edu)

Introduction: New observations reveal the presence of a pair of distributary fans where two valleys enter a 40-km unnamed crater, centered at 77° 40' E and 18° 25' N. Possible modes of origin for such fans are volcanic activity [1], as deltas in a crater lake [2] or as an alluvial fan. We believe that the evidence suggests that the deposits described in this work are fluvial deltas, deposited in a standing body of water that may have persisted for a long period of time.

Recently, two papers have described the strong evidence for a delta or fan in a crater northeast of Holden crater [3, 4]. The features discussed herein differ from the “Holden NE” crater delta in two fundamental ways: the Holden NE fan deposit is much larger than the small deltas observed here, and the Holden NE fan has been inverted by subsequent exhumation. The new features examined are potentially significant for understanding the martian water cycle. This is especially true given recent results that question the fluvial-lacustrine interpretation of (at least one) proposed crater lake/delta deposit on Mars [1].

Observations: We compiled MOC, THEMIS, and MOLA data for the region in the ArcMap GIS environment. Figure 1 shows a THEMIS IR mosaic of the region colored with MOLA topography. Valleys enter the crater from the west and north, and led to the formation of two apparent fan complexes. A primary reason we interpret these features as having formed subaquously is the topography of the crater. On the eastern margin of the crater, there is an outlet channel which has eroded to a level of approximately -2395 m elevation. The delta deposits have median elevations ~50 meters below this level.

The outlet channel appears to significantly incise the eastern rim. Thus, the hypothesized lake appears likely to have had a much higher stand in the crater than the current lowest elevation of the notch. Two lines of evidence suggest that this higher stand was at an elevation of -2260 (+60, -20) meters. First, there appears to be a small portion of the northern delta with a maximum elevation just below the -2260 m contour, well above the present elevation of the exit breach. After lake level fell the northern valley cut, and was diverted by topographic barrier. Second, a break of slope in MOC images E1301894 and E1101284 is observed at approximately the -2260 m contour.

Figure 2 is a sketch map of the deltas and the valley networks in the observed region. The deltas are ~50 km² in area. An estimate of the volume of sediment in each deposit is 1-3 km³, although the original topography of the underlying surface is uncertain. The smooth deposit mapped in Figure 2 may also be finer-grained sediment which was deposited off the deltas themselves.

Details of the MOC narrow angle image E1301894 shows a section of the northern valley and its fan, revealing a mottled texture probably related to weathering. Some suggestion of layering in the eroding material is evident. The fan related to the western valley is comparatively well preserved, with several possible leveed distributary channels. It has a “birdfoot” morphology which bears a resemblance to elongate, fluvial-dominated deltas on Earth [5]. Unfortunately, no high resolution images of this western delta are available.

Input Valleys: The two valleys that led to the formation of the crater lake both stretch for substantial lengths, with the western valley extending at least 250 km to the west and the northern valley reaching ~70 km to the north and west. The westernmost valley drops ~2.4 km from its headwaters to the crater, which gives it an average gradient of 1%, which is substantial for a martian or terrestrial valley. The valleys have well-developed meanders, possible cutoff segments, and potential oxbows most easily observed with the THEMIS IR night data, presumably because of thermal inertia contrast of eolian deposits on the valley floors with the surrounding material. As in other locations on the martian surface, few major tributaries are observed here despite the long length of these valleys. The meandering nature and long length of these valleys suggest (but does not require) an extended period of activity.

There is a substantial lack of knowledge of the depth and width of flow in martian valleys given the fact that we typically have no evidence of interior channels [cf. 6]. This certainly holds for the valleys which were the inputs to the hypothesized crater lake. Nonetheless, the volume we infer for the crater lake using present topography (without correctly for sedimentation or other infilling) is ~500 km³, and even for very large flows it is hard to fill such a volume without persistent flow of water on the surface. We have used the Manning equation scaled for martian gravity [7] to infer very rough estimates of water flux, which suggest minimum flow times on the order of years. This is consistent with the minimum filling calculations made for the “NE Holden” system [4].

We estimate that filling the crater (and delta formation) took much longer, probably at least thousands of years. There are two major reasons for this belief: (1) These small input valley networks (forming meandering and cutoff channels) are very different from outflow channels (which presumably formed catastrophically); instead, they appear much more like fluvial valleys on Earth. (2) Extrapolation of our volume estimates for the deltas suggest that a conservative upper limit on total sedimentation volume is on the order of 10 km³. If we divide this into the volume we believe the crater lake must once have contained (~500 km³) (a minimum volume of water that flowed through the system) gives us an upper limit on the sediment to water ratio of ~0.02. This is an order of magnitude smaller than some sediment water ratios cited for larger channels on Mars [8], and more consistent (though still larger) than usually observed in fluvial systems on Earth [9].

Conclusion: These observations argue for persistent flow on the surface of Mars; catastrophic formation of the
delta observed here seems highly unlikely. It also provides new evidence for the existence of at least some fluvial-lacustrine environments on Mars, in contrast to the volcanogenic origins interpreted for some delta features [1]. Future work will further test interpretations suggested here by: (1) examining individual MOLA tracks to obtain better absolute constrains on elevations, (2) investigating the implications of the thermal properties of various units, and (3) examining the input valley properties and regional context in greater detail.


Figure 1. THEMIS daytime IR mosaic with MOLA topography (ranges from white -1200 m [and above] to purple -3000 m [and below]).

Figure 2. Sketch map of valley networks in the region of the hypothesized crater lake, along with delineation of the extent of the deltas and the smooth deposit (see text).
PRELIMINARY ANALYSIS OF THEMIS INFRARED DATA ASSOCIATED WITH THE MARS NORTH POLAR BASAL UNIT. K. E. Fishbaugh¹, ¹International Space Science Institute, Hallerstrasse 6, CH3012 Bern, Switzerland, kathryn.fishbaugh@issi.unibe.ch

Introduction: The north polar basal unit (BU), lying stratigraphically between the polar layered deposits (PLD) and the Vastitas Borealis Formation (VBF) has been hypothesized to be an Amazonian-aged frozen deposit of both eolian [1] and small, ancient polar cap deposit origin [2]. This unit has a relatively low albedo and patchy layering (Fig. 1) and appears to be the main if not sole source for the north polar dunes which comprise the largest sand sea on Mars [1,2,3]. Study of the origins of this unit has led to important insight into north polar geologic history as a whole. Periodic migration of sand northwards from lower latitudes to create the layers of the BU may have been controlled by the formation and erosion of the ice-rich, low latitude mantling layer described by [4,5]. Additionally, a possible higher obliquity (~45°) during much of Mars’ past [6] may have contributed to the lack or smaller size of a polar cap during the time of BU formation. In this study, I begin an analysis of Mars Odyssey THEMIS data which lend further insight into the characteristics of this unit and its relationship with surrounding deposits.

THEMIS Data: I have begun preliminary analysis of THEMIS infrared data of the BU by comparing the relative differences in temperature between major polar features. Two examples are illustrated in Figs. 2 and 3. Each figure shows the THEMIS IR image, a temperature map created using the Brightness Temperature Record associated with that image, and the context MOLA shaded relief map taken from the THEMIS website (http://themis-data.asu.edu/). Both images were acquired during northern summer. The temperature maps were created by converting the DN values of the IR image pixels to brightness temperature using the formula

\[ \text{Temperature} = \text{scaling} \times \text{factor} \times \text{DN} \times \text{value} + \text{offset} \]

The scaling factor and offset (or minimum temperature) can be obtained from the .IMG header file for each image. The temperatures were then put in 5-degree bins in order to highlight major variations, so that a temperature of 217.3 K, for example, appears as 220 K on the map. All temperatures are in degrees Kelvin, and both figures use the same temperature scale.

Fig. 2 encompasses PLD, a likely exposure of the BU, the dunes of Olympia Planitia, and a few knobs (e.g., A) associated with VBF plains. These knobs may be part of what has been interpreted as kame and kettle terrain deposited during retreat of the polar cap [7,8]. It is immediately obvious that the dune-covered areas in Fig. 2 are relatively much hotter than the layered terrain and the knobs. This could be due to two main factors: the small grain size of the dunes may give them a relatively low thermal inertia [9], the dunes have a relatively low albedo, and Olympia Planitia slopes southward. Thus these dunes have been heating up all day; note that the local time for both images (~18:00) is within daylight hours during northern summer. Similarly, the scarp that contains a likely exposure of the BU is also quite warm as is the area at the scarp base, as warm in some spots as the dunes. Again, part of the warmth may be due to the fact that this is a south-facing scarp. However, the temperatures may also be warm because the BU consists of material similar to that comprising the dunes. This may also argue for a relatively low ice content at least in the upper few microns to centimeters of the unit (the depth to which THEMIS observes in the IR).

If we focus in more closely on the layered terrain, we see that south-facing trough walls appear warmer than north-facing as evidenced by the orange and dark maroon banding (e.g., at B). Flat areas on the surface of the ice (e.g., at C) appear the coolest. Interestingly, the bottom left corner of the image exhibits a feature in the plains which is as cold as portions of the ice surface. Either there is a shallow (at most µm’s-cm’s depth) exposure of ground ice or a very thin surface exposure of ice here (e.g. frost). This investigation must be continued by comparing THEMIS IR data with visible images to search for surface frost. Alternatively, this cold spot could be an exposure of bedrock with a relatively high thermal inertia, so high that the rock was not significantly heated during the day. It is important to note that [7] and [10] have hypothesized that cap retreat may have left remnants of ice (partially buried) in this area.

Fig. 3 shows PLD and associated troughs, a section of the head of Chasma Boreale, and a portion of the VBF plains just beyond the cap. This area of the cap appears to be much warmer than the area near Olympia Planitia. The reason for this is not immediately apparent. The banding (e.g., at A) which may have been due to heating of southward-facing walls in the area shown in Fig. 2 does not directly correspond to trough geometry in this region (e.g., compare to shaded relief). Instead, it appears that the warm areas of the troughs lie on the flat areas just adjacent to the south-facing walls, and the main flat areas between troughs are not as easily distinguished in temperature from the troughs the same as Fig. 2.

The chasma floor (B) and scarp containing a BU exposure (C) appear to be the hottest regions of the image. Since the floor is as warm as the scarp, it is unlikely that high sun exposure on the scarp is the exclusive reason for the warmth. Instead, dunes on the chasma floor as well as the presence of the BU within that scarp (documented by [2]) are more likely heating factors. Warm spots on the cap itself (D) could be due to dust or ground ice, a possibility which will be investigated by comparison with visible images.

Strangely, the plains in this region appear only slightly warmer than the cap, the reason for which has yet to be understood. Note that the tiny exposure of the crater (E) is warmer than the surrounding plains. Since the crater ejecta and outside of the rim are sloping away from the sun, the high temperature is probably due to exposure of relatively high thermal inertia bedrock in the crater rim and ejecta which has retained its warmth.

Conclusions and Future Work: Preliminary analysis of north polar THEMIS IR data has begun to yield interesting results. Dunes and exposures of the BU appear to be relatively warm and of similar temperatures, consistent with earlier conclusions that they may be composed of the same material [1,2]. While in some locations, north-facing trough walls are cooler than south-facing, that relationship is not always immediately apparent. However, the troughs in both examples are cooler than the BU. The polar cap in Fig. 2 is not much cooler than the surrounding plains in that area so that plains are not distinguishable from ice. This must be investigated further. Another interesting feature of the polar plains is the cool spot pictured in Fig. 3 which could possibly be a sign of ground ice.

Further analysis of these and other THEMIS data should in-
volve noting characteristics of and relationships between major polar features. These data can also be put into ESRI Arcview to create temperature map and THEMIS image mosaics which can be analyzed simultaneously with geo-referenced data from other instruments. The temperatures from the images can then be compared to temperatures of specific materials predicted by thermal models, thus allowing an estimate of the thermal inertia of various features (e.g. to search for ice vs. bedrock). This analysis will provide further insight into the origin of the BU and the geologic history of the north polar region, extending the work done by [2].


**Fig. 1.** An example of the BU (portion of MOC image E0201209) wherein the BU is the dark, layered material lying beneath the brighter polar layered deposits. The dark material at its base has been reworked into dunes in other locations. Arrow indicates sun direction. Below is a MOLA shaded relief context map showing location of entire image. This image lies in a trough near Olympia Planitia(83.85°N, 237.80°W).

**Fig. 2.** THEMIS infrared image I04626009, the associated temperature map, and MOLA shaded relief context image. Local time is 18.690. Arrow next to scale bar indicates sun direction. Temperature scale is in Kelvin and applies to both figures. See text for description of annotations.

**Fig. 3.** THEMIS infrared image I04396009, the associated temperature map, and MOLA shaded relief context image. Local time is 18.533. Arrow next to scale bar indicates sun direction. Temperature scale is in Kelvin and applies to both figures. See text for description of annotations.
HABITABLE ZONES IN EXTRASOLAR PLANETARY SYSTEMS: THE SEARCH FOR A SECOND EARTH. S. Franck, W. von Bloh and C. Bounama, Potsdam Institute for Climate Impact Research, PF 601203, 14412 Potsdam, Germany (franck@pik-potsdam.de, bloh@pik-potsdam.de, bounama@pik-potsdam.de).

Introduction: The search for extrasolar planets is one of the main goals of present research. Up to now, more than about 130 extrasolar giant planets are known to orbit around Sun-like stars including several multiple-planet systems. These giant planets, with hydrogen and helium as the main constituents, have atmospheres too turbulent to permit the emergence of life and have no underlying solid surfaces or oceans that could support a biosphere. The distribution of masses of all known exoplanets lets scientists suppose that there must be a multitude of planets with lower masses. Even if it is today beyond the technical feasibility to detect Earth-mass planets we can apply computer models to investigate known exoplanetary systems to determine whether they could host Earth-like planets with surface conditions sufficient for the emergence and maintenance of life on a stable orbit. Such a configuration is described as dynamically habitable.

In our approach habitability does not just depend on the parameters of the central star, but also on the planet itself. In particular, habitability is linked to the photosynthetic activity of the planet, which in turn depends on the mean global surface temperature and the planetary atmospheric CO2 concentration. Our integrated systems approach is applied both to the solar system and to the extrasolar planetary systems 47 Ursae Majoris (UMa) and 55 Cancri (Cnc).

Model Description: On Earth, the carbonate-silicate cycle is the crucial element for long-term homeostasis under increasing solar luminosity. Furthermore, on geological time-scales the deeper parts of the Earth are considerable sinks and sources for carbon. In addition, the tectonic activity and the continental area change noticeably. Therefore, we favour the so-called geodynamical models, which take into account both the growth of continental area and the decline in the spreading rate. Our numerical model couples the stellar luminosity, the silicate-rock weathering rate, and the global energy balance to allow estimates of the partial pressure of atmospheric and soil 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 CO2 sink in the atmosphere-ocean system and the metamorphic (plate-tectonic) sources [2].

Results for the Solar system: The results for the estimation of the HZ for the solar system are summarized in Figure 2, where we have plotted the width and position of the HZ for three different points in time (past, present, future). In about 500 Myr the inner boundary reaches the Earth’s position and the biosphere ceases to exist. The outer boundary decreases in a strongly non-linear way.

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Figure 1: Box model of the integrated systems approach [1]. The arrows indicate the different forcings (dotted lines) and feedback mechanisms (solid lines).

Figure 2: Habitable zone (green shading) for the solar system at three different time steps. The orbits of the three terrestrial planets, Venus, Earth, Mars, are shown. The solid green lines describe the evolution of the inner and outer boundary of the HZ [3].
Results for extrasolar planetary systems: The first system, 47 UMa, has been identified to host two Jupiter-mass planets at respectable distances from the host star, which has properties very similar to those of our Sun, including mass, effective temperature, spectral type, and metallicity. The star of the second system, 55 Cnc, has an outer planetary companion orbiting at Jupiter distance and two inner giant planets at very small orbits (hot Jupiters). We have investigated whether these extrasolar planetary systems could host Earth-like planets on stable orbits within the habitable zone. In Figure 3 we show the results of our calculations of the HZ for a likely value of the central star luminosity (color shaded) and the gray-shaded range of orbital stability. The intersection of the two areas describes the interesting parameter range where an Earth-like planet on a stable orbit can exist within the HZ. It is evident that an almost completely ocean-covered planet (“water world”) has the highest likelihood of being both habitable and orbital stable. If the planet is covered with more than 50% continental area, then habitability and orbital stability cannot be found for the entire assumed range of stellar age. For a continental area of more than 90% of the total surface, no habitable solutions also meeting the requirement of orbital stability exist.

Figure 3: The habitable zone around 47 UMa for the likely value of luminosity. The colored areas indicate the extend of the HZ for different relative continental areas. The gray shaded area indicates the permissible parameter space as constrained by the possible stellar age and the orbital limit at 1.25 AU [4].

Also in the case of 55 Cnc an almost completely ocean covered planet has the highest likelihood of being habitable [5]. In Figure 4 we show the HZ for the systems 47 UMa and 55 Cnc in comparison with the Solar system for an Earth model with constant continental area of about 30%.

Conclusions: In the present investigation we have shown that the existence of a dynamically habitable Earth-like planet is principally possible both in the system 47 UMa and in the system 55 Cnc. This likelihood depends critically on the percentage of the planetary land/ocean coverage and is significantly increased for planets with a high percentage of ocean surface (“water worlds”).

A consortium of Institutions from Russia, Italy and USA proposes to start using aboard landers the Attenuated Total Reflection (ATR) spectroscopy technique, so far not used in planetary missions. The ATR-spectrum is similar to an absorption spectrum of a sample being in contact with an ATR-prism. Simplicity of sample preparation and absence of atmosphere contribution are among the main advantages of this method. The proposed SIATR experiment (Soil Investigation by means of ATR) has two main tasks: mineralogical analysis of Martian soil and search for primitive forms of Martian life. The analyzed spectral range is 2.5–25 µm with resolution of 5 cm$^{-1}$. ATR spectra obtained with SIATR would be complementary to ATES’ data, providing also a “ground-truth” reference for it.

The SIATR instrumentation includes a miniature Fourier-spectrometer with a source of IR radiation and a number of small changeable compartments, each with its own ATR-prism, allowing analysis of multiple samples. We consider SIATR as a “Group 1” (analytical laboratory) investigation. Mineral powder samples will be delivered to SIATR by MSL SA-SPAH system.

First, “as is” ATR analysis of the sample is performed, providing basic mineralogical information. Then, a modification of spectrum with heating of the sample allows to decide on water and other volatiles abundance. Similarly, adding some water may discover clays, etc.

To detect living or dormant bacteria, potentially survived on Mars, the SIATR instrument is equipped with two additional systems. The first allows culturing of the bacteria, which propagate on the ATR-prism resulting in progressive deepening of absorption bands, characteristic to alive cells (e.g., “Amide-2” absorption of proteins at 1550 cm$^{-1}$). The culturing is achieved by providing optimal temperature, optimal chemical environment, etc. If no propagation is detected, the second technique, namely layer-by-layer ATR-analysis, allows detection of spores and mummified cells.

The ATR spectra allow to detect biomolecules themselves (proteins, DNA/RNA, lipids, carbohydrates…), not just products of cell’s metabolism.

Extensive laboratory experiments have demonstrated the effectiveness of the ATR-spectroscopy for the specified goals, both mineralogical analysis, and the detection of live cells.

**APPLICATION OF ATR-SPECTROSCOPY ABOARD LANDERS.**


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Possible sequence of SIATR measurements is as follows:

- A new ATR-cell (a ~7x9x20mm small box with the ATR-prism as the bottom side) is set in working position
- The reference spectrum is acquired
- A sample of the soil is delivered into the ATR-cell
- “As is” ATR-spectrum of the soil sample is acquired (basic mineralogy analysis)
- Sample is heated, water ice (if any) melts out, volatiles evaporate
- Altered mineral ATR-spectrum is acquired
- Liquid water is added to the sample, altered mineral ATR-spectrum is acquired
- An appropriate chemical environment for culturing of the bacteria is provided in the ATR-cell
- The ATR-cell is warmed up to +30°C for some time (maybe several days)
- The lid is opened, water evaporates, the matter sediments onto ATR-prism
- The biological ATR-spectrum is acquired (the bio-sequence may be repeated)
- A new, clean ATR-cell is set, the sequence is repeated with new soil sample

SIATR concept. SIATR is a “Group 1” (analytical laboratory) investigation. Mineral powder samples will be delivered to SIATR by MSL SA-SPAH system.
EXCITATION OF FREE OSCILLATIONS ON MARS. T.V. Gudkova and V. N. Zharkov. Institute of Physics of the Earth, Russian Academy of Sciences; B. Gruzinskaya, 10, Moscow 123995, Russia. gudkova@ifz.ru

Introduction. Since the outer layers of Mars are very heterogeneous, as in the case of the Moon [1], it is difficult to construct a spherically symmetric model of its interior structure (a model of a zeroth approximation), by using only a few seismometers for recording seismic body waves. First of all, they will allow to obtain the structure of the crust at the sites of their location. That is why, in this paper the excitation of free oscillations of the planet will be considered.

The estimate of oscillations amplitude. The level of tectonic and geological activity on Mars suggests that it should be seismically more active than the moon but less active than the Earth. Phillips and Grimm [2] and Solomon et al. [3] considered that only the thermoelastic cooling of the lithosphere could generate marsquakes. They found that more than 10 events of seismic moment greater than $10^{21}$ dyne cm, and more than 250 events of magnitude greater than $10^{20}$ dyne cm may be expected per year. A few (2-3) per year should have a moment greater than $10^{24}$ dyne cm. A $10^{20}$ dyne cm quake is the upper bound of the estimate of the activity on Mars [2]. Their estimates of seismicity are consistent with conclusions by Golombek et al. [4]. Golombek et al. have determined the seismicity on Mars based on all shear faulting visible at its surface. They have concluded that Mars is seismically active today.

To judge whether the free oscillation method can be used to study the Martian interiors, it is necessary to estimate the amplitudes for different types of free oscillations during marsquakes and to determine how these amplitudes depend on focal depth and excitation processes based on the available estimates of the Martian seismic activity and the sensitivity of current instruments. Currently available broadband seismometers can measure accelerations [5]

$$a_{N,E} = -\omega_0^2 u_{N,E} \approx 10^{-8} \text{ms}^2.$$  \hspace{1cm} (1)

Thus, the problem is to find the modes that satisfy condition of Eq.(1) and to assess their diagnostic capabilities.

We have calculated the amplitudes of torsional and spheroidal oscillations for sources at different depths (0-300 km) and with different focal mechanisms for a trial model of Mars. The displacement components $u_N, u_E$ are proportional to the seismic moment $M_o$ of the source. That is why, to estimate the values of the displacements for different seismic moments, they are calculated for a unit seismic moment $M_o$.

We have considered two possible locations of a marsquake: in Olympus region ($135^\circ W, 18^\circ N$) and in Valles Marineris ($80^\circ W, 5^\circ S$) and located a seismometer at "Gusev" crater ($14.64^\circ W, 18^\circ S$) and $175.06^\circ E$.

Figure shows the amplitudes of horizontal displacements $u_N$ and $u_E$ for the fundamental tones of torsional oscillations for two different focal mechanisms. We see from Figure that the displacements of the torsional fundamental modes with $\ell \leq 20$ lie in the range of $10^{-12}$ to $10^{-11}$ cm for a unit seismic moment. For a marsquake with $M_o = 10^{22}, 10^{24}, 10^{25}$ dyne cm the amplitudes of oscillations lying above the corresponding curves are about $10^{-9}$, i.e. they satisfy Eq.(1). And, consequently, the torsional modes with $\ell \geq 3$ (if a marsquake with $M_o = 10^{25}$ dyne cm occurs), with $\ell \geq 6$ ($M_o = 10^{24}$ dyne cm), and with $\ell \geq 12$ ($M_o = 10^{23}$ dyne cm) could be detected by currently available instruments. The torsional modes with $\ell \geq 3$, 6 and 12 can sound the Martian interiors down to 1600, 1100 and 700 km, respectively [6].

The displacement amplitude for the overtones is smaller than that for the fundamental modes. Only a marsquake with $M_o = 10^{25}$ dyne cm can excite overtones of torsional oscillations with $n=10$-$20$ and $\ell = 2$, the amplitudes of which satisfy the condition of Eq.(1).

The noise level on Mars can reach significant values [7], and in this case torsional modes will not be observed for seismic moments described above. Torsional modes will be observed if a real seismic event has a larger moment or a seismometer is placed by a penetrator deeper into the ground in order to eliminate wind effects.

A marsquake with a seismic moment $10^{25}$ dyne cm is required for the spheroidal oscillations to be detected. In this case, the spheroidal modes with only $\ell \geq 17$ could be detected by currently available instruments. The spheroidal modes with $\ell \geq 17$ can sound the outer layers of Mars down to 700-800 km [8]. For a marsquake with a higher seismic moment ($10^{26}$ dyne cm) the spheroidal modes with $\ell \geq 26$ could be detected. The spheroidal modes with $\ell \geq 26$ can sound the outer layers of Mars down to 2000 km [8]. Only a marsquake with $M_o = 10^{26}$ dyne cm can excite overtones of spheroidal oscillations with $n=5-25$ and $\ell = 2$.

These results are in agreement with the results obtained in [5], where the authors concluded that normal mode detection would be clearly successful for $10^{25}$ dyne cm seismic moment and $10^8$ ms$^{-1/2}$ noise level and the moment may be reduced to $10^{24}$ dyne cm for a noise level of $10^{10}$ ms$^{-1/2}$.

Conclusion. In the future the Netlander mission will have a geodesy and seismic payload and it is of great importance for studying Martian interior. The paper is related to the excitation of normal modes and the possibilities of detecting such modes by future Mars missions.

We would like once more to emphasize the importance of the information on normal mode frequencies, in order to determine the very deep structure of Mars. A good installation of a broadband seismometer is mandatory to provide this information. It is found down to what depth the normal modes can sound the planetary interiors. A marsquake with a seismic moment of $10^{25}$ dyne cm is required for spheroidal oscillations (with $\ell \geq 17$) to be detected. These spheroidal modes are capable sounding the outer layers of Mars down to a depth of 700-800 km. These results are in agreement with the results obtained by Lognonné et al. (1996), who concluded that normal mode detection would be clearly successful for a $10^{25}$ dyne cm seismic moment marsquake and $10^9$ ms$^{-1/2}$ noise level.

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Amplitude of displacements $u_N (a, b)$ and $u_E (c, d)$ for the modes of torsional oscillations with $\ell = 2-20$ and $n=0$ versus frequency $f = 1/T$ and degree of oscillation $\ell$ for two focal mechanisms. $M_0$ is equal to 1. The focal mechanisms are 45°, 45°, 45° (a, c) and 90°, 90°, 90° (b, d) for the dip, strike and slip angles. The seismometer coordinates are 15°S, 185°W. The epicenter coordinates are 18°N, 135°W (Olympus), the epicentral distance is 59.3° (open circles - a focal depth of 0.3 km and open squares - a focal depth of 300 km) and 5°S, 80°W (Valles Marineris), the epicentral distance is 103.1° (filled circles - a focal depth of 0.3 km and filled squares - a focal depth of 300 km).
HIGH PURITY GE GAMMA-RAY SPECTROMETER ON JAPANESE LUNAR POLAR ORBITER SELENE.

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The SELENE project, the first mission of Japan to the moon, is in progress. The mission is to be launched in 2006. SELENE is a lunar polar orbiter at its altitude 100km, and in operation for one year. If possible it will be extended by more one year at a lower altitude.

This note serves to describe a gamma-ray spectrometer (GRS) onboard SELENE. Its structure is shown in Fig.1 and its appearance including a radiator is shown in Fig.2. This is aimed to remote sensing elemental materials over the entire lunar surface and to make global mapping. In particular, we are interested in the major elements, such as Mg, Al, Si, Fe and so on, and natural radioactive elements like K, Th and U etc. By the measurements, we study the origin, evolution and structure of the moon. Also it is interested in a signature of hydrogen, because it would prove evidence of existing water on the surface.

GRS consists of three parts; gamma-ray detector (GRD), compressor driver unit (CDU) and gamma-ray and particle electronics (GPE). Since GRD plays a major role of GRS, this note is exclusively concerned with GRD.

GRD is expected to identify elements uniquely and to be operated more than one year. For this purpose, a highly purified Ge detector is employed as a main detector, by which gamma-rays can be detected in the energy range from 0.1 to 12 MeV with a considerably high detection sensitivity. Its active volume is 252cc (manufactured by Eurisys Measures). The Ge crystal is encapsulated with an aluminum container, canister...

The Ge detector should be kept below 90K, in order to avoid possible damages from radiation. It is retained at ~80K, by means of a mechanical cryostat. The main part of the cryostat is a Stirling cryocooler, which is composed of compressor and cold head. The cryocooler is driven with 17V at 52Hz. Then its cooling capacity is 2W at 80K with an input power of 53W.

The cryocooler has two characteristic features: One is a dual-opposed-pistons compressor, with which effects of mechanical vibration is considerably suppressed. Accordingly, the energy resolution of GRD is expected to be better than 3keV at 1MeV. The other is a long-term operation. Up to now, it has been achieved a normal operation beyond 33,000hrs in a laboratory life-time test.

The Ge detector is surrounded by a BGO crystal and a plastic scintillator. The former is of a horse-shoe shaped. The asymmetric shape makes it efficient to veto gamma-rays from the SELENE vehicle and leakage photons from the Ge detector, as well as to avoid charged particles. The latter faces the lunar surface opposed to the vehicle and to veto charged particles.

Performance of the Ge detector is carried out by using the flight model.

GRD is well examined against vibrations at launch, which was simulated by applying an AT level.

As for the cryocooling performance, since the canister has a heat load of 1.8W in total, the cooling capacity is sufficient to keep the whole system at ~80K, when the ambient is at room temperature.
Indeed, it takes less than 24 hrs to arrive at ~80K from room temperature.

The energy resolution was examined by cooling the detector at ~80K. As one of results, two gamma peaks from a $^{60}$Co source are shown in Fig.3. The energy resolution was achieved to be 3 keV fwhm at 1.33 MeV.

From this result, it is anticipated to identify $2.223\text{MeV}$ photons from the capture reaction, $n+p \rightarrow d+\gamma$, by discriminating $2.210\text{MeV}$ photons from Al and $2.235\text{MeV}$ photons from Si, which are abundantly populated on the lunar surface.

Fig. 3. Observed photon peaks of 1.17 and 1.33 MeV photons from a $^{60}$Co source.
ANTARCTIC DRY VALLEYS: MORPHOClimATE ZONATION, VARIABLE GEOMORPHIC PROCESSES AND IMPLICATIONS FOR ASSESSING CLIMATE CHANGE ON MARS. J. W. Head¹ and D. R. Marchant², ¹Dept Geol. Sci., Brown University, Providence RI 02912 USA; ²Dept. Earth Sci., Boston University, Boston MA 02215 USA.

Introduction: The Antarctic Dry Valleys are generally perceived of as a hyperarid cold polar-desert environment with characteristics that can aid in the interpretation of geomorphic features and climate conditions on Mars. Here we further explore this analog by outlining three major microclimate environments within the Dry Valleys that are defined on the basis of differences in annual surface temperatures (and the resultant behavior of meltwater), relative humidity, and soil moisture (and the resultant types of subsurface ice): 1) a Coastal Thaw Zone, with a traditional active layer and ice wedge polygons similar to those of the Arctic; 2) an Inland Mixed Zone showing variable soil moisture and both composite and sand wedge polygons; 3) an Upland Frozen Zone, with low soil moisture (~<5% (ultraxerous) and without segregation ice and a traditional active layer. The subtle but important variations in these parameters among the three micro-climate environments result in considerable variation in the distribution and characteristics of geomorphic features and processes at all scales. These guidelines and analogs can be readily applied to the interpretation of similar features on Mars to help deduce individual current microenvironments and to document their change during climate fluctuations.

Morphoclimate 1: Coastal Thaw Zone (Subxerous); Macroscale topography. Low-rolling hills (~<500 m relief) and wide valleys (~8 km) dominate the macroscale topography of the coastal thaw zone. Bedrock slopes average from 5° to 20°. North-facing slopes are less steep by several degrees than south-facing counterparts. Glaciers and perennial snow banks lose mass dominantly by sublimation. A significant fraction (15-20%) of ice melting occurs at the steep cliffs that front most alpine glaciers in the coastal zone. Gullies in this zone are generally deeper and possess more rounded interfluves than counterparts in the inland mixed zone. The well-developed gullies of the coastal thaw zone likely form from a combination of runoff from snowmelt, freeze-thaw, and salt weathering. The magnitude, rate, and location of snowmelt in gullies vary considerably within the coastal thaw zone. Provided sufficient snow/ice exists in gullies, the major factors that control gully hydrology include slope angle, aspect, shielding (terrain obstructions), cloud cover, and surface albedo. Topography, at all scales, is the dominant control on solar radiation. At the largest scale, solar radiation, as averaged over monthly timescales, is greatest for north-facing slopes. This finding is consistent with geomorphic observations that show that 1) north-facing gullies are better developed than south-facing gullies and 2) that north-facing slopes are generally less steep than south-facing slopes. Slope gradients decrease with increasing solifluction, salt-weathering, and freeze-thaw processes.

Meso-scale topography. Unconsolidated sediments superposed on the bedrock topography of the coastal thaw zone are seasonally wet and display evidence for lateral flow (solifluction) and melting of subsurface ice (thermokarst). This zone displays a traditional active layer down to ~30-60 cm depth; moisture in this layer varies from ~6%-77%. The relatively mild summer conditions and traditional active-layer dynamics permit the development of rills, channels, debris flows, ephemeral ponds, and intermittent streams. The beds of intermittent streams, particularly low-gradient beds (~<20°), commonly display a boulder pavement that grades laterally into a hyporheic zone, which is saturated and permits downstream throughflow. If sufficient meltwater exists, sediments within the hyporheic zone may move downslope via solifluction. The largest and best developed streams flow from the snout of glaciers in the coastal thaw zone. Soils developed in the coastal thaw zone are classified as Anhyturbels and contain salts enriched in sodium chloride. Contraction-crack polygons are common in this zone. The seasonal influx of liquid water into open thermal-contraction cracks, and subsequent expansion on freezing, leads to the development of ice-wedge polygons with raised rims. The width of ice wedges (as measured across their top surface, ~<2.0 m) is generally smaller than that found in the Arctic, reflecting the limited availability of liquid water in the Dry Valleys. Rock glaciers are rare in the coastal thaw zone. This could reflect 1) the paucity of exposed bedrock cliffs that could produce sufficient rockfall to accumulate a surface lag on glacier ice or 2) the lack of extensive talus and colluvium on valley walls that could develop flow through the formation and deformation of segregation ice.

Morphoclimate 2: Inland Mixed Zone (Xerous); Macroscale topography. Heavily incised slopes dominate the macroscale topography of the inland mixed zone. On average, these slopes are steeper (by as much as 5-10°) and rougher (as measured over 10-m-scale baseline) than those of the coastal thaw zone. Gully density is about twice that of the coastal thaw zone (even in areas of similar granitic-bedrock lithology) and interfluves show greater angularity than those near the coast. There is a transition from smooth, low density gullies in the coastal thaw zone to jagged, high density gullies in the inland mixed zone. Although minor variations in bedrock structure could play a role in gully density and morphology, we suggest that excessive snowfall near the coast, accompanied by relatively high meltwater runoff, freeze-thaw, and salt weathering may account for the relatively large size of gullies and apparent rounding of interfluves in the coastal thaw zone. The low density would thus reflect the progressive development and evolution of master gullies, which form at the expense of minor gullies and troughs that are incapable of trapping requisite snowfall. If correct, then the relatively high density of gullies in the inland zone may reflect a paucity of snow accumulation (rather than insufficient elapsed time). Insufficient meltwater runoff, freeze thaw, and salt weathering in the inland mixed zone could prevent generation of mature gully networks. We note that in contrast to the coastal thaw zone, small talus cones, rather than streams, occupy the base of many gullies.

Meso-scale topography. Slow moving gelifluction lobes with concave longitudinal profiles and ribbed, steep fronts (~<30° in some cases) represent the dominant form of mass wasting in the inland mixed zone. Gelifluction lobes may show plug-like flow with considerable velocity variation. Rock glaciers occur sporadically within the inland mixed zone. These features are morphologically similar to gelifluction lobes described above, but contain a significant percentage of buried ice that moves through shearing and plastic deformation. The rock glaciers commonly form downslope from local accumulations of snow and ice, are generally tongue-shaped in plan view, and display alternating surface ridges-and-furrows and steep termini when active. Of 32 rock glaciers surveyed in a region roughly coincident with the inland mixed zone, 38% are stagnant. Although the origin of ice in rock glaciers is commonly debated, most rock glaciers in the inland mixed zone are likely cored with secondary ice, reflecting the development and deformation of subsurface ice within pre-
existing colluvium. Although some have noted that ~15% of the surveyed rock glaciers are “in transition with alpine glaciers”, the link to these extant glaciers is likely through melting, percolation, and subsequent refreezing of meltwater in pre-existing colluvium, rather than through progressive burial of glacier margins. Apart from gelifluction lobes and rock glaciers, the dominant mesoscale feature of the inland mixed zone are sand-and composite-wedge polygons. Sand-wedge polygons form in a manner analogous to ice-wedge polygons except that contraction cracks are filled with sand, rather than with ice. Composite polygons contain wedges with alternating lenses of ice and sand. The growth and evolution of sand- and composite-wedge polygons have been studied extensively. The retention of open contraction cracks at the ground surface and the availability of eolian sand to fill such cracks are determining factors in polygon growth. Because cracks tend to be wider and remain open longer in soils with cohesive near-surface horizons (ice and/or salt cemented), sand-wedge polygons are most active near the margins of perennial snow banks and/or in regions that experience minor snowmelt. The sands and gravels that line the troughs of sand-wedge polygons are generally coarser-grained and show less compaction than sediment at polygon centers. Where sand-wedge polygons form over buried ice, this textural variation is particularly important in modulating the rate of underlying ice sublimation, with the highest rates occurring beneath coarse-grained sediment at polygon troughs. Ultimately, the observed variation in soil-moisture conditions in the inland mixed zone (from ~3% to >20%) gives rise to a patchy distribution of Anhythel and Anhyturbels, the latter showing evidence for some cryoturbation. In the field, boundaries between the coastal thaw zone and the inland mixed zone are plotted at the first indication of long-term ground stability (ancient soils, relic sand wedges, and ancient in situ ashfall, as measured from the coast), even though we recognize that isolated regions with traditional active-layer cryoturbation occur sporadically throughout.

**Mesoscale topography.** Straight (rectilinear) slopes topped by steep cliffs dominate the macroscale topography of the stable upland frozen zone. On average, these slopes are steeper (averaging ~25-35°) than those of the inland mixed- and coastal thaw zones, but this may be due in part to the local occurrence of cliff-forming lithologies. Minor variations in slope gradients are linked to the distribution and partial melting of perennial snowbanks. We have observed minor snowmelt on low-albedo rocks, even when atmospheric temperatures are <10°C. Although water from snowmelt can coat surface rocks and infiltrate the upper ~10 cm of coarse-grained soils, it is insufficient to incise channels and/or produce rills in unconsolidated sediment. Most of the meltwater that intermittently moistens soils in this zone evaporates within hours to days and does not freeze to form segregation ice. Gullies are rare in the stable upland frozen zone and occur concentrated only at a few localities, such as at the margin of low-albedo dolerite sills. The presence of ash-rich avalanche deposits, as much as 6.4 Ma and 11.3 Ma, as well as bedded ashfall dated at 12.5 Ma on steep valley walls, indicates that many colluvial slopes in the stable upland frozen zone are today inactive, and have been so since at least late Miocene time. Widespread colluvial deposits, most of which are at the angle of repose, are unmarked by rills, gelification lobes, and rotational slumps.

**Mesoscale topography.** Rock glaciers dominate the mesoscale topography of the stable upland frozen zone. Most rock glaciers occur downwind from dolerite-capped cliffs. Unlike rock glaciers of the inland mixed zone, which form via deformation of segregation ice, these rock glaciers originate through direct burial of glacier ice. Surface debris accumulates from rockfall and/or sublimation of dirty glacier ice. The latter brings englacial material to the ice surface at rates dependent on the thickness, porosity, and permeability of the overlying till cover. The stratigraphic contact between ice and overlying till is smooth and dry and lacks physical evidence for melting and subsequent refreezing - even where the buried ice lies just ~35 cm below the ground surface. It has been proposed that stagnant glacier ice beneath a thin till cover (~50 cm thick in central Beacon Valley, Quaternary Mountains is as much as 8.1 Ma.

Portions of rock glaciers in the stable upland frozen zone show morphologic characteristics signaling active flow, including steep frontal lobes and concentric surface ridges. However, horizontal ice-surface velocities for these rock glaciers are an order of magnitude lower than that measured for rock glaciers in the inland mixed zone. For example, the maximum surface velocity for active regions of the Mullins Rock Glacier in the Quaternary Mountains is ~40 mm per year. This low surface velocity likely reflects minimal snow and ice accumulation and emphasizes the important role of the katabatic winds in concentrating windblown snow from the Polar Plateau. Mass balance calculations, along with estimates for direct snowfall in this microclimate zone, suggest that >85% of the snow that falls onto the accumulation area of Mullins Valley rock glacier sublimes back to the atmosphere. Beheaded rock glaciers (those where deep topographic hollows, rather than ice, occupy snow-accumulation areas) commonly occur in the stable upland frozen zone - but not in the inland mixed zone. The topographic hollows, which commonly are lined with transverse, recessional moraines, mark the former position of alpine glaciers that, due to insufficient till cover, have completely sublimed; regions downslope, where debris cover has been/is sufficient to reduce sublimation, buried glacier ice still remains but commonly shows minimal evidence for lateral flow. Sublimation polygons, a special type of sand-wedge polygon, forms in tills overlying buried ice in the stable upland zone. Coarse-grained sand wedges that form at polygon margins allow for enhanced vapor diffusion of underlying ice, relative to that beneath fine-grained, compact, and undisturbed till at polygon centers. Ultimately, this enhanced sublimation leads to the development of ice-cored till mounds separated by deep “sublimation troughs” (>3 to 5 m deep). A negative feedback prevents runaway sublimation and complete loss of glacier ice: as sublimation troughs deepen, they become preferred sites for collection of windblown snow; the downward flux of vapor from the base of these snowbanks, particularly during months when subsurface ice temperatures drop below atmospheric temperatures, creates a thin layer of ice that caps the buried glacier and effectively shuts down loss of remaining glacier ice.

Soils within the stable upland frozen zone are extremely dry (<3% soil moisture) and lack stratigraphic evidence for traditional active layer deformation. They are best classified as Anhythel. As for most regions in the Dry Valleys, intermittent moistening of near-surface soils from minor snowmelt on low-albedo rocks leads to the development of subsurface salts. Old soils >10 Ma contain salt horizons as much as 20-cm thick. Salts are enriched in nitrates and sulfates, reflecting the dominance of westerly katabatic winds off the Polar Plateau. Detailed chemical analyses show that below 20-cm-to-30-cm depth many soils of the stable upland frozen zone retain undisturbed salt horizons that accumulated as much as 13 Ma. This indicates long-term soil stability, limited leaching, and minimal cryoturbation over million-year timescale.
Mission Summary: ARES is a proposed Mars Scout Mission designed to use an airplane platform to fly over the surface of Mars and make measurements of key science objectives. ARES extends and complements NASA’s Mars Exploration Program (MEP) by returning benchmark measurements in two critical scientific themes: 1) Crustal Magnetism and 2) Near-Surface Atmospheric Chemistry which embody six primary science objectives:

1. High spatial resolution crustal magnetic survey.
2. Crustal magnetism source structure.
3. Underlying geology and mineralogy
4. Role of water vapor in the Mars atmospheric chemical cycle.
5. Chemical coupling between the atmosphere and surface.
6. Atmospheric chemical and isotopic composition and evolution.

ARES provides a fundamental scientific insight into the origin and evolution of magnetic fields, early planetary crustal formation processes, the role of water in the Mars atmospheric chemical cycle, Mars’ volatile and climate history, and the viability of Mars as a biosphere conducive to past or present life.

ARES offers the opportunity to target and explore specific regions of highest scientific interest, returning previously unobtainable measurements from a vantage point 1.5 km above the surface across 500 to 850 km of diverse terrain.

ARES provides NASA with the opportunity to elicit an immediate and significant scientific response while inspiring the next generation of explorers through a large and complementary Education and Public Outreach (E/PO) effort.

Science Payload: A comprehensive suite of science instruments provides critical, high-priority COMPLEX and MEPAG measurements:

- Magnetometer (NASA GSFC) Magnetic field with 2 km spatial resolution and 1 nT sensitivity.
- Context Camera (Malin Space Science Systems) Contiguous imagery of science traverse over 3 km horizon swath.
- Point Spectrometer (NASA LaRC) Surface spectra of ground track with 2.0 x 2.7 m spatial resolution, 430-1020 nm spectral range with 9 nm spectral resolution.
- Mass Spectrometer (Univ. of Texas, Dallas) Atmospheric constituents measured in-situ with 1 ppb sensitivity.
- Atmospheric Data System (Aurora Flight Sciences) Regional-scale pressure, temperature, density, and wind.
- Video Camera (Malin Space Science Systems) Nar video rate imagery of critical deployment events, aerial platform and science traverse.

Science Advancements: ARES science objectives, measurement requirements and data products flow directly from NASA’s strategic plan for Mars Exploration, the COMPLEX and MEPAG reports. ARES science return is derived from orders of magnitude improvement in spatial resolution and precision.

- Two orders of magnitude higher spatial resolution magnetic survey than provided by Mars Global Surveyor (MGS), with the ability to resolve the crustal magnetism source structure.
- One order of magnitude higher spatial resolution spectroscopy than Mars Reconnaissance Orbiter (MRO) CRISM instrument.
- First direct measurement of near surface water vapor and chemically active gas concentrations.
- More than one order of magnitude higher precision isotopic ratio measurement than Viking Landers.

Mission Architecture: ARES implementation includes robust technical margins and flight proven systems to achieve flexibility and success. The ARES mission delivers an airplane to Mars with a spacecraft derived from Genesis heritage, and an entry and descent system derived from Pathfinder and Mars Exploration Rover. The rocket-propelled airplane autonomously completes a 500 to 850 km pre-planned, inertially-navigated science traverse. Science and critical event data is relayed to the ARES spacecraft as it performs a Mars flyby and is returned to Earth within 18 hours of the airplane flight through the 34 m DSN subnet. All data is disseminated and archived within 6 months of receipt.

Launch Vehicle: Delta 2925
- Launch Dry Mass: 522 kg
- Launch Wet Mass: 678 kg wet (35% margin)

Mission Design:
- Flyby spacecraft design and operations
- Direct entry trajectory
- Critical events link margin: 7 dB
- Airplane to S/C link margin: 12 dB
- S/C to Earth link margin: 3 dB

Airplane
- Mass: 66.5 kg dry (32% reserve), 125 kg wet
- Altitude: 1.5 km AGL, Ground Speed: 140 m/s
- Wing span: 6.2 m
- Maximum Duration: 101 min
- Maximum Range: 850 km
- Peak Power Margin: 53%, Cost Reserve: 49%

Airplane Robustness: ARES airplane size allows:
- Use of traditional design, test and manufacturing techniques.
- Credible aerodynamic performance and stability predictions.
- Flight-proven systems and a large payload volume margin.
- A majority of the implementation risk to be contained in a single well-defined and testable event: airplane deployment.
- Robust stability and control margins to accommodate large uncertainties in the Mars environment.

Management Approach: Streamlined, effective management approach ensures rapid decision-making with strict accountability. ARES team members have well defined responsibilities.

- Lockheed Martin Astronautics: Spacecraft, Entry System.
Figure 1. Artist's conception of ARES performing its survey over the surface of Mars.

Figure 2. Artist's conception of ARES collecting data over the surface of Mars.

Figure 3. Video still from the successful deployment of a half-scale model of ARES at Mars-relevant atmospheric pressure above Oregon, USA.

Figure 4. Scientific payload onboard ARES.

Figure 5. ARES bridges critical scale and resolution measurement gaps in the core MEP.
Background and History: The successful Spirit and Opportunity rover traverses highlight the context provided by the rover moving from place to place and underscore the basic importance of human telepresence at scales commonly experienced by geologists on Earth. Indeed, recent developments in computer science (visualization, rendering and immersive virtual reality) and acquisition of high-resolution data sets of planetary surfaces have provided the capability to place scientists virtually on the surface of planets, where they can attack important multidisciplinary scientific problems that have heretofore gone unexamined.

In the past, high-resolution terrain visualization and other forms of planetary data visualization have taken separate paths. High-resolution 3D representations of terrain data have typically been computed off-line as fixed sequence movies, while at the same time planetary data sets are commonly presented for query in mapping format. While many presentation tools exist for producing batch animations and still images, we know of no interactive tool that offers the functionality proposed by ADVISER—namely, real-time visualization and user interaction with high-resolution planetary datasets including geology analysis tools that facilitate exploration of the data to help address scientific questions. Therefore, we see the demonstration of this synergism as a fundamental first step in establishing the importance of these techniques so that they can be developed further and be routinely applied to scientific problems.

Establishing the scientific imperative for visualization and immersive virtual reality (IVR): Geologists explore the Earth primarily through fieldwork and analysis of the geological record at various points on the surface of the Earth. They then integrate these individual points of understanding about the Earth’s surface through more synoptic analyses, often aided by the integrating perspectives seen from image and topographic data acquired from Earth orbit. Planetary geoscientists commonly work toward an understanding in the reverse order. The distances and times involved dictate that the first data from individual moons and planets comes from flybys and orbital spacecraft, perhaps in some cases evolving towards the deployment of a few landers and rovers, and for the Moon, human explorers. Now that we have global data sets for the Moon, Mars and Venus, comprehensive regional data sets for many of the outer planet satellites, and soon to be obtained global data for Mercury, we can begin to undertake the detailed exploration of planetary surfaces that is required for the full understanding of the evolution of planets.

How do we accomplish this? In only a very few cases can we expend the resources to put a lander and rover, and thus our eyes and ears, down onto the surfaces of the planets. The successful Pathfinder and Mars Exploration Rovers are testimony to the exciting results that can be obtained by such real surface exploration. Fortunately, developments in advanced visualization and immersive virtual reality environments have created the ability to place the geoscientist back down on and near the surface, to visit virtually any part of the planet they wish to see, and to regain the perspective that is the foundation for the understanding of the geological relationships necessary to unlock the record of the history of the planets.

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The ADVISER Problem Solving Environment: ADVISER is a “problem solving environment” (PSE) for planetary geosciences. We define the PSE as a set of tools that provide the planetary geoscientist with the capability to explore and analyze data as if they were on or near the surface of a planet. The ADVISER PSE has four basic parts:

1) Geoscientist on the Surface: Visualization capabilities that enable the placement of the geoscientist onto the surface and near surface environment through immersive virtual reality (IVR) (Figure 2) or related desktop capabilities using topographic data and surface rendering programs.
2) Importation and Visualization of Multiple Data Sets: On-demand importation, co-registration and overlay of relevant image format data sets to enhance the eyes of the geoscientist and their ability to correlate and interpret data for scientific analysis. This includes high-resolution data sets such as MOC, THEMIS, HRSC, OMEGA (and derived data products), CRISM (in the future) and synoptic GRS, TES, and other types of derived regional data sets (slope, mineralogy, temperature, etc).

![Image 1](image1)

Figure 2. Using our prototype system running in a Cave immersive virtual reality system, two participants collaboratively examine a trough and a distant polar plateau. Note the already detailed Mars Orbiter Laser Altimeter (MOLA) terrain is made clearer by attention to lighting effects. Fractal geometry added to the closest terrain can be turned on or off; while not accurate terrain data, its important benefit is enhanced stereo viewing of the 3D form through the subtle textured look it creates.

3) Development of the Field Kit: This is analogous to the geologist's field kit and will consist of a set of software functions that has the capabilities commonly carried out by the geologist in the field using things such as the Brunton compass, altimeter, etc. It will consist of selectable items including elevation of any point chosen, relative elevations, slope determinations, strike and dip of planes defined by several points on the surface (say a continuous layer in the polar layered terrain or an outcrop in Valles Marineris), ability to produce variable lighting and different illumination geometries, ability to determine wind directions and velocities.

4) Development of Ancillary Virtual Field Instruments: This is analogous to the additional tools that the geoscientist carries in the field such as cameras, GPS, and field notebooks. Specifically, we have developed the capability for: a) Virtual Photography: This capability permits documentation of individual images or video streams from a menu and the ability to store them for incorporation in electronic form for export to other remote collaborators, storage for further analysis or insertion in manuscripts and meeting presentations.

b) Virtual GPS: This virtual global positioning system permits exact location at any point in the analysis and permits tracking of traverses and storage of these data so that they can be readily repeated or exported to other systems for collaborators to use. c) PDA: This is the equivalent of the field notebook and permits the investigator to record a host of information including notes and documentation as well as any of the data derived from the Field Kit (Figure 3). This is a major part of the written record of the use of the exploration tool, and serves as the basis for the documentation of the analysis and solution of the scientific problems. These data form a permanent record and permit the synthesis of information for professional publication. Together, these four parts form the foundation of the ADVISER PSE.

**Planetary Geoscience Demonstration Projects:**
We are undertaking a three-pronged approach of geo-sciences investigation to demonstrate the scientific usefulness of visualization and IVR in planetary geosciences with the ADVISER PSE. The three projects will use the visualization of large martian terrain and image/map databases and will provide the basis for developing a visualization system generalized to complex GIS terrain, image and volumetric data. This system will integrate state-of-the-research geometric and volumetric rendering software with parallel graphics hardware to allow interactive exploration and analysis of data at the highest available resolution and with synthesis of all available modalities. Research will be carried out to develop science domain-specific instrumentation to support interactive query and analysis. The demonstration projects for our geosciences investigation will be a) Mars Polar Evolution, b) Mars Tropical Glaciers, and c) Noachian Hydrological Cycle.

![Image 2](image2)

Figure 3. Navigating terrain using a wireless PDA. Here a user can click on a point they wish to explore, and then they are automatically navigated to that area. Fine-grain navigation could be accomplished in a point-and-fly style metaphor subsequently. The PDA can also be used for a variety of other user interface elements. The terrain appears blurred because it is displayed as a field-sequential stereograph for viewing with LCD shutter glasses.
LARGE IGNEOUS PROVINCES ON MARS: ASCRAEUS MONS
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Large igneous provinces are common on most of the terrestrial planets, including the Earth, Mars, Venus and the Moon. However, there are fundamental differences in the characteristics of large igneous provinces on these planetary bodies. For example, while large shield volcanoes are absent on the Moon, Mars is characterized by extremely large shields that have formed on the stable lithosphere. Venus on the other hand shows evidence for a rapid, planet-wide and catastrophic resurfacing event and Head and Coffin [1997] argued that such an event could be the equivalent of a planet-wide large igneous province. Large igneous provinces (LIP) are characterized by voluminous emplacement of mostly mafic plutonic and volcanic rocks that are not related to seafloor spreading. Rather, large igneous provinces such as the Deccan basalts or the Columbia River basalts are formed by transient large-scale activity occurring in geologic settings such as continental flood basalt provinces, volcanic passive margins, oceanic plateaus, ocean basin flood basalts, and large seamount chains [Head and Coffin, 1997]. Numerous studies indicate that large igneous provinces are closely linked to mantle dynamics [e.g., Coffin and Eldholm, 1994], but details such as emplacement rate, relationship to tectonics, and crustal processes remain under discussion.

On a dynamic planet such as Earth with its constantly moving plates and renewal of the lithosphere, mantle plume dynamics are often obscured. In contrast, one-plate planetary objects such as the Moon, Mars, Mercury and Venus can illustrate the long-term influences of mantle plumes and their variations under different thermal condition in space and time [Solomon, 1977]. Because the lithosphere is not recycled on these planets, we have a very long geologic record that allows one to investigate the chronology and episodicity of large igneous events and provinces. For these reasons, Head and Coffin [1997] argued that the planetary examples of large igneous provinces, in concert with detailed studies of terrestrial large igneous provinces, should help to develop and test models for the origin, formation, and emplacement of LIPs, and to decipher the influence of plate tectonics and deeper mantle and crustal processes on LIPs.

Mars is the focal point of numerous recent space missions, including Pathfinder, Mars Global Surveyor (MGS), Mars Odyssey, Mars Express, and the two MER rovers. In concert with early spacecraft data from the Mariner and Viking missions, the returned data indicate that 58% of the surface of Mars, an area of ~0.84 x 10^8 km^2, are covered with volcanic material [Tanaka et al., 1988]. Greeley et al. [1987] estimated that approximately 2 x 10^8 km^3 of volcanic rocks were emplaced on the surface of Mars. Stratigraphic studies show that most of the geologic surface activity, including volcanism, took place in the first half of the solar system history, with some volcanism and eolian and glacial activity continuing well into more recent times. In fact, the most recent spacecraft data show evidence that some lava flows might be as young as 10 - 100 million years [Hartmann and Neukum, 2001]. Today Mars is characterized by the largest known volcanoes in the solar system. A large number of these volcanoes sit on top of an extremely large volcanic rise that covers about 20% of the planet's surface. The Tharsis rise forms a broad dome of ~4000 km in diameter, rises as much as 10 km above the surrounding terrain, and covers an area of >6.5 x 10^6 km^2 [Head and Coffin, 1997]. Detailed mapping of the Tharsis rise indicates that it formed over 10^8 - 10^9 years, a much longer period of time relative to many terrestrial LIPs, which formed over 10^5 - 10^6 years. The Tharsis bulge is commonly interpreted to be the result of a long-lasting large mantle diapir that due to the lack of plate tectonics on Mars, had enough time to uplift the lithosphere and initiate tectonic faulting and volcanism [e.g., Banerdt et al., 1992; Breuer et al., 1996; Harder, 1998; Smith et al., 1999; Zuber et al., 2000; and references therein].

A comparison of Olympus Mons, the largest volcano of the Tharsis rise, with the Hawaiian shield illustrates the enormous dimensions of this volcano. Its base is several hundreds of kilometers wide, approximately the size of Arizona, and its summit is at more than 20 km elevation, three times as high as Hawaii. Its volume of 2 x 10^8 km^3 (above its base) is an order of magnitude larger than that of Hawaii (1 x 10^7 km^3), which is composed of several individual shields. In addition to Olympus Mons, there exist numerous other large volcanoes on Mars, including the three Tharsis Montes, Ascreaeus, Pavonis, and Arsia Mons, as well as Alba Patera, Elysium Mons Hadriaca Patera, and Amphitrites Patera, to name only a few. Very often these volcanoes exhibit a wide range of rift zone development, internal deformation related to lithospheric loading and flexure, flank and slope failure, and summit caldera development [Carr, 1973, 1981; Hodges and Moore, 1994; Head and Wilson, 1994; Crumpler et al., 1996].

How do we account for these unique characteristics of Martian shield volcanoes? On Mars, a one-plate planet, the lithosphere has been stable and has not moved laterally over very long times in Martian history. Thus the regions of melting in the mantle concentrated their effusive products in a single area, rather than having them spread out in a conveyor-belt-like fashion, as in the case of Hawaii. Over time, these melt products accreted vertically into huge accumulations that loaded the lithosphere and caused flexure, deformation and edifice flank failure. [Carr, 1973, Head and Coffin, 1997]. Another factor contributing to the large height of the Martian volcanoes is the very thick elastic lithosphere associated with these volcanoes; the volcanic load and the underlying lithosphere did not subside at a rate that would limit their heights.

New high-resolution multispectral stereo data from the HRSC camera on board the European Mars Express mission allow us to take a very detailed look at Martian large igneous provinces. The HRSC camera is a linescan.
LARGE IGNEOUS PROVINCES ON MARS: ASCRAEUS MONS  

HIESINGER, H., HEAD III, J.W.

The HRSC camera acquires images at spatial resolutions of about 10 m/pixel and is complemented by a Super Resolution Channel (SRC) with a 1024 x 1032 pixel frame CCD, which obtains images of about 2.3 m/pixel from an altitude of 250 km at periapsis. The latest HRSC images can be found at http://berlinadmin.dlr.de/Missions/express/. We used HRSC data from orbit 16 to investigate lava flows on Ascraeus Mons, one of the three Tharsis Montes [Hiesinger et al., in prep.]. The Tharsis Montes are the locations of some of the youngest volcanic deposits on Mars [Scott and Tanaka, 1986]. Compared to earlier studies, the high spatial resolution of the HRSC data allowed us to map 25 late-stage lava flows and to measure their dimensions, as well as their morphological characteristics in greater detail. In the HRSC images we observe several lava flows with well-defined leveed channels on the flanks of Ascraeus Mons, some of which are truncated by the collapse of the calderas and extend for tens of kilometers downslope. On the basis of morphologic similarities between terrains on Ascraeus Mons and terrestrial shield volcanoes, Zimbelman and McAllister [1985] proposed that individual prominent flows on Ascraeus Mons are a’a flows and the planar areas adjacent the flows are pahoehoe flows. Our estimates of the yield strengths for the young flows are on the order of ~2.8 x 10^4 Pa. These values are comparable to estimates for terrestrial basaltic lava flows, and are in good agreement with estimates of Zimbelman [1985] derived for a small number of lava flows on Ascraeus Mons. Our investigation indicates that the effusion rates for the studied Ascraeus flows are consistent with findings of Zimbelman [1985] that indicate effusion rates of 18-60 m³s⁻¹, with an average of 35 m³s⁻¹. On the basis of our estimates of the effusion rates and the measured dimensions of the flows, we calculated that the time necessary to emplace the flows is on average on the order of hundreds of days. Viscosities were estimated on the basis of yield strengths and effusion rates, yielding average values of ~5 x 10⁶ to ~6 x 10⁵ Pa-s.

In summary, with the new data we have the opportunity to better understand the environments, associations and styles of emplacement of LIPs on Mars and also to gain a better understanding of the relations to the internal structure and the implications for the plume structure. In addition, the new data allow us to estimate more precisely the areas and volumes of LIPs, the duration and rates of their emplacement, their petrologic evolution, their relation to the geologic history of Mars and their influence on the atmosphere and the environment.


Figure 1: Topography from Mars Orbiter Laser Altimeter (MOLA) on board Mars Global Surveyor spacecraft. (Credit: MOLA Science Team; http://ltpwww.gsfc.nasa.gov/tharsis/map_lab.html)
Is Mars cheating about its ice? J. Helbert\(^1\) and J. Benkhoff\(^1\), \(^1\)Institute for Planetary Research, German Aerospace Center DLR, (Rutherfordstr. 2, Berlin-Adlershof, Germany, joern.helbert@dlr.de), Research and Scientific Support Department, ESTEC, (Keplerlaan1, 2201AZ Noordwijk ZH, The Netherlands).

**Introduction:** While Mars has been considered for a long time a dry place, this view has changed in recent years. This started mainly after the MOC imagery showed features like the gullies and morphological features which can be associated with glacial activity. Now the motion was discussed that at least small amounts of water or ice had been present in the recent past on Mars. Still, the common notion was that Mars today is a dry place. With the excellent dataset of the Gamma and Neutron spectrometer (GRS and HEND) on board of Mars Odyssey this view had to be corrected. The instrument detected water abundance of at least 5wt% in the equatorial regions of Mars and this water is found within the first 2m below the surface, the penetration depth of the instrument.

**Water on Mars:** There are three main explanations for this observed amount of water which are not mutually exclusive. Some of the water measured is most likely adsorbed water. While it is still unclear how much water the Martian soil can adsorb, this mechanism can not explain the high abundances measured in some place. We might see highly hydrated minerals. Some of the suggested minerals are indeed capable of holding large quantities of water. The last and maybe most exciting possibility are near surface ice deposits. However if it is ice, the question is, how did it survive close to the surface under the hyper-arid conditions we encounter on present day Mars. And how much ice is there really on present day Mars?

**Ice on Mars:** Until today we have seen ice only at the polar caps and only this year did we get the first direct measurements of ice abundances by the PFS and OMEGA instrument on Mars Express. We do not have any direct evidence for ice at lower latitudes. From the GRS and HEND measurements we know that the polar caps extend under the surface much further than previously expected. One might assume therefore that near surface ice deposits we see at low latitudes are literally only the tip of an iceberg and the Mars might have a global ice reservoir in shallow depths. If this would be the case, Mars would be a wet planet which is just temporarily frozen. Another less dramatic scenario is the assumption that ice deposits at low latitudes are remnants of the last Martian ice age. The change in the obliquity of Mars can lead to a redistribution of ice across the planet. So maybe we observe today a transition state, in which we only see the dwindling remains of equatorial glaciers. If the ice within the top 2m has survived until today, this would however imply that these regions have been covered by large amounts of ice during the last ice age. Both ice related scenarios would imply that Mars has, or at least had in the very recent past, large quantities of ice on or close to the surface.

**Enrichment of ice:** While working on model calculations for the stability of ice on Mars today we discovered a possible third scenario. Most models used to study the stability of ice in the Martian soil assume a homogenous soil with constant thermo-physical properties with depth. Furthermore most of the models do not actually model the water vapor transport in the subsurface, but instead using an equilibrium approach and derive the depth of the ice table based on the subsurface temperatures. With the Berlin Mars near Surface Thermal model we have worked on a different approach. The model actually models how the ice table would move over time, including a detailed treatment of the diffusive processes and the energy transport.

We have studied cases where the soil consists of layer with very different thermo-physical properties. One of the scenarios we have looked at is a low thermal conductivity dust layer on top of a sand layer with a significantly higher thermal conductivity. Such configurations can be found for example in the Terra Arabia region. For this case we observed the formation of an ice lens at the boundary between the material, effectively closing the pore space and significantly reducing downward diffusion. This leads to an actual enrichment of ice within the top layer. Depending on the thickness of the dust layer and the parameters used we can get an enrichment of a factor of 2 or more in the first 2m below the surface.

**Conclusions:** We will discuss the implications of such a scenario on our understanding of ice on Mars. The ice enriched layer might in a sense fool the GRS and HEND instrument and it might show a morphological behavior very similar to a rock glacier. The process of enriching the ice significantly slows the movement of the ice table to greater depth and can therefore stabilize ice over several thousands of years close to the surface.

This means we might indeed see today the remains of the last Martian ice ages, but the amount of ice moved across the planet can be significantly smaller than previously thought. Furthermore some of the “young” glacial feature we see today would have been formed not during, but after the last ice age and might even exist until today.
MULTI-RING BASINS: MODELING TERRESTRIAL ANALOGUES. B. A. Ivanov, Institute for Dynamics of Geospheres, Russian Academy of Science, Leninsky Prospect., 38-1, Moscow, Russia (baivanov@idg.chph.ras.ru).

Introduction: History of the lunar surface is a part of the early evolution of terrestrial planets. Important footprints of the early lunar history are impact multi-ring basins. Similar giant impact structures are known on Earth. The aim of the presented project is the analysis of terrestrial giant impact structures origin as the starting point for the following analysis of origin and structure of lunar multiring basins and their role in formation of lunar surface.

Lunar and Terrestrial Basins: Lunar basins created ~3.8 Ga ago and earlier [1] definitely play an important role in formation of the modern lunar landscape [2]. Currently lunar basins may be studied by remote sensing only. Ancient impact basins on Earth have not survived. Only a few large impact structures are currently known including large impact craters Chicxulub, Sudbury, and Vredefort [3]. Following simple scaling rules [4] morphologic types of impact structures on different planets are similar for crater sizes with the same value of gD (g is gravity acceleration, D is the impact structure diameter). Hence, in the first approximation, terrestrial impact structures with diameters 100 to 300 km may resemble lunar craters with ~6 times larger diameters, i.e. craters with diameters 600 to 1800 km. The similarity definitely cannot be an exact one as target layering (crust thickness, for example), thermal profile and other important parameters may complicate the simple scaling. However, the study of largest terrestrial craters as possible analogues of small lunar basins looks attractive as we have a lot of data on crater subsurface structure [3].

Numerical Modeling: In the presented project numerical modeling of formation of selected terrestrial impact structures is carried out with following comparison of model results (predictions) with available geological and geophysical data accumulated for known largest terrestrial craters. SALEB hydrocode is used to model the vertical impact of a spherical projectile into a target, presented with 2 or 3 layers of different materials (e.g. upper crust, lower crust and mantle) initially balanced in a gravity field. ANEOS equation of state in tabulated form and other important parameters may complicate the simple scaling. Hence, in the first approximation, terrestrial impact structures with diameters 100 to 300 km may resemble lunar craters with ~6 times larger diameters, i.e. craters with diameters 600 to 1800 km. The similarity definitely cannot be an exact one as target layering (crust thickness, for example), thermal profile and other important parameters may complicate the simple scaling. However, the study of largest terrestrial craters as possible analogues of small lunar basins looks attractive as we have a lot of data on crater subsurface structure [3].

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Perspectives: The presented results of the numerical modeling proved the basis to generate a set of predictions, which may be compared with observations. An iterative process of the model improvements to fit (and explain) observations allows us in a close future to find the best fit to each of investigated impact structures. The best model approaches will allow us to make more or less robust estimates with the following numerical modeling of lunar impact basin formation.

Acknowledgements: The project is supported by RFBR grant # 04-05-64338.

Fig. 1. Chicxulub. The comparison of the “best fit” model and geophysical model [1].

Fig. 2. Chicxulub: modeled geometry of melt and ejecta zones (initial position of Lagrangian tracers) and final position of initially horizontal layers of sediments (blue) and crystalline rocks (brown). Also the stream tube of ejecta deposited at the distance of Yax-1 drillhole is shown.

Fig. 3. Popigay: 100-km in diameter crater in crystalline rocks. Computed position of impact melt is shown as red dots.

Fig. 4. Vredefort: The proper diameter of the middle crust “neck” at the assumed erosion level of ~10 km corresponds to the rim basin diameter of 180 km.

Fig. 5. Vredefort: the comparison of computed diameters of middle crust rock “neck” at ~8 to 10 km erosion level (red circle) and annual syncline (blue circle) with the simplified geological map [13]. Pink dashed circle is for estimated rim position.
**Martian Upper Crust Strength Estimates.** B. A. Ivanov, Institute for Dynamics of Geospheres, Russian Academy of Science, Leninsky Prospect., 38-1, Moscow, Russia (baivanov@idg.chph.ras.ru).

**Introduction:** Impact craters observed on the surface of Mars give us an important information about the geologic history of the planet. For example, crater size-frequency distribution allows us to estimate retention age of specific areas, extent and morphology of fluidized ejecta blankets reflect (at least, qualitatively) the presence of ice/water in upper layers of Martian crust. At the same time the presence of ice water in target rocks should influence also crater morphology and morphometry. However, the conversion of observed details of impact structures demands better knowledge of cratering mechanics in Martian rocks. The presented project is aimed to make a review of possible mechanical properties of near surface Martian rock. Plausible hypothesis about mechanical properties should facilitate the analysis of observational data on Martian impact craters, especially if one use the technique of computer numerical modeling of impact cratering.

**Upper crust as a target for impact cratering:** The upper crust of Mars is assumed to be brecciated during the late heavy bombardment period. The brecciated autochthon rocks may be covered by ejecta, volcanic, aeolian, and, in some areas, by marine/lake sediments. Porous space in all rocks, listed above, may be filled with water (brine) and water ice depending on the local pressure-temperature conditions. The cratering mechanics depends on target density, compressibility, initial strength of material, and dissipative properties (internal friction, effective viscosity) of fragmented material, involved into a crater-forming motion. The presence of ice in some layers of target rocks open a possibility to influence cratering in two opposite directions: low-temperature ice in permafrost layers increase the porous rock strength, while possible shock melting of ground ice make, in general, result in the strength/friction decrease. Below we make a draft of a scheme to construct initial conditions for the formation of impact craters of various diameter in different geologic provinces of Mars.

**Temperature and pressure gradient:** Unknown Martian heat flux is commonly estimated as ~30 mW m⁻² [1]. With a lot of assumptions the heat conductivity may be estimated as ~2 W m⁻² K⁻¹ [1]. The resulting near-surface thermal gradient is estimated as ~15 K km⁻¹. Simple pressure estimate give \( p = \rho g z \approx 9.3 \text{ Mpa} \) for depth \( z \) of 1 km (assuming density \( \rho \approx 2500 \text{ kg m}^{-3} \) at gravity \( g = 3.72 \text{ m s}^{-2} \)). Hence the range of pressures along the assumed porous space of the upper 10 km of Martian crust [1] is growing from atmospheric pressure at the surface to ~100 Mpa (1 kbar). The temperature grows up from the surface mean annual value (~150 to 170K at high latitudes and 210 to 220 K at low latitudes) at +150 K approximately.

**Water/ice content:** According to the model of hydrosphere [1] one can list possible variation of the saturation state of porous rocks: (i) upper layer, filled with ice which exist only in high latitudes (low T, low ambient pressure); (ii) dry (desiccated) layers in low latitudes; (iii) water saturated low horizons (T above freezing point of water/brine); (iv) possible intermediate layer of dry porous rocks if the total water column does not fill all the porous space. The model [1] assumes 100% saturation in layers (i) and (iii). One should have reasonable estimates of rock strength for all listed layers before to start any modeling. For dry porous material one can use a numerous experimental data for terrestrial rocks. As an example the shear strength of basalt specimens in triaxial tests is presented in Fig. 1.

**Ice strength:** Shear strength data for pure ice are presented in Figs. 2-4. The strain rate dependence of strength as well as structure of ice result in relatively large data scatter. However it is evident that the ice strength grow approximately 5 to 10 times when temperature decrease from melting point to ~150K. For lower temperatures the strength is approximately constant.

**Permafrost strength:** The permafrost strength depends on temperature (being larger at low temperatures) and size-frequency distribution of rock/soil particles reflecting available particle surface absorbing (unfrozen) water. In general the permafrost strength follow the ice strength trend. For many materials strength is 1.5 to 2 times above the pure ice strength at the same temperature. After failure broken material behave as Coulomb material with a friction coefficient depending on ice content (i.e. on the rock porosity). For high ice content the friction is close to pure ice friction which is ~twice less than general rock dry friction. For stress state invariants of shear \( T_s = [(\sigma_1 - \sigma_2)^2 + (\sigma_2 - \sigma_3)^2 + (\sigma_3 - \sigma_1)^2]/6 \) and pressure \( p = (\sigma_1 + \sigma_2 + \sigma_3)/3 \) the friction coefficient \( (T_s/p) \) for rocks is of ~0.7, while for pure ice it is of ~0.5 (at low \( p \)) to ~0.2 (for \( p \geq 20 \text{ Mpa} \)). However for large strain and strain rate the dynamic friction of ice is a complicated dynamic process, possibly with shear melting (e.g. [4]).

**Perspectives:** The compiled data shown in Figures allow us to construct a set of possible model targets for Martian impact cratering. Depending on latitude, assumed change with depth of porosity, one can present these targets as layers, where the initial strength is controlled by presence or absence of ice below freezing point and presence/absence of water.
saturated fractured rocks in upper ~10 km of Martian crust. It should improve previously published models (e.g. [5]) where dynamic strength of Martian crust assumed to be constant (10 to 20 MPa).


Fig. 1. Shear strength of initially intact basalt specimen in triaxial tests. One can assume that after brittle failure strength of fragmented material is controlled by dry friction (Bayerly law [6]).

Fig. 2. Low-temperature ice shear strength (black dots) and friction (compressive strength of specimens with a fracture) [2,7].

Fig. 3. Low pressure range of data shown in Fig. 1 with the addition of data for large temperatures ([3] for –40°C, pink; [8] for –20°C, crashed ice).

Fig. 4. Unconfined 1D strength of ice and frozen sand, silt, and concrete samples as a function of temperature. At temperatures of ~220K (~50°C) frozen silt may be stronger than frozen concrete.

Fig. 5. Tensile strength of ice and permafrost specimens dependence on temperature for most fast tests. Black and gray symbols are for pure ice. Blue symbols and curves are for permafrost.
PLANETOLOGY FOR EVERYBODY: THE SERIES OF UNDERSTANDABLE TO ALL STORIES AT “VOKRUG SVETA”, THE NATION’S FIRST COGNITIVE MAGAZINE

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Introduction: The series of papers aimed to tell general public the elementary stories about the planetary bodies of the Solar system have been published at “Vokrug Sveta” magazine in 2003-2004. The review of the endeavour is presented.

The Magazine Overview: “Vokrug Sveta” (“Around the World”) popular science magazine have been founded in 1861 (the same year the serfage have been canceled in Russia). “Vokrug Sveta” is considered to be the nation’s first cognitive magazine. During 143 years 2769 issues have been published by this

Fig. 1. Cover of August 2003 issue of “Vokrug Sveta”, featuring the story about the first manned flight to the Moon. The main title: ‘WALKING ACROSS MARE TRANQUILLITATIS: What Jules Verne failed to take into account’.

Microsymposium date. The current monthly number of copies is about 240,000. It provides an estimation of the potential readers as 2,000,000 persons. The magazine contains 200 pages of A4 format in polychromatic printing on glossy paper. Each issue usually contains about 15 rubrics, some of which are regular and some – variable. One of the regular rubrics – “Planetarium” – is devoted to astronomical and space research themes. The online version of the magazine take place at Internet with the archive of issues beginning from No. 1/2002.

Popular Planetologic Papers: Since 2003 there is a plan to undertake a continuos publication of papers about the planets of the Solar system. Such papers are planned to appear at “Vokrug Sveta” 3 to 4 times during the year. By this date six papers has been published: “The distorting mirror of the Earth” (about planet Venus; No. 6/2003), “The Grand Odyssey” (about the first manned flight to the Moon; No. 8/2003), “The space Liliputians” (about asteroids; No. 10/2003), “And more than a year the day is lasting” (about planet Mercury; No. 12/2003), “Discovered twice” (about planet Uranus; No. 6/2004), “Disturbances of the heavenly solidity” (about earthquakes outside the Earth; No. 9/2004). Three times the planetary themes find the place on the cover of the magazine (Fig. 1, 2, 3).

Fig. 2. Cover of June 2004 issue of “Vokrug Sveta”, featuring the story about the planet Uranus. The main title: ‘THOSE, WHO GAVE BIRTH TO THE GIANTS: Where Uranus is rolling to’.

Style of Presentation: The stories about planets are presented at understandable to all form. It suppose absence of scientific formulae and complicated termi-
nology. The facts are telling with close-to-everyday speech. Pictorial examples and comparisons with well-known things take place to help the understanding of unfamiliar facts. The story contains contemporary review of the knowledge about the planet, either about the aspect of planetary nature (e.g. earthquakes), numerical data, description of a planet, interesting features from the history of investigation, plans of the future spacecraft flights. The story is both cognitive and amusing. There are a lot of color pictures, mainly photos, in each paper. Other the types of illustrations include maps, diagrams, crossections, etc. Most of them are original, designed specially for this paper, either are redesigned from black-and-white schematic drawings into colorful attractive view. Each paper, published so far, usually occupies 10 pages. The beginning two-page spread is aimed to feature the principal idea of the story with the large and attractive illustration. The text is arranged in two categories: the main story and the additional parts. The latter are the small stories, which are independent from the main text. Each paper provides reader with comprehensive and up-to-date story on the selected topic.

Fig. 3. Cover of September 2004 issue of “Vokrug Sveta”, featuring the story about the earthquakes outside the Earth.

The main title: ‘SHAKING OF THE SUN: What does space seismology study’.

Online Edition: Three months after publication of the hard-copy edition, an online edition is placed at the Internet on the magazine’s site www.vokrugsvetaru.ru. The online version involves the main text, the additional stories, and most of the illustrations. To fit the net requirements the paper is rearranged and illustrations obtain smaller size. Some of the illustrations are designed to be presented at larger size after clicking on them with the computer’s cursor. Some illustrations, placed in parer version of the magazine are omitted in online version. Currently there are five planetological papers of the series described here, which are placed at online version of “Vokrug Sveta”. They could be found on the magazine’s site either by the issue numbers in “Archive” section, or through the “Search” engine with the author’s name (Burba – in Russian letters).

Distribution: The paper copies of the magazine are distributed by subscription (podpiska@maart.ru) and on newsstands. The latter distribute the current issue of the magazine beginning from the first day of each month. The distribution of the previous month’s issue is stopped just when the new issue appears. The earlier issues may be obtained by the postal order from vokrugsvetarzdzr.ru, either could be bought from the store at Moscow (sl@maart.ru).

Acknowledgements: The papers on planetology, appeared by this day at “Vokrug Sveta”, obtained numerous responses, which are positive in general. Gratitude for the comments and consultations are expressed to A. Basilevsky, O. Kuskov, V. Shashkina, E. Zabalueva, (Vernadsky Institute, Moscow), V. Trubitsyn, A. Dolitsky (both – Shmidt Institute of Physics of the Earth, Moscow), K. Shingareva (Moscow University of Geodesy, Aerial Survey and Cartography), S. Pugacheva and the Department of the Moon and Planets of Sternberg State Astronomical Institute (Moscow), Yu. Shkuratov, D. Stankevich (both – Kharkov University, Ukraine), G. Makarenko (Physical Institute, Moscow), M. Rosiek, J. Blue, A. Wasserman (US Geologic Survey, Flagstaff, Arizona, USA), G. Fink (USDA Forest Service, Flagstaff, Arizona, USA), A. Markin (Freelance popular science author on space research, Moscow), E. Tishkovsky (Institute of Experimental Physics, Novosibirsk), Laboratory of Meteoritics of Vernadsky Institute (Moscow), A. Gurshtein (Mesa Colledge, Grand Junction, Colorado, USA), L. Ksanfomaliti (Institute of Space Research, Moscow), I. Chernaya (Planetarium of Moscow Palace of Children’s Creative Work).
**Major Episode of the Hydrologic and Volcanic History of Hesperia Planum, Mars.**

**Introduction:** Hesperia Planum (HP), which is a high-standing volcanic plateau of Hesperian age (1300×1700 km, area ~1.5 10^6 km²), is in NE part of the Hellas basin rim. The region of HP and surrounding uplands hosts a rich array of volcanic and fluvial landforms suggesting that the interaction of volcanic and fluvial processes is the main theme of both the evolution of HP and probably the history of deposition in the Hellas basin. Here we outline the most important features in the region of HP and try to correlate temporally processes that have led to their formation using the whole set of imagery and topographic data available to date (Viking, MOC, THEMIS, HRSC, and MOLA-1/64-gridded topography).

**Topography of Hesperia Planum:** The surface of HP forms a broad and shallow depression. The flat surface of it has about the same elevation (~1.2 km above MPR, except for three areas: 1) Tyrrhena Patera, which is ~3.5 km above the surface of HP, 2) area in the SE part of HP that represents a basin between ~35°~40°S and 225°~240°W (“Morpheos basin”), which is ~750~800 m deeper than the rest of the surface of HP, 3) region in the SW corner of HP, which is a depression ~200 km wide (“SW trough”) running toward the floor of Hellas.

The mean of the measured differences in elevation between the surface of HP and the adjacent uplands is ~450 m for the major portion of the HP boundary. Within the SW trough, however, the surface of HP is up to ~3 km lower than the surface of the uplands and the trough represents a "bottle neck" that breaches the uplands and connects HP with the Hellas basin. Dao, Niger, and Harmakhis outflow channels are concentrated in the trough.

**Volcanic plains, impact craters, and volume of HP:** The vast Hesperian plains make up the surface of HP [1-4]. The characteristic features of the surface of HP are wrinkle ridges that typically form polygonal networks of structures. The ridges are generally linear but in places they form unusual circular patterns. We interpret these circular ridges as structures formed by the deformation of lava plains against rims of impact craters. This interpretation is supported by the observation of the true flooded rims of craters predating emplacement of the lava plains in HP.

The initial height of the rim of the flooded craters is the measure of the thickness of the lava fill and the MOLA data allow precise determination of the shape of impact craters on Mars [5]. We have conducted a regional survey of the flooded craters in HP and found 43 such features (from 6.5 to 63 km). The mean rim height is estimated to be ~325±73 m (±1σ); the maximum value for the height is found to be ~495 m. These values give the total volume of the lava fill within HP in the range of ~0.4 to ~0.7 10^6 km³.

The other very important aspect of the flooded craters is that they characterize the morphology of the floor of HP prior the lava filling. We have compared the size frequency distribution (SFD) of the flooded craters in HP with SFD of craters in a typical Noachian terrain (Terra Tyrrhena) and in classical volcanic provinces of the Hesperian age, Syrtis Major and Lunae Planum. We also have tested a hypothesis if the combined population of the flooded and the exposed craters in HP would make the SFD to be more similar to that of the cratered uplands.

The Terra Tyrrhena curve shows the highest crater density and the Hesperian curves are practically identical and lie significantly lower. The curve for the exposed craters in HP corresponds well to the SFD of Syrtis Major and Lunae Planum. The SFD of the flooded craters in HP almost exactly mimics the distribution of the craters on the surface of both Hesperian regions meaning that the SFD of the flooded craters in HP belongs to the family of the Hesperian distributions. When the exposed and flooded craters in HP are combined, it provides a negligible shift toward the higher crater density, which is not significantly different (±10) from the SFD of the other Hesperian units.

The crater statistics strongly suggest that the Noachian population of impact craters in HP was erased before emplacement of the plains. The large-scale topographic depression of HP, thus, may have partly to wholly been formed in the post-Noachian time. If by that time the region of HP was not a depression, the maximum depth of the topographic basin that later had been formed in this area can be estimated as the sum of the mean topographic difference between the surface of HP and surrounding uplands (~500 m) and the thickness of the plains (~250 to ~500 m). This gives the range of the depths from ~750 to ~1000 m and, correspondingly, the maximum value of the total volume of material missed in the HP is from ~1.1 to 1.5 10^6 km³.

If a depression in the area of HP existed during Noachian then the minimum value of the thickness of material removed from this area can be estimated by the rim height of the larger impact craters characterizing the surface of the Noachian terrain. This height is ~300 m for the craters in wide range of diameters from 100 up to 1000 km [5]. For this value, the total volume of the materials removed from the floor of Hesperia Planum is ~0.45 10^6 km³.

**Fluvial features in HP:**

1) **Small valley networks:** The Noachian units around HP are among terrains that are most dissected by small valleys [6, 7]. The local to regional topographic gradients govern the orientation of the small valleys and the area of HP appears to be the principal sink for the valley effluents. The valleys are abruptly terminated by the contact between the uplands and HP. The absence of the deltas, fan-shaped deposits, and the channels cutting the plains means that the formation of the valleys took place before emplacement of the lava plains.

2) **Large outflow channels:** Three large outflow channels, Dao, Niger, and Harmakhis Valles, cut the surface of HP in the SW trough. The fourth channel, Reull Vallis, runs from Morpheos basin across the northern edge of Promethei Terra. The fifth, unnamed, channel is in the SE part of HP. This relatively short channel appears at ~32°S, 246.5°W, and runs southward disappearing at the northern edge of Morpheos basin at ~35°S, 246°W.

Dao and Harmakhis start in distinct closed depressions. The source regions of Niger Vallis and the unnamed channel are less distinct and marked by circular and elongated depressions suggesting both the subsidence of

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**Reference:** Microsymposium 40, abstract 37, 2004 (letter format)
the surface and subsurface flows [8]. Reull Vallis begins full-size at the western edge of Morpheus basin and has no distinct source region. Formation of all these channels clearly postdate emplacement of the volcanic material in HP.

3) Viscous flows: The viscous flows are abundant in the southern portion of the region under study. The most spectacular flows are lobate debris aprons around massifs of the uplands in the northern part of Terra Promethei. The aprons are absent both around the upland massifs within HP and in the uplands north of about 38°S and east of about 250°W [9]. The viscous flows are common features in the Dao-Niger system and along the lower reaches of Reull Vallis. The position of the flow sources relative to the surface defines two types of them: sub-surface and on-surface flows. Both type postdate formation of the channels.

The subsurface flows occur on the walls and at the heads of the large outflow channels within the SW trough. A range of features accompanying these flows (subsidence and breakup of the surface of the lava plains, pit chains, shallow trough and zones of graben marking their edges, arcuate concentric scarps concave toward the channels) indicates that the flows were originated from beneath of the composite layer of the HP lava fill. The viscous flows of the other type occur almost exclusively within the northern portion of Promethei Terra near middle and low stretches of Reull Vallis. The flows are superposed on the surface of surrounding plains and partly fill channel of Reull. The sources of these flows are on the surface and there is no evidence for the subsurface sources.

Discussion: One of the first recognizable episodes in the history of the HP area is the formation of small valley networks that dissect the surrounding uplands. The lava plains of HP clearly embay the valleys implying that they most likely continued to the original floor of HP and stored there their effluents. Although the source of the valleys is unknown, the very attractive hypothesis of their formation is the base melting of thick ice sheets [10]. If this was the case, a large amount of ice accumulated around HP and possibly within it established the source for the later fluvial activity in the region of HP.

The SFD of flooded craters in HP strongly suggests that its surface has the Hesperian size-frequency distribution of craters before the lava fill. Thus, the area of HP probably should undergo an episode of massive removal of materials that erased the older crater record. The hypothesis of magmatic erosion of the volatile-saturated regolith at the initial stage of volcanism in HP [11] offers a good explanation for such an event. The total volume of material removed from HP before the main episode of the on-surface volcanism is estimated to be from ~0.45 to ~1.5 x 10⁶ km³. All these materials probably went into the neighbor Hellas basin where they may form a layer about 0.5-1.5 km thick.

The second episode of water release postdated emplacement of the lava plains and the centralized volcanoes such as Hadriaca Patera and led to formation of the outflow channels. For the Dao-Niger system, Harmakhis Vallis, and the unnamed channel in SE portion of HP there is the good evidence that the sources of the flows were beneath the layer of the lava plains. The volume of material removed from these channels is ~0.02 x 10⁶ km³ [12] or only ~ 1.5-5.5% of the total volume of material possibly eroded from HP. Reull Vallis is different from other channels because its source apparently was on the surface of the plains within the depression of the Morpheus basin. The possible scenario of evolution of Reull Vallis and area around it is presented in the separate abstract [9].

At the apparently last stage of the fluvial activity, the viscous flows played the most important role. The flows are concentrated almost exclusively in the areas cut by the largest outflow channels (Dao-Niger, Harmakhis, and Reull Valles) and superposed on their floor and partly fill the channels but the total volume of the flows is small comparing with the amount of material eroded away from the channels. The most important feature of the flows is that they have distinctly different source regions. The flows that occur at the Dao-Niger system are originated from the subsurface (on-surface flows are absent) and the flows around Reull begin from the sources that were on the surface (subsurface flows are absent).

The different position of the flow source regions suggests the different explanation of their formation. The subsurface flows are most likely related to the remainders of volatiles in the reservoir that was almost emptied during the massive erosion in HP and formation of the outflow channels. The on-surface flows may have been formed due to formation of a transient water reservoir within the Morpheus basin that was filled from the subsurface source by the unnamed channel. Reull Vallis then drained the basin and its effluents were re-accumulated in the east part of the Hellas rim where the on-surface flows now prevail [9].

Conclusions: The hydrologic history of HP appears to begin with the accumulation of volatiles around and in the HP basin and formation of a large reservoir there in the late Noachian. The reservoir was then emptied in three different modes that reflect diminishing of the amount of the stored volatiles: 1) the massive areal erosion, 2) the outflows concentrated in a few places, and 3) dispersed viscous flows. Volcanism within HP probably played the major role in mobilization and release of the volatiles. It is appears to be likely that the volcanic activity had induced the main episode of erosion in HP [11] and it is also possibly that later magmatism was triggering the outflow channels [8,13]. Formation of the viscous flows probably is not related to volcanic activity and represents flows from the largely depleted initial reservoir of volatiles within HP.

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ARE MCCs DHOFAR 225 AND DHOFAR 735 OF CM3-TYPE? M.A. Ivanova¹, L.V. Moroz², M. Schmidt³, U. Schade⁴, F. Brandstätter⁵, M.A. Nazarov¹ and G. Kurat⁶. ¹Vernadsky Institute of Geochemistry and Analytical Chemistry, Kosygin St. 19, Moscow 119991, Russia (venus2@online.ru). ²German Aerospace Center, D-12489 Berlin, Germany. ³Heidelberg University, D-69120 Heidelberg, Germany. ⁴BESSY GmbH, D-12489 Berlin, Germany. ⁵Natural History Museum, A-1014, Vienna, Austria.

**Introduction:** Dhofar 225 and Dhofar 735 are metamorphosed carbonaceous chondrites (MCC) with some similarities to the Antarctic MCCs Belgica-7904 (CM) and Yamato-86720 (CM) [1]. Based on our previous data and new results obtained using in situ synchrotron IR microspectroscopy (SIRM) we discuss the possible genesis of Dhofar 225 and Dhofar 735 by dehydration of matrix phyllosilicates. In addition, we studied a new Ca,Fe-oxysulfide [2].

**Results:** In texture and petrography, Dhofar 225 and Dhofar 735 are similar to CM chondrites [1]. However, Dhofar 225 contains the first Ca,Fe-oxysulfide found in nature [2]. Its best-fit stoichiometry and low analytical total indicate a formula of \((\text{Ca}_{4.66} \text{Fe}^{2+}_{0.34})_5\text{Fe}^{3+}_{6}\text{S}_5\text{O}_9\). Another possible formula is \(\text{Ca}_4\text{Fe}^{2+}_5\text{S}_4(\text{OH})_4\text{O}_3\), but the Ca,Fe-oxysulfide inclusions appear to lack OH because they are stable under the electron beam. Moreover, absorption bands of structural OH at 2.7 μm were not detected in these grains by SIRM.

Matrices of Dhofar 225 and Dhofar 735 are very fine-grained, similar to the MCC’s, they have high EPMA totals, are depleted in Fe and S, and contain small grains of olivine, troilite, taenite, and tetrataenite [1,2]. The bulk composition of Dhofar 225 is low in H2O (1.76 wt.%) and Fe.

No signatures of O-H bonds (in structural OH and/or bound H2O) at 2.7-3 μm were detected in the Dhofar 225 and Dhofar 735 matrices by SIRM, suggesting a lower content of hydrated phases, phyllosilicates and tochilinite, as compared to those in CMs. The O-H absorption bands were identified by SIRMs in the matrix spectra of CMs Cold Bokkeveld, Murray and Mighei, and in tochilinite inclusions of Murray, studied for comparison. Further evidence for the dehydrated state of the Dhofar 225 and Dhofar 735 matrices is the position and shape of strong bands around 10 μm due to Si-O vibrations, being consistent with fine-grained Fe-rich olivine. The positions and shapes of the Si-O bands in the IR spectra of the typical CM2 matrices are different, being consistent with mixtures of Fe-rich and Mg-rich phyllosilicates.

**Discussion:** Dhofar 225 and Dhofar 735, the first non-Antarctic MCC meteorites, expand the MCC group and have similar oxygen isotopic compositions (Fig. 1) [2]. They apparently have experienced heating after aqueous alteration. No water-bearing mineral was detected in their matrices, indicating that phyllosilicates were dehydrated. The materials were heated above 245 °C, since tochilinite disintegrates into troilite and oxides at 245 °C. The absence of this phase distinguishes MCCs from CMs. The presence of dolomite in Dhofar 225 and Dhofar 735 indicates an upper limit of the temperatures reached of ~700 °C. Although all MCCs underwent aqueous alteration followed by heating to different degrees [3], they may be considered as CM3 chondrites according to the general characteristics of the petrological type 3 chondrites. However, the meteorites differ from CMs also in the bulk abundances of some refractory (enriched in Ti and Al), siderophile (depleted in Ni and Fe), and moderately volatile elements (enriched in P and K). It is unlikely that these differences in bulk chemistry are the result of metamorphism. Therefore Dhofar 225 and Dhofar 735, as well as MCCs, may represent a separate group of carbonaceous chondrites of type 3 (also supported by oxygen isotopes [2]) Fig. 1.

THE METAMORPHIC HISTORY OF CO AND CV CHONDRITES INVESTIGATED BY THE THERMOLUMINESCENCE METHOD

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INTRODUCTION

The CO chondrites are similar to type 3 ordinary chondrites in several respects [1]. They are both chondritic in bulk composition, with non-volatile elemental abundances generally within about 30% of the CI values. Thus the two groups are mineralogically very similar, consisting of olivine, pyroxene, plagioclase, metal and sulfide. Like the ordinary chondrites the CO chondrites appear to constitute a metamorphic sequence [2-4]. However, they also differ from ordinary chondrites in several respects. They are isotopically different [5, 6], element ratios show small but significant differences [7, 8], they contain refractory amoeboid inclusions, and their chondrules are smaller [2, 9]. Unlike type 3 ordinary chondrites, CO chondrites often contain primary calcic feldspar [10], presumably associated with the refractory inclusions. Keck and Sears [3] also found that the thermoluminescence (TL) sensitivity of the (110-120)°C peak increased by a factor of 100 with increasing metamorphism, while the TL sensitivity of a second TL peak at 230°C was not metamorphism-dependent. They suggested that the first peak was caused by feldspar formed by devitrification of chondrule glass, a situation analogous to that of type 3 ordinary chondrites [11, 12], while the 230°C peak was due to primary (i.e. non-metamorphic) feldspar, perhaps associated with refractory inclusions. Compositional equilibration between refractory inclusions and the ferromagnesian components, and variations in the homogenization of matrix olivine, suggests that the CV chondrites have suffered various levels of parent-body metamorphism [13-15]. Since the CV chondrites consist of both oxidized and reduced subgroups, a single metamorphic series is precluded although two parallel series are possible [13]. The petrographic, mineralogical and bulk compositional differences among the CV chondrites indicate that the TL sensitivity of the ~ (110-130) °C TL peak is reflecting the abundance of ordered feldspar, especially in chondrule mesostasis, which in turn reflects parent-body metamorphism [16].

The purpose of the present paper was to study of CO and CV chondrite metamorphism using the TL-device of the Vernadsky Institute and the scaling procedure proposed by [1, 16].

EXPERIMENTAL

The measurements of TL induced by X-rays were carried out for 21 carbonaceous chondrites. Nine CO3, eight CV3 and four CK chondrites were studied. Samples weighing from 0.7 up to 1.0 g were crushed and powdered in a jasper mortar. Then a magnetic fraction was removed from the powders using a manual magnet. Three 2 mg aliquots of each non-magnetic fraction were measured using the TL device. After measurements of natural TL (the heating up to 500 °C), the samples were irradiated with X-rays for two minutes and then induced TL was measured. The experimental procedures have been described in more detail earlier [17-19].

RESULTS AND CONCLUSIONS

The results of induced TL measurements are given in Table, where I_{TL} is the TL peak height at the temperature of about 130 °C. The values of I_{TL} were normalized to I_{TL} of the Dhajala chondrite (I_{TL} Dhahala = 1). The subtypes determined by this study and others [1, 16] are shown also in the table. The star symbol (*) marks the recommended petrographic type. The glow curves of TL of CO, CV and CK meteorites of different types - are shown on Fig. 1. The majority of investigated chondrites have a composite shape of glow curves with peaks in the temperature range of 110-130 °C. There are also some peaks at > 150 °C. However the Coolidge is different from others. It shows only two peaks at ~ 130 °C and ~ 150 °C (Fig. 2). Such shape of TL peaks is most typical for glow curves of ordinary chondrites. The chondritic subtypes obtained in this study and determined by [1, 16] for the same meteorites coincide very well (Table, Fig. 3) and, therefore, the method applied in our laboratory is suitable for determination of the metamorphic grade of carbonaceous chondrites.

Here we report first the subtype others Acfer 202, Dar Al Gani 078, and Dar Al Gani 303 CO chondrites, the SaU 085 CV chondrite, and Dhofar 535 ungrouped chondrite (Table). The received subtypes do not conflict with results of petrographic and other investigations. In addition, subtypes of Dhofar 015 (CK3), Ningqiang (CK – ungr), Karoonda –(CK4), and Maralinga –(CK4) were first determined. The obtained results indicate, that CK chondrites are unique among metamorphosed chondrites. They demonstrate no detectable induced TL. It confirms mineralogical data on very unusual feldspar occurring in these meteorites.

REFERENCES

Table. Observed results of a peak height (ITL) of glow curves (130 OC) and degree of metamorphism carbonaceous chondrites.

<table>
<thead>
<tr>
<th>Meteorite</th>
<th>Type</th>
<th>$I_{TL}$ (ITL Drajala=1)</th>
<th>Subtype</th>
</tr>
</thead>
<tbody>
<tr>
<td>This paper</td>
<td>Others</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Felix CO</td>
<td>0.106</td>
<td>3.4</td>
<td>3.2-3.5 (3.4*) [1]</td>
</tr>
<tr>
<td>Isna CO</td>
<td>0.356</td>
<td>3.6</td>
<td>3.6-3.8 (3.7*) [1]</td>
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<td>Kainsaz CO</td>
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<td>3.1-3.5 (3.2*) [1]</td>
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<td>3.4-3.7 (3.4*) [1]</td>
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<td>3.3-3.6 (3.4*) [1]</td>
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<td>3.5-3.8 (3.6*) [1]</td>
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<td>Allende CV</td>
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</tr>
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<td>3.0</td>
<td>3.0-3.3 (3.0*) [16]</td>
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<td>Coolidge CV</td>
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<td>3.7</td>
<td>3.8-3.8 (3.8*) [16]</td>
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<td>3.3</td>
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<td>3.0-3.3 (3.3*) [15]</td>
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<td>Leoville CV</td>
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<td>3.0-3.6 (3.0*) [16]</td>
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<td>3.5 [20]</td>
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<td>DAG 303 CO</td>
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<td>DAG 078 CO</td>
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<tr>
<td>Dhofar 535 Ungr.</td>
<td>0.030</td>
<td>3.2</td>
<td>—</td>
</tr>
<tr>
<td>SaU 085 CV</td>
<td>0.171</td>
<td>3.5</td>
<td>—</td>
</tr>
<tr>
<td>Dhofar 015 CK</td>
<td>0.212</td>
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<td>—</td>
</tr>
<tr>
<td>Karoonda CK</td>
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<td>—</td>
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<td>Maralinga CK</td>
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</tr>
<tr>
<td>Ningqiang CK</td>
<td>0.029</td>
<td>3.1</td>
<td>—</td>
</tr>
</tbody>
</table>

(*) - Recommended petrographic type.
Vernadsky Institute of Geochemistry and Analytical Chemistry Russian Academy of Sciences, 119991, Moscow, Kosygin Str. 19. cosmo@geokhi.ru. tel.: (095) 137-86-14.

Introduction: In order to test cosmic radiation condition and shock-thermal history of the Brenham pallasite, three other meteorites on this class were selected for study. Olivine crystals from the Eagle Station, Marjalahti and Omolon were examined by the track and thermoluminescence (TL) methods. The Brenham pallasite is unusual in the two main aspects: (1) Although this meteorite nominally classified as a pallasite, it has a solar type gases rich fraction [1] that was not observed in any other meteorites of this class. (2) Olivine crystal microstructure of the Brenham, probably, reflects the process of brecciation.

Track measuring: In this first-step study we measured tracks in one sample (~0.1 g) of the Brenham pallasite. The tracks might be expected in the meteorite crystals from some sources: U238 and extinct Pu244 spontaneous fission fragments, and galactic (ρGCR) and solar (ρSCR) cosmic-ray very heavy (VH-group) nuclei. The possibility of presence in meteorite matter of the last track-source group is correlated with process of their precompaction solar wind irradiation. The chemically etched tracks, the number of which per crystal was small, 0 to 7, and gave track densities (ρ) ranging for 53 of studied crystals (sizes of ~50-300 μm) in the interval from 10^3 cm^-2 up to 1.5x10^5 cm^-2. Obtained results are shown in Fig. 1 and Table 1. As it seen from the histograms in all three size-group of olivine crystals, the very wide track-density range (up to three order of magnitude) are observe. The main part of grains (totally ~80%) have ρ =10^3 cm^-2, that corresponds to galactic cosmic ray VH-nuclei. A higher ρ-values can be attributed only to solar cosmic ray VH-nuclei, the total irradiation dose of which in individual olivine grains is not the same. Note the sufficient, near twice, increasing of the mean ρav for grains of the smallest size-fraction.

On the base of obtained track data it can be assumed, that solar type gases in Brenham material could be due to solar wind ions, implanted during of pre-compaction stage of this meteorite parent-body formation or/and in later regolith stage. However, the last scenario for pallasites is unreal. sizes: mean size- fractions of ≥ 200 μm, 100 – 200 μm and ≤ 100 μm, respectively.

TL measuring: TL has been studied in the vicinity four pallasites. Bulk powder olivine samples by weight of ~2 mg were annualized. For the measurement of artificially induced by 55 KeV X-rays TL it was used TL equipment described earlier [2]. Figure 2 shows the typical TL glow-curve shapes. As it seen in all cases the high temperature (T ≥ 250 °C) peak occurs, whereas the very high region (T<370 °C) is essentially depleted in TL glow. Note the significant difference of the glow-curve shapes between Omolon and Murchison meteorites. The first demonstrate presence of high-density point dislocations in the individual olivine grains, that is not observed in second as taken for comparison. TL glow-curve shape for Brenham also differ from other pallasites. The total difference were observed in comparison with TL glow-curves, measured in artificially shocked olivine (our unpublished data). In the latter case (see Fig. 3) the low temperature wide peaks (near 100-200 °C) is predominant. Now we can only to note that these preliminary observations indicate artificially induced TL, chiefly characterized the microstructure of the crystals under investigation, are: (a) high variable in different pallasites, and (b) the level of possible shock influence, recorded in pallasites olivine is very low.

Conclusions: (1) The shape of the track-density distribution, observed in the Brenham olivine crystals, supports assumption, that this distribution resembles determined irradiated on pre-compaction stage crystals. These tracks probably represent an addition of the solar cosmic ray VH-nuclei. (2) The existence of difference in the track- and TL-parameters in olivines for different pallasites tells us that a variety of processes were occurred in the pre-compaction history of these meteorites.

Partially the work is performed under supporting of RFPI Grant 04-05-64930.

Fig. 2. The TL induced by X-ray irradiation in olivine from pallasites Brenham, Engle Station, Omolon and Marjalahti.

Table 1. Track parameters in the Brenham olivine grains.

<table>
<thead>
<tr>
<th>Crystal fraction</th>
<th>N grains</th>
<th>N tracks</th>
<th>S, 10^2 cm2</th>
<th>( \rho_{\text{ee}}01,3 ) cm2</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>10</td>
<td>57</td>
<td>1.4</td>
<td>4.1 ± 0.5</td>
</tr>
<tr>
<td>B</td>
<td>37</td>
<td>56</td>
<td>1.5</td>
<td>3.7 ± 0.5</td>
</tr>
<tr>
<td>C</td>
<td>6</td>
<td>8</td>
<td>0.12</td>
<td>7.0 ± 2.5</td>
</tr>
</tbody>
</table>

Fig. 3. The induced by X-ray TL glow curve in the experimentally shock loaded olivine sample under pressure about 20-30 GPa.

Summary: We analyzed polarimetric data for the Martian disk obtained by the Hubble Space Telescope (HST) at the 2003 opposition of Mars. We found transient polarimetric features and studied their relation with surface and atmosphere. We found optically thin clouds that are strongly polarizing in ultraviolet light. The clouds likely are related to condensing ice in the atmosphere, since such features can be formed by micron-scale particles of regular shape and similar size.

Introduction: Planetary surfaces and atmospheres affect the polarization state of scattered solar light. The degree of linear polarization P depends on wavelength λ and illumination/observation geometry (phase angle α). Size of scatterers, their composition, shape, and orientation are responsible for polarization degree variations. The Martian surface is composed of scatterers (soil particles) of different sizes and aggregates of small grains. The atmosphere contains molecules and small particles (permanent sub-micron dust haze) of the clean atmosphere, mists and clouds consisting of ice crystals, and comparatively large suspended particles from dust storms. Earlier Earth-based polarimetric observations of Mars in visible light are presented in [1].

HST polarimetric data: The HST observation program #9738 [2] was carried out at the time of closest Earth-Mars encounter as Mars passed within 0.372 AU of Earth. Five series of images of Mars were taken on Aug. 24, just before the closest approach and on Sept. 5, 7, 12, and 15. The phase angles α at the observation moments were 6.4, 8.2, 9.7, 13.6, and 15.9°, respectively. The image scale was about 7 km/pixel at the disk center (the highest spatial resolution ever achieved from Earth). Disk center in all images was located at 19°S 20-35°W showing the distinctive features of Valles Marineris, Terra Meridiani and surroundings. The season on Mars was summer in the southern hemisphere (the acentric longitude of the Sun Ls = 247 – 261°).

Within this observation program, for the first time, HST observations of Mars included imaging polarimetry. The observations were done with the Advanced Camera for Surveys (ASC) with 3 polarization filters [3]. We used polarization data obtained in combination with two ultraviolet wide-band spectral filters F250W and F330W, and a wide-band blue filter F435W. The number in the filter name is approximately the effective wavelength in nm.

Data processing: The standard dark current, flat field, and geometric distortion calibrations were performed routinely by the HST data retrieval facility. We then identified and removed the cosmic-ray tracks from the images using an original heuristic algorithm for cosmic rays identification [4]. After spatially coregistering the images obtained with the three polarizers, we derived Stokes parameters of scattered light. To perform the latter step, we derived polarimetric calibration information from calibrating observations of a standard star (HST programs 9586, 9661, 10055) and observations of Mars themselves, making use of symmetry requirements for the center of the disk (see [5] for details).

Transit polarization clouds: Using the nor- malized Stokes parameters we made images of P for the whole Martian disk for all observation dates. These images reveal transient high-polarization features. These features vary in size, polarization degree, and location. They were best expressed on Sept. 5 and 7, and disappeared after that. The detailed view of polarization degree distribution in the southwestern part of the planetary disk on Sept. 7 is presented in Fig. 1. The most prominent polarization features (red-color-coded) have polarization as high as 3%, the polarization plain is oriented close to the scattering plane. Visual inspection confirms that these features coincide with faint semitransparent clouds.

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faint semi-transparent clouds show the polarization excess. To illustrate this conclusion we compared albedo and polarization profiles over the most interesting region characterized by dramatic changes in polarization degree and cloud density. Fig. 2 shows an image in the UV-330-nm filter for the same part of the disk as in Fig. 1. A complicated clouds system (light shades) with gradual brightness changes and rather dark surface details visible through clear atmosphere are seen.

Fig. 2. Albedo image of the southwestern part of the Martian disk in the UV filter F330W, $a=10^\circ$. Red outlined region is discussed in text.

Fig. 3. Distribution of polarization degree for the area shown in Fig. 2. UV filter F330W.

The distribution of $P$ for the same filter is presented in Fig. 3. To suppress inevitable noise and accentuate major polarization changes we slightly smooth the $P$ distribution with a Gaussian filter. The red-outlined rectangle in Figs. 2, 3 cover the area from the northern part of the Argyre Basin to the highest Thaumasia Planum and include the major portion of albedo and polarization dynamic range in the images. Fig. 4 shows the albedo-polarization correlation diagram for the outlined region. It is seen that both the transparent atmosphere (dark areas) and bright clouds show typically low polarization values, while the faint clouds have abnormally high polarization degree. Note, however, that in other portions of the disk, there are intermediate-brightness semi-transparent clouds without polarization anomalies.

Fig. 4. Correlation diagram albedo-polarization for the outlined region in Figs. 2 and 3.

**Atmospheric properties from TES:** We retrieved Mars Global Surveyor Thermal Emission Spectrometer (TES) atmospheric dust opacity (9.7 $\mu m$) and ice opacity (12.1 $\mu m$) data for the dates. Unfortunately, no MGS tracks crossed the areas of high polarization. The data show that the atmosphere is rather dusty (which is typical for this season), but no particular large-scale dust-lifting events occur in the imaged part of Mars. TES data show a slightly higher ice opacity in the entire western portion of the disk where the polarimetric clouds are observed.

**Discussion:** Low polarization of bright nontransparent clouds is caused by well developed multiple scattering between aerosol particles in these clouds. This, in fact, explains their high brightness. Multiple scattering in the optically thin clouds is insignificant, and polarization produced by single scattering is preserved. Scattering in a clear atmosphere is dominated by molecular scattering, which follows the Rayleigh scattering law. Therefore, the polarization is normal to the scattering plane and quadratically approaches zero when the phase angle tends toward zero. Thus, a clear atmosphere produces low $P$ in our case.

High polarization of the optically thin clouds suggests that they are made of scatterers having regular shape and similar size. The drop-off in the absolute polarization that occurs when the observation wavelength increases, suggests an upper particle-size limit in the submicron region. Such aerosols may be the result of the initial stages of nucleation of H$_2$O ice crystals on submicron dust.

TECTONICAL AND CHEMICAL DICHOTOMY OF MARS AND EARTH: COMMON AND DISTINCTIVE FEATURES. G. G. Kochemasov, IGEM RAS, 35 Staromonetny, Moscow 119017, Russia,
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As planetary bodies both Mars and Earth share common features developed in all celestial bodies [1-3 & others] just because they all move in non-round (elliptical, parabolic) keplerian orbits and rotate. A fundamental statement: “Orbits make structures” could be unfolded in form of 4 theorems: 1. Celestial bodies are dichotomic; 2. Celestial bodies are sectorial; 3. Celestial bodies are granular; 4. Angular momenta of different level blocks tend to be equal [2]. All these tectonic features are a result of standing waves warping spheres of rotating celestial bodies in 4 directions. The fundamental wave1 (long $2\pi R$, where R is a body radius) makes the ubiquitous tectonic (and, as a consequence, chemical) dichotomy. Overtones of the wave1 make smaller sectoral structures (most important wave2 warps spheres giving them features of a structural octahedron). Individual waves, inversely correlated with orbiting frequencies, make tectonic granulas. Hypsometrically (tectonically) different level blocks of a rotating body tend to keep their angular momenta compensating changing radius by changing density of composing blocks materials.

A principal difference between Earth and Mars, among others, is in their orbiting frequencies. 1/1 year for Earth makes granula size $\pi R/4$ (wave4), 1/2 years for Mars makes granula size $\pi R/2$ (wave2). 4 waves in the terrestrial great planetary circle produce 8 alternating highs and lows. 2 waves in the martian great circle produce 4 alternating highs and lows. 4 waves warp the terrestrial sphere more or less evenly, 2 waves warp the martian sphere inevitably in such a manner that it is extended in one direction and squeezed in the perpendicular one (Fig.4). That is why Earth could be compared with a watermelon and Mars with a melon. This wave shaping immediately tells on the planetary relief range. On Earth it is about 20 km, on Mars ≈30 km (or 56 km brought to the Earth’s diameter). The dichotomy relief (an average between the northern lowlands and southern highlands) is about 6 km. The Earth’s dichotomy relief is not so sharp.

The fourth theorem is satisfied at Earth by an average density range between lowlands (the western Pacific hemisphere) and highlands (the eastern continental hemisphere) about 0.25 g/cm$^3$, that is the difference between densities of oceanic tholeiites and continental andesites (an average composition of continents). No doubt, at Mars this density range must be higher, say, 0.45 g/cm$^3$ [4]. Martian lowlands composed of Fe-basalts are denser than the Earth’s oceanic lowlands (tholeiites). Consistently, the martian highlands must be less dense than the Earth’s continents, this means less dense than andesites. In [4, 5, 6] before the “Pathfinder” mission we have proposed contrasting rocks at the martian lowland/highland contact, mentioning albitites, syenites, granites as highland lithologies. “Pathfinder” found at the contact andesites. We interpreted them as contact rocks (compare to the circumpacific andesite belt at Earth) and insisted at lighter (less dense) rocks inside the highlands.

Albedo and gravity (MGS) data and gamma-spectrometry (Odyssey) do not contradict to these not dense light alkaline and acid rocks. But in situ study of highland rocks is possible only now. “Opportunity” on Meridiani Terra and “Spirit” at Gusev crater indeed discovered light layered rocks. But rinds of salts (sulphates, chlorides, bromides) and eolian Fe-rich sediments (hematite) cover outcrops and separate blocks causing difficulties to analysers trying reach fresh rocks. It seems that drilling to depths about 5-10 mm is not always enough to get desirable rocks. Nevertheless, partially published in Internet results indicate at silicate rocks rich in Al (high Mg and Fe can be partially explained by surface rinds), containing Na, K, Ca. Ti. Al/Ca steadily increases from lowland basalts (“Viking” data) to highland rocks. Mg/Fe does the same. On the whole, one might think about sharp transition of Fe-rich lowlands to Al, Mg, alkali-rich highlands. Some flood-basalts can be found on highlands but these basalts are Mg-rich in contrast to Fe-rich dense basalts of lowlands (The same situation but not so sharp with flood-basalts at Earth: Mg/Fe in them increases from oceans to continents, that means their density diminishes in this direction).

In [7, 8] we paid attention to high chlorine in martian rocks and soils (0.3-0.6%) attributing this to alkaline rocks (at Earth, normally, these rocks are the highest in Cl among magmatic rocks). Accumulation of Cl means the utmost magmatic fractionation. Namely, this kind of fractionation is expected wide-spread at Mars. At Earth there are only a few examples of such advanced fractionation. The best example is probably the Lovozero alkaline ring complex at Kola peninsula. Perfectly thinly layered ring complex of nepheline syenites occupies an area of 650 km$^2$. Alkaline pyroxene-rich rocks alternate with nepheline-rich ones. There are many Ti, Nb, Zr –minerals. In some rocks sodalite prevail over nepheline and Cl content in rocks can reach upto 2.5% (tavite). On an average the Lovozero massif contains 0.2% Cl.

Feldspathoids (nepheline, sodalite) are often replaced by zeolites (hydrothermal alteration and
weathering). These water-containing minerals are very wide-spread at Earth and probably at Mars. In [9] we proposed that zeolites can be water-sinks and suggested that the martian near-equatorial anomalous hydrogen can be related to zeolites (water-containing salts are already discovered by two landers).

So, “dull” entirely basaltic Mars does not exist any more. But only a few years ago the majority of planetologists believed in it! The comparative wave planetology from the beginning [4] insisted at highly fractionated Mars. It means that the regular wave planetology having the predictive power really exists and adequately reflects natural processes.

Fig.1 shows similar wave produced tectonic dichotomies of Mars (N-S) and Earth (E-W); similarities are in shape of subsiding segments and their relative areas (~1/3 of the globes surfaces). Fig.2 shows dichotomy of Mars and the wave produced antipodality of remarkable martian tectonic features: “lows” on highlands and “highs” on lowlands being parts of tectonic sectors. Fig.3 shows how tectonic sectors look in nature. Fig.4 explains successes and failures of martian landers: the overall score is fifty/fifty.


Fig.1 Similar formation of Mars’ and Earth’s tectonic dichotomy: a model of wave interference. A – Vastitas Borealis of Mars. Crustal thickness inside the contour is less than 50 km [10] (as viewed from inside the globe what makes the contour mirrored). B – Pacific basin. C – Flat geometric model of wave interference (4 wave directions). One needs mentally to wrap up it around the globe. Fig.2 Martian hemispheres with the dichotomy boundary. Antipodality of Hellas (1) to Alba patera and Tempe terra (2) and Argyle (3) to Elysium planum with Phlegra montes (4). Fig.3 Mars’ sectoral structure. Hubble Space Telescope image on 26.8.2003, eleven hours before the closest approach of Mars to Earth (Image courtesy of NASA, J. Bell & M. Wolff. ESA Bulletin # 115, Aug. 2003, cover). Fig.4 A scheme of successful (Spirit, Opportunity) and failed (Beagle2) landings on the martian surface through heterogeneous atmosphere produced by wave processes (wave2 structure).
The new wave planetology [1 – 4 & others] gradually presses the habitual impact planetology by means of its predictive power. Impacts are not able to explain regular shapes and structures of celestial bodies constantly appearing in images sent by the planetary missions and to give an adequate explanation of chemical fractionation processes related to morphology of the bodies. The wave planetology firmly states: “Orbits make structures”, and proves this by numerous comparative data of cosmic experiments. The newest Cassini project already gave excellent images of the saturnian rings an the outermost satellite Phoebe. In [5] we stated that the Cassini project by its detailed studies will bring data supporting the wave planetology. Various interfering wave structures of the rings confirm this prediction. Fig.3 shows as crossing waves create evenly spaced round features of equal sizes, not impact craters! If the ring structures are wave-induced then satellites have to be affected by waves as well.

Earlier we have shown relations between waves and shape (structure, compositional regularity) in form of 4 theorems of planetary tectonics [3]: 1. Celestial bodies are dichotomic (the fundamental wave1 makes this); 2. Celestial bodies are sectoral (effects of overtones); 3. Celestial bodies are granular (effects of individual waves lengths and amplitudes of which are inversely proportional to orbiting frequencies); 4. Angular momenta of different level blocks tend to be equal (in a rotating body wave produced tectonic blocks of different planetary radii tend to keep their angular momenta by adjusting their densities). Small satellite Phoebe obeys all these requirements. It is dichotomic: one side of it is flatter than the opposite one. This can be seen in a shape model produced by the Space Science Inst., Boulder, Colo (Fig.4 ). The convexo-concave shape is typical for small celestial bodies [6]. Phoebe reveals perfect sectoring (Fig.5, 6; compare to Miranda’s sectors, Fig. 8-10) and some polyhedral outlines made by wave2 longing made a compare to Miranda’s sectors, Fig. 8-10) and some polyhedral outlines made by wave2 longing made a

\[ \frac{1}{2.65} \times 7.5 \]

18 km) is observed on its surface as a typical crater satellite with many cavities not able to compact them (say, heavy troilite) are gone away and the small Amalthea situation is worse: even sulphur compounds drastically different despite their similar sizes, rotations and possibly densities. Phoebe is an icy satellite, Amalthea has a similar density but produced by completely different “torturous” outgassing processes (a “candle-end”).

Outgassing plays very important roles in development of planetary scenarios. The mysterious asteroid belt has some unanswered questions. Why it
is “layered”; why asteroids are flat, not isometric; why larger asteroids rotate faster than smaller ones. The impact planetology has difficulties answering them. The wave planetology, having as an example the sequence of the inner planets, states that the inner asteroid belt orbiting faster than the outer parts is better outgassed. So, “leavings” are denser and acquire metallic character. In the outer parts prevail less outgassed carboniferous C-asteroids. Larger asteroids keep their near original mass and rotation velocities. Sweeping out volatiles means mass reduction (bodies become smaller), loss of angular momentum and hence slower rotations (compare Mercury and Venus, from one hand, and Mars and Earth, from another*). Universal flat bodies of asteroids – a result of squeezing them by the fundamental wave that bulges out one hemisphere and presses in the antepodean one. The convexo-concave shape is a consequence of the universal wave shaping [6].

*A new finding of this kind is, possibly, fractionation of saturnian rings revealed by “Cassini”: icy fragments of the inner parts are “dirty”, of the outer parts –purer, ice-rich. Explanation: enhanced degassing of the inner parts leaves dirtier particles and surfaces.


Fig.1 Amalthea, leading side, PIA01074. Diamond shape and intersecting wave warps. Indivisible through structure. Fig.2 Amalthea, trailing side, PIA01074. Cellular texture, evenly spaced “spots”.

Fig.3 Saturnian ring, a part of the NASA/JPL image “Cassini enters Saturn’s orbit”, soi-3, 20.07.04. Intersecting waves create ring structures. Fig.4 Phoebe’s dichotomy and bumpy topography. Outlined are low areas making tectonic granulation ~πR/3. The view centered at 90° west. Drawing after the NASA/JPL/Space Science Inst. colorful graphic “The true shape of Phoebe”, pia06070, 23.06.04. Fig.5 Phoebe sectoral structure. The NASA/JPL/Space Science Inst. image “Battered Moon”, pia06066, 12.06.04. Fig.6 Drawing stressing the sectoral structure of Phoebe (see Fig.5). Fig.7 Phoebe drawing after the NASA/JPL/Space Science Inst. image “Battered Moon”, pia06066, 12.06.04. Faintly visible intersecting grooves make cellular appearance (granulation) of the Moon’s surface. Also visible are on the whole polyhedral outlines of the satellite. Fig.8 Miranda’s sectoral structure (NASA/JPL P-29541). Subsided squeezed sector (below) shows layered formations; uplifted extended sector (up) is full of degassing craters; to the left and right are neutral sectors. Fig.9 Drawing of the Miranda’s tectonic sectors (after Fig.8). Fig.10 Scheme of the wave interference origin of the sectors in Fig.8, 9 & Fig.5.
CHARACTERS OF CHANGES THERMOPHYSICAL PARAMETERS OF THE MARTIAN
POLYGONAL TERRAINS IN DEPENDENCE FROM THEIR GEOGRAFIC POSITION.
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Introduction: We have present results of Mars Climate Database (MCD) data analyses based on
General Circulation Model (GCM). The last one was produced for Earth and later it was adopted for
Martian conditions [1]. The adoption adjust with new data of last missions which had investigated
climatic dependences on the planet. Thermal surface data of thermal flux to surface were analysed for
sites there we previously have found maximum polygonal pattern grounds concentration. Finally we
have received data of above mentioned characteristics change in dependance of latitude and longitude.
We have made average-out of these parameters on seasons and produced their model processing. As a
result we have a graphs of thermal profiles of polygonal sites with depth of zero annual amplitude
attenuations. In our opinion it allow to come to the polygonal classification on new base, which have
founded on factorial material and approved two-layers model of martian surface.

Observation: Our investigation have included analyses of thermal changes graphs in dynamic
monitoring during martian year for sites with co-ordinates (first site - 65N, 48E; second site - 65S,
48E; third site - 65S, 30E), there we had found polygonal pattern ground terrain, which were refered
by us [2] to ice-cracking process results (Fig. 1,2,3). We have produced

![Figure 1](https://example.com/figure1.png)

Figure 1. Surface temperature dependance on martian seasons for three research sites.

![Figure 2](https://example.com/figure2.png)

Figure 2. Thermal flux to surface data dependance on martian seasons for three research sites.

![Figure 3](https://example.com/figure3.png)

Figure 1. Solar radiative flux to surface data dependance on martian seasons for three research sites.

In previous papers we based in our model calculation on indirect data received through analyses of
heat inerter maps, width of cracks and so on. In present paper we have used factorial material, what let us
increase precision of thermal fields for polygonal sites. We determinated some dependences of depth
of zero annual amplitude attenuations for characters sites, which were chosed on the base of previous
polygonal classification, connected with comparision with earth analogues and on the using of veri-

analogic analyse for input thermal flux MCD data for these sites. These data have became the basement for
 calculation analyses of thermal fields of investi-
CHARACTERS OF CHANGES THERMOPHYSICAL PARAMETERS.

I.A.Komarov¹, V.S.Isaev², O. Abramenko³

fied methods of terrestrial classification (Rac, Kaplina, Romanovsky).

Conclusions. These results have allowed us to make conclusions about some zoning in thermal parameters changing in dependence with their geographic position and their influence on thermal regime formation for frost ground, composed with sites. Further research in this direction will allowed to solve the task of morphologic zoning of Marsian surface and as a result to microrelief map issue, based not only on external factors but on thermal regime and thermal parameters too. In our research we took in account the works of other scientist [4], hence we think that our approach to a problem have deeper basement for zoning tasks.

References:
MAPS OF PARAMETERS OF THE POSITIVE POLARIZATION MAXIMUM FOR THE LUNAR DISK.
V. V. Korokhin, and Yu. I. Velikodsky. Astronomical Institute of Kharkov National University. Sumskaya Ul., 35, Kharkov, 61022, Ukraine. E-mail: dslpp@astron.kharkov.ua.

Maps of parameters of the maximum of positive polarization: The surface of the Moon is an example of atmosphereless celestial bodies’ surface. Due to the facts that albedo of the Moon varies in wide range and the lunar surface is available for observations from the Earth in practically full range of phase angles, it is possible to study different dependences of optical parameters. For example, the dependence of degree of positive polarization (and maximum of positive polarization $P_{\text{max}}$ in particular) on albedo is studied well enough. However the distribution of $\alpha_{\text{max}}$ over the lunar disk and correlation with other optical parameters are not practically investigated.

Therefore the maps of maximum of positive linear polarization degree $P_{\text{max}}$ and of its phase angle $\alpha_{\text{max}}$ have been constructed for the eastern hemisphere of the Moon, which are based on a set of polarimetric observations of the lunar surface. The observations were carried out at Kharkov Observatory in 2 wavelengths $\lambda_{\text{eff}}=461 \text{ nm}$ ($\Delta \lambda=106.4 \text{ nm}$) and $\lambda_{\text{eff}}=669 \text{ nm}$ ($\Delta \lambda=125.0 \text{ nm}$) with an imaging CCD–polarimeter [1] and a camera lens of 3 cm diameter, and 30 cm focal length. For approximation of phase dependence of polarization the modified Rayleigh function has been used:

$$P(\alpha) = \frac{(\sin^2(\alpha - \Delta \alpha))^W}{1 + \cos^2(\alpha - \Delta \alpha) + \text{dePol}},$$

where $\Delta \alpha$ is a maximum shift parameter, $W$ is a maximum width parameter, $\text{dePol}$ is a depolarization parameter.

The solutions for $P_{\text{max}}$ and $\alpha_{\text{max}}$ (Fig.1) are obtained using observations at 10 different phase angles from $45^\circ$ to $123^\circ$ with fixed values of the parameter $W$ ($W=0.75$ for $\lambda_{\text{eff}}=461 \text{ nm}$ and $W=0.88$ $\lambda_{\text{eff}}=669 \text{ nm}$). Those values of $W$ have been calculated as averaged from previous solution with $W$ variation, because of obtained maps of $W$ parameter are very noisy.

Now we have data of polarimetric observations at phase angles $143^\circ$ and $155^\circ$. Hopefully it allows us to construct the reliable maps of the parameter $W$. The maps of spectral indices $C_{P_{\text{max}}} = P_{\text{max}}(669\text{nm})/P_{\text{max}}(461\text{nm})$ (Fig.2a) and $C_{\alpha_{\text{max}}} = \alpha_{\text{max}}(669\text{nm})/\alpha_{\text{max}}(461\text{nm})$ (Fig.2b) have been constructed too. All the maps are represented in the external perspective projection (distance=221.1739 of $R_{\text{Moon}}$, image radius=225 pix) and are accessible at http://www.univer.kharkov.ua/astron/dslpp/moon/polar as FITS-files. A pixel size is equal to about 8 km on lunar surface.

Data processing was fully carried out using our "IRIS" software complex (http://www.cyteg.com).

A histogram of $P_{\text{max}}$ distribution over the lunar disk has distinct maximum, $P_{\text{max}}=7.3\%$ for $\lambda_{\text{eff}}=461 \text{ nm}$ and $P_{\text{max}}=5.25\%$ for $\lambda_{\text{eff}}=669 \text{ nm}$, corresponding to highlands. Distribution of $P_{\text{max}}$ for maries is more diffuse. The range of $P_{\text{max}}$ variations is $4.0...21.0\%$ for $\lambda_{\text{eff}}=461 \text{ nm}$ and $3.0...15.0\%$ for $\lambda_{\text{eff}}=669 \text{ nm}$.

A histogram of $\alpha_{\text{max}}$ distribution is distinctly bimodal, with the first peak at $\alpha=99.7^\circ$ (highlands), and the second one at $\alpha=104.1^\circ$ (mares) for $\lambda_{\text{eff}}=461 \text{ nm}$. For $\lambda_{\text{eff}}=669 \text{ nm}$ we have $\alpha=96.8^\circ$ and $\alpha=101.2^\circ$, respectively. The histogram is narrower in blue light, $94.0^\circ ... 106.0^\circ$, as compared to red light ($90.0^\circ ... 105.0^\circ$). The maximum of polarization occurs at larger phase angles in the blue band.

Fig. 1. Maps of phase angle of maximum of positive polarization of the Moon

Distribution of $P_{\text{max}}$ and $\alpha_{\text{max}}$ over lunar disk:

Fig. 2a: $C_{P_{\text{max}}} = P_{\text{max}}(669\text{nm})/P_{\text{max}}(461\text{nm})$; Fig. 2b: $C_{\alpha_{\text{max}}} = \alpha_{\text{max}}(669\text{nm})/\alpha_{\text{max}}(461\text{nm})$.
Fig. 2. a) map of spectralpolarimetric index $C_{p\text{max}} = \frac{P_{\text{max}}(669\text{nm})}{P_{\text{max}}(461\text{nm})}$ b) map of $C_{\alpha\text{max}} = \frac{\alpha_{\text{max}}(669\text{nm})}{\alpha_{\text{max}}(461\text{nm})}$

**Distribution of spectral indices $C_{p\text{max}}$ and $C_{\alpha\text{max}}$ over lunar disk:** A histogram of $C_{p\text{max}}$ distribution over the lunar disk has distinct maximum, $C_{p\text{max}} = 0.70$. The range of $C_{p\text{max}}$ variations is $0.65 \ldots 0.77$. The map of $C_{p\text{max}}$ shows significant correlation with lunar details. $C_{\alpha\text{max}}$ has practically constant value over the lunar disk, $C_{\alpha\text{max}}=0.976$ ($\sigma=0.006$). There are no any lunar details on this map.

**Analysis of data:** The analysis of relationships between various optical parameters of the lunar surface was carried out. It was established that: 1) dependence $\alpha_{\text{max}}$ on logarithm of albedo and on logarithm of $P_{\text{max}}$ shows significant linear correlation; 2) the parameters $\alpha_{\text{max}}$ and $P_{\text{max}}$ depend on wavelength via albedo changes only; 3) correlation diagram “spectro-polarimetric index $C_{p\text{max}}$ – albedo” has two branches: there is anticorrelation for mares and correlation for highlands.

**Conclusions:** Obtained maps may be useful for progress of methods of remote sensing of surfaces of the Moon and other atmosphereless bodies and for verification of models of positive polarization of light scattered by regolith-like surfaces.

**References:** [1] Korokhin V. V. et al. (2000) *Kinematika i fisika nebesnykh tel*, 16, No 1, 80-86.
HRSC STUDY OF AN UNNAMED MARTIAN IMPACT CRATER AT 24.5° S, 80.9° E. J. Korteniemi1, V.-P. Kostama1, M. Aittola1, T. Öhman2, T. Törmänen1, H. Lahtela1, J. Raitala1, G. Neukum3 and the HRSC Co-Investigator Team, 1Planetology group, Univ. of Oulu, P.O. Box 3000, Oulu, Finland, <jarmo.korteniemi@oulu.fi>, 2Dept. of Geosciences, Univ. of Oulu, Finland, 3 Inst. of Geosciences, Freie Universitaet Berlin, Germany.

Introduction: An unnamed 95 km crater (Figs. 1, 2) is located on a gentle (0.5°) regional slope on the NE Hellas basin rim. The area is a mix of modified Hellas ejecta (Nh) and cratered highland (Npl1), in places with fluvial channels (Npld) [1].

The crater floor has details which can only be fully identified from HRSC orbit 389. Using the different HRSC channels together with MOLA, MOC and THEMIS data sets, we were able to develop a plausible scenario for the development of the identified features.

Crater floor morphology: The crater floor is divided into 6 morphologically distinct units (Fig. 3).

Crater Floor (CF): The crater floor unit, a 72 km ring, is almost uniformly at an elevation of -1150 to -1200 m below MPR, and 1400 m below the average level of the surrounding plains. The flat CF has some minor texture, and embays the HC unit (see below).

Central massif (CM): A large 35 x 35 km plateau with steep, up to 60° slopes, raises roughly at the crater center. Remnants of similar material can be found at the edges of the HC unit in a semi-circular pattern (fig.2; arrows). Topographically CM is two-fold; central parts are in average 100 m lower than the E, W and S corners or the N remnants. In average the massif is 200 m above the average CF elevation. Numerous small 100 to 1000 m craters cover the entire plateau surface. The W edge of the plateau slopes gently to a 50 m deep depression (Fig. 2;a).

Central Peak (CP): At the center of CM, a different 6 x 8 km CP unit is found. It stands 50 m above the surrounding lowest areas of the CM plateau. The CP material is more rugged than CM and decays at its W part into a knobby terrain. The material is noticeably brighter than the surrounding CM material, with the crater density being approximately the same as in CM.

Honeycomb terrain (HC): A unit with distinctive honeycomb-like ridges and intervening pits creates a crescent-shaped area E of CM. It is best visible in the 17 x 17 km SE area (Fig. 2;b). The ridges are at approximately the same level as CF, with pits ranging 400-1200 m in diameter and 100-200 m in depth. The 40 x 17 km (Fig. 2;c) portion of this unit is degraded, the pit depths ranging mostly from 10 to 100 m.

Knobby material (KM): The crater floor between HC and the E rim is identified as knobby terrain, characteristically 30-100 m lower than CF. The unit is divided into two categories with a subtle transition but different qualities; KM1 (darker, small knobs; fig.2;d) and KM2 (brighter color, yardangs; Fig. 2,e).

Bright material (BM): A unit of very high albedo is recognized in the CM walls and on the bottom of KM pits. This material is always present only as outcrops and thus embayed by other deposits.

Proposed history: Crater formation: The shape of the CM unit and the degraded mesas around HC suggest the presence of a deformed large ring structure (diameter 50 km) at the center of the crater, coinciding with the average Martian peak-ring crater size [2,3]. However, none of the other same size craters in the region have peak rings. The older ones are filled with deposits, while the younger ones have only central peaks. This implies that the formation of this particular crater was unusual due to some unknown factor, perhaps target material at the impact time. Parts of the CM unit (an unusually large central...
peak structure) may be the result of a massive outburst of impact melt during the crater formation.

Deposits and sedimentation: The crater is about 300-500 m deeper than other apr. same size and age craters in the region. This suggests that the crater was originally filled to some extent with deposits, creating a thick layer of material on the original floor. These deposits were later eroded away (see below). The remnant CM unit could also be at least partly the result of this deposition/erosion process. The deposition type is not known, but most probable candidates are layers of lavas, pyroclastic materials and aeolian deposits. There is no undisputable evidence of fluvial or glacial activity, but the floor morphology implies that at least parts of the sediments were rich in volatiles (see below).

Erosion: Sediments were later eroded, revealing a underlying brighter unit (BM) under the deposits and on the slopes of craters. This unit may be the original crater floor, or one of the first deposits on it, as it is layered and embayed by later processes. The CP unit is interpreted as the original central peak of the crater, which may continue to the E under CM sediments.

Tectonism: The erosion of CM and other units has followed the regional weakness directions [4], as seen in the 8.5 km polygonal crater remnant CR1. The tectonic lineaments inside the crater are mainly radial or concentric to the Hellas basin. This is in accordance to our earlier works on the W side of Hellas [5]. There is also a component which is directly radial to Hadriaca Patera in the SE.

Tectonic activity may also explain some small features. The 2 x 7 km plate radial to Hellas in the NE of HC unit (Fig. 2:f) is homogenous, tilted 11°, and clearly the result of tectonism. The depression between CM and crater rim, and the 300 m high gentle slope descending into it from the plateau can be explained by a rigid plate sinking into the crater floor. Additionally, the local slope winds could have gathered material on the E side of the depression to camouflage the perhaps steeper CM-depression slope.

Volatile: The HC unit is probably a result of volatile escape from the crater floor. If volatiles were abundant either in the original floor or sediments superposing it, they gathered in the deepest parts of the floor, i.e. where the HC unit now resides. These sediments may have initially cracked, creating polygonal fractures, which later were filled with more competent material. This happened either from sedimentation above, or from volcanic dykes from below. As the ice/water sublimated or the material became more easily erodable from these reservoirs, karst-like depressions were created between the hardened material. The two coeval 8 km craters in the south support the hypothesis of local volatiles. CR3 has a ballistic ejecta field, while CR4 is a rampart crater [6]. The KM unit has moraine-type features.

Aeolian features: The transition between HC and KM unit has abundant N(NW)-S(SE) yardangs due to wind activity. The latest period of activity in the crater is also aeolian; small dark dunes seen in MOC NA images superpose all the crater floor units.

Fig. 3. Geology of the crater. HL: Highland unit; R: crater rim; W: crater wall; CF: crater floor; CM: central massif; CP: central peak; HC: honeycomb material; KT: knobby terrain; BM: bright material; CR: crater; Lines: tectonics.

Conclusions: The proposed geological crater history is complex. Due to an unknown factor it has a (degraded) peak-ring while the neighbor craters do not. After its formation the crater was filled with two deposit types; volatile-rich and more erosion-resistant ones. The subsequent erosion was influenced by local zones of weakness created by Hellas basin and possibly by Hadriaca Patera. Erosion/weathering removed volatile-rich material, creating the honeycomb-pitted terrain and revealing outcrops of the underlying, perhaps original crater floor BM. Primary N-S wind direction is deduced from the KM yardangs and small dunes superposed on all units.

Acknowledgements: We gratefully acknowledge the efforts made by the MEX-HRSC Photogrammetry Team in processing the digital image data.

Introduction: The eastern rim region of the Hellas impact basin (Fig. 1) is characterized by several large outflow channels: The adjoined Dao and Niger Valles near Hadriaca Patera and to the south, Harmakhis and Reull Valles [1]. Of these, only Reull does not connect to the basin itself. Its main channel ends abruptly close to the source of Harmakhis. A massif with large debris apron shields the possible connection between these two channels. Reull Vallis also differs in its source region. While the other channels emerge from clearly defined depressions, Reull does not. It seems to originate from the southern lava plains of Hesperia Planum. This studied region, a large topographic low located S-SW from Morpheos Rupes has also another channel (proposed by [2] to be upper part of Reull Vallis). It emerges from beneath the Hesperia Planum first as subsurface flow, then continues as on-surface channel down to the topographic depression, Morpheos Basin (MB) (Fig. 2). In this ongoing study we propose that this depression, the evolution of Reull Vallis, and the adjoined plains are tightly linked. Our study also adds a new piece to the puzzle of the Martian watery past.

Morpheos Basin: The basin itself is approximately 200-300 km by 900 km in size. In general, the basin is located ~500-700 m deeper than the surrounding highs, including the northern part of Hesperia Planum. The lowest regions of this basin are located close to the starting point of Reull Vallis. The channel emerging from beneath the Hesperia also enters the basin next to the deepest parts.

The general morphology of the floor of MB does not differ significantly from the lava plains of Hesperia. However, some evidence suggest the presence of volcanites within the basin floor material, predominantly in the lowest regions. The local impact craters (Fig. 3) show distinct morphologies: rampart ejecta, softened floor material which is usually lineated, indicating viscous movement. In addition, the southern massifs con-
some explanation to the subdued morphology of the plains as well as to the local phenomenon: the possible glacial features (e.g. Fig. 4) and the debris aprons concentrated around the massif remnants. The presence of these features is confined within the region connected to the Reull Vallis [3]. Within the middle and lower stretches of the channel, viscous flows from on-surface sources merge with the channel floor material [4].

**Proposed evolution of the Morpheos basin-Reull Vallis connection:** In previous studies [4,5,6] it has been proposed that some material was transported from the Hesperia region creating a void that was partly filled later by the lavas. Some volatile rich material evidently remained (ice lenses, volatile rich sediments, water) resulted in subsurface flows (~32°S, 246.5°W) which changed to an on-surface flow towards MB and Morpheos paleolake. After basin filling water breached the lowest point to the east and formed the Reull Vallis channel. Any natural dam caused overflowing to the adjacent plains resulting in secondary channel for awhile before the main route was cleared. MB was drained but some water remained (ice, sediments, glaciers) both in Morpheos Basin as well as in the Reull region.

**Conclusions:** Based on our observations, we propose that Reull Vallis was formed by a large outflow from an on-surface paleolake in Morpheos Basin rather than from a subsurface melting similarly to the other Hellas outflow channels. Comparable paleolake reservoirs with distinct outflow channels can also be found elsewhere on Mars [7].

The formation and flooding of Reull also affected to the evolution of the surrounding plains. Debris aprons, viscous flows from on-surface sources, deformation of the local features and glacier-like (or ice-rich) material in the craters and massif flanks within the region suggest that some of the water remained in the region for at least some period of time.

**Acknowledgements:** We gratefully acknowledge the efforts made by the MEX-HRSC Photogrammetry Team in processing the digital image data.

DYNAMICS OF CONFINED LIQUID MASS, SPREADING ON PLANET SURFACE. A. A. Kostrikov, Laboratory of Comparative Planetology, Vernadsky Institute, 19, Kosygin St, 117975, Moscow, Russia (kostrikov@geokhi.ru).

Introduction: Ice or permafrost covers (completely or partially) appreciable number of planets and satellites of Solar system. By several ways subsurface water can spring away [1-4]. What are the basic features of the confined water mass spreading? Generally speaking behavior of liquid mass (spot) poured out on the surface of planet depends on many parameters. This investigation had been carried out to solve a simple problem of non-viscous liquid spreading on smooth surface of rotating planet.

Model: The model bases on shallow-water equations with a source. It is proposed that source has a circle form and input of liquid decreases with time exponentially. In dimensionless form task solution depends on three non-dimensional numbers: \( Fric = \frac{C_d}{f} \sqrt{\frac{g}{3}}, T_s/T \) and \( R_{im}/L \), where \( g \) is acceleration of gravity, \( f \) Coriolis acceleration, \( C_d \) friction coefficient, \( H=ST \), a scale of spot depth, \( S \) intensity of source, \( T \) a scale of discharge duration, \( T=1/f \) a time scale, \( R_{im} \) source circle radius, \( L=\sqrt{gH/f} \) is length scale. (All values are non-dimensional below.) For simulation of the spot flow the method of large particles has been used [5-7].

Numerical experiments: Simulations of liquid spreading on Mars, Earth, Europa and Ganymede with \( R_{im}=10 \text{ km}, S=10^{-3} \text{ m/sec}, T_s=10^3 \text{ sec} \) and \( C_d \) from \( 10^{-7} \) to \( 10^{-4} \) have been made. Values of other input parameters and obtained scales are in Table 1.

<table>
<thead>
<tr>
<th>Planetary Body</th>
<th>( g ), m/sec^2</th>
<th>( f ), sec^{-1}</th>
<th>( T ), sec</th>
<th>( H ), km</th>
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<td>50</td>
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<td>79</td>
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Table 1. Values of parameters and scales.

Behavior of liquid spot depends on magnitude of \( Fric \). If \( Fric<0.1 \) the spot performs radial motion. If \( Fric<0.1 \) liquid mass starts to oscillate from center to periphery and back, clock- and anticlockwise. Velocity vector moves clockwise at any point of spot. At sufficiently small \( Fric \) the spot looks like a contractive and unclasping ring (Fig. 1). The origin of these waves is the same as Poincare’s waves [8]. In \( t=2 \) spot volume increasing drops abruptly and then oscillations begin with period of 3.78 and amplitude equals to 0.45. Worthy of note that in contrast to amplitude this value of period is practically constant for all experiments with oscillations. The spot starts to spread strongly with decreasing \( Fric \) under 0.001. Following its front wave crest moves to spot periphery. If \( Fric > 0.005 \), mean position of crest approaches spot center. Further amplification of \( Fric \) results in confused oscillations and their vanish at \( Fric > 0.2 \).

An objective view of spot flow gives dimensional data (Table 2), where \( R_{pl} \) is radius of planet or satellite. One might see that in order to increase upto tenfold initial extent (120-150 km) it will be necessary a little more then 4 terrestrial days for spots on Mars and Earth, 16 days on Europa and a month on Ganymede.
Dynamics of Confined Liquid Mass: A. Kostrikov

<table>
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<td>2634</td>
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</table>

Table 2. Dimensional characteristics of spot movement.

Discussion and conclusions: Being used interval for friction coefficient $C_d$ can appear moved to the range of too small values (For example, $C_d = 0.0025$ for smooth concrete canals [9]). Indeed, you must use appropriate values of $C_d$ to model water flows with rates smaller 10 m/sec. The point is that the higher speeds generate cavitation cushion made of gas bubbles, which reduces friction sharply [10]. Besides, if an ejected mass is a steam and gas mixture originally, it flows at lesser friction.

One can obtain small values of $Fric$ not only by reduction of friction coefficient, but by lessening of acceleration of gravity, or increasing of speed of planet rotation or ejected liquid mass. Thus erupted liquid will spread on planet like a pulsing ring in case that its mass is sufficiently large, and this ejection takes place in high latitudes of a small, fast-rotating planet with smooth surface. Conclusion follows that requirement $Fric < 0.1$ must be fulfilled to realize such structures.

Very likely that these structures can leave behind corresponding ring-shaped trace.

ARE MARTIAN NORTH POLAR CAP SPIRALS TRACES OF ANCIENT ICE SHEET COLLAPSE?
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Introduction: The surface of north polar cap of Mars is essentially heterogeneous unlike flat terrestrial ice sheets [1]. Troughs up to one kilometer deep with gently (no more 10-15°) sloping are seen all over the ice cap. The unique feature of the trough system is its helical appearance (Fig. 1).

Laboratory experiment: A simple laboratory experiment has been run to prove the hypothesis. A disk made of wet clay that measures 14 centimeters across and two tenths of a centimeter at the center (tapering down to 0 at the outer edge) had been thrown counter clockwise. In two minutes of rotating and subsequent drying spiral cracks made their appearance (see Fig. 2). There is a long crack on the right part of the figure, a shorter and S-like cracks in the center. Every crack has a clockwise cockling. One can see a qualitative resemblance of the spiral structure of Martian north polar cap and this clay model.

Model: A simple model of crack progression in viscous ice sheet on rotating planet has been developed. Model crack trajectory depends on Coriolis parameter $f$, initial coordinates, initial velocity ($u_0$, $v_0$) and ice resistance coefficient $\tau$.

Results: An example of model trajectory for $x_0=0$, $y_0=600$ km, $u_0=60$ m/sec, $v_0=0$ m/sec, $f=10^{-4}$ sec$^{-1}$, $\tau=2 \times 10^{-5}$ sec$^{-1}$, visualized on Fig. 3. One can see the model trajectories depended on physical characteristics of ice and Coriolis force.
crevasse looks like a spiral, coiling around center. Qualitative resemblance of helical troughs is evident. It makes sense to verify if they fit quantitatively. An analysis of trough pattern (Fig.1) has been made in order to take angles $\phi$ between its tangent directions and local meridian lines – “spiral inclines” (see Fig.3). Fig.4 shows these data, lines of polynomial interpolation and modeling dependency $\phi\left(\psi\right)$. One can see the scatter of points is sufficiently large at first sight. Trend line says that spiral inclines are above mean (67°) on the sheet periphery, approaching 90°. They diminish near pole and close to 70° in the wide enough latitude range (82.5°-86°). It is interesting that the model line loops a sufficient large loop.

**Discussion:** Model curve behavior is defined by starting conditions and parameters of the problem. Suppose angular rotation velocity of Mars practically did not change since hypothetical collapse of polar sheet, and so used value of Coriolis parameter deserves credit. As a matter of fact the size of ice sheet could be different, though, one can hope, its distinction from present one is a little. The used values of initial speed and resistance coefficient $c$, for their parts, depend on value of ultimate stress, kinematic coefficient of ice viscosity and ice density. Terristial ice investigations show, that ice density varies in sufficiently close limits and only a little smaller of the used value $10^{3}$ kg/m$^3$, but the ranges of ultimate stress and kinematic coefficient of ice viscosity are wide [11]. Mean values have been used in this research. Generally speaking, on account of faint maturity of the cracking dynamics theory application of simple models seems justified. Being examined trough pattern closely, one can notice that no trough traces from sheet margin to the pole continuously, that they, as rule, consist of several sections. This says that cracking happened step by step. Continued spread of ice sheet resulted in a rise of stretching stress at crack vertex that, in its turn, sooner or later, set going spasmodic initiation of a new crack. Flatness of trough slopes denotes that collapse accompanied by creation of immense crevasses, took place a long time ago. After reduction of geothermal flux to the previous level bed friction rose, ice sheet spreading dropped down, cracking stopped, and accumulation began to play a key role in sheet surface modification. Thus, regardless of the fact that the ice spreads away slowly (order of speed magnitude is mm/year [7]) smoothing irregularities of its surface, accumulation process drives helical troughs irrepressibly north. If accumulation process did not start up, deep troughs would close in $10^5-10^6$ years [7].

By the way, absences of spiral troughs on the surface of terrestrial ice sheets can stands for remoteness or lack of their collapses. However one can think that being disappeared to the end of last glacial period Laurentian and Fennoscandian ice sheets could left footprints (moraineas?) of their helical structure. One must take into account that underlying topography could disfigure this structure substantially.

**Conclusions:** Thus, it follows from this investigation that being defreezed at bed the Martian north pole ice sheet began to transform, as a matter of fact, to an ice body resembling ice shelf. This transformation was accompanied by drastic amplification of radial tension that came to breaking of ice entirety, to emergence of deep crevasses all over the sheet. This planetary scale process was so intensive that being influenced by Coriolis force crack trajectories deviated to the right, forming spirals. After bed temperature fell down and sheet collapse ceased, obtained relief began to undergo a smoothing owing to continuous slow ice spreading and mass transfer from the warmed by sun north crack slope to the shady south one. This process transformed the helical structure of crevasses to the helical structure of troughs.

**Acknowledgments:** The author met with support from Dr A. Basilevsky. Thank to Mr. Ken Turner, who run a laboratory experiment with the clay slab, very much. Discussions with V. Bogush, M. Krass, L. Ingel, S. Netreba were helpful.

THE CRATERS SHOEMAKER AND FAUSTINI AS COLD TRAPS FOR VOLATILES
E. A. Kozlova¹, V. V. Shevchenko¹ . Sternberg State Astronomical Institute, 119899, Moscow, Russia

Introduction: In 1994, the “Clementine” spacecraft launched by NASA explored the Moon for 70 days. As a result, a radar experiment made it possible to discover areas with anomalous radar properties (Nozette et al., 1996). The “Lunar Prospector” spacecraft, launched by NASA toward the Moon in 1998, was equipped with a neutron spectrometer for detecting possible deposits of volatiles in the polar areas of the Moon. In the region of the south pole of the Moon, maximum hydrogen content was found in the areas coinciding with such craters as Faustini (87.2° S, 75.8° E, D = 45 km) – 160.3 ppm and Shoemaker (88° S, 38° E, D = 56 km) - 146 ppm, [2]. The average level of the contents of hydrogen for the Moon makes 50 ppm.

We computed the permanently shadowed areas and temperatures inside the craters for the variation of the position of the lunar pole of rotation with respect to the ecliptic pole with a period of 18.6 years. We estimate the total permanently shadowed area in the lunar northern polar region as 28260.2 km² [3]. The total permanently shadowed area in the region of the south pole of the Moon is smaller than in the region of its north pole and we estimate it as 22168.5 km². According to our results, the total permanently shadowed area in the polar craters of the Moon is equal to 50428.7 km², or 0.13% of the total area of the lunar surface [3].

For an estimation of temperature of a surface in a crater we took into account the direct solar flux coming to an illuminated element of the crater surface, the thermal flux from the interiors of the planet, the flux reflected from the illuminated surface of the crater, the secondary reflected light flux from element of the inner surface of the crater, the infrared flux incident onto crater surface element. We do not take into account the thermal flux from the adjacent elements of the crater; however, this flux can be ignored because of the low thermal conductivity of regolith on the Moon. In addition, we also ignore the effect of the solar wind, because it barely penetrates into permanently shadowed areas [4].

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 [5]. And sulfuric compounds remain stable if the maximum temperature does not exceed 220K [5].

Crater Shoemaker (88° S, 38° E, D = 56 km) was found on the images received with the help of observatory Goldstone [6]. In view of position of the Earth above lunar horizon at the moment of radar-tracking measurements the structure of a crater has been constructed. Average depth of a crater has made 2.75 kms, an inclination of walls - 13°. The area of permanently shaded part of a crater makes 59 % from all area of a crater [3]. The diagram of distribution of the maximal temperatures in a crater shows, that in northern part of a crater there is an area where the temperature never exceeds 70 K. The average temperature of this area – 40 K Thus, this part of crater Shoemaker can be “a cold trap “ for water ice or NH3. In the central part of a crater the maximal temperatures reach at northern slope 100 K, and 140 K at southern. Average temperatures in this area do not exceed 50 K and 70 K accordingly. In such conditions in this part of a crater open deposits of water ice or deposit of water ice under a layer of regolith, and as connections of sulfur can be kept. The maximal temperatures in area of a southern slope of a crater exceed 180 K.

On images of area of South Pole of the Moon, received by observatory Goldstone [6] (fig. 1) as possible “cold trap”, we have allocated crater Faustini (87.2° S, 75.8° E, D = 45 km). The maximal temperatures in northern part of a crater does not exceed 87 K, the average temperatures in this area change from 47 K up to 57 K. This area can be a cold trap for the deposits of the water ice which has been not covered with a layer of regolith. The maximal temperatures in a significant part of northern half of crater do not exceed 100 K and can contain open deposits of water ice. The maximal temperatures inside south part of the crater Faustini do not exceed 200 K so connections of sulfur can contain in any part of a crater.

We found that craters Shoemaker and Faustini containing permanently shadowed areas in which the temperature allows volatiles to remain stable for a long time coincide with the areas of high hydrogen content according to the “Lunar Prospector” data and can be “cold trap” for volatiles, including water ice.

References:
Figure 1. The radar image of lunar south pole region obtained with the 3.5 centimeter wavelength Goldstone Solar System.
PERMANENT CO$_2$ DEPOSITS ON MARS AT LOW OBLIQUITY. M. A. Kreslavsky$^{1,2}$ and J. W. Head$^2$

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**Introduction:** It has long been understood that secular variations of spin and orbit parameters of Mars strongly influence the climate of the planet [e.g., 1] through variations in their spatial and temporal insolation pattern. Changing parameters that control the insolation are obliquity $\theta$, eccentricity $e$, and, for epochs of large eccentricity, season of perihelion, which can be quantified as areocentric longitude of the Sun from the moving equinox at perihelion $L_p$. Among these three parameters, obliquity has the strongest impact on major climate characteristics. The obliquity of Mars oscillates quasi-periodically, varying up to $-\pm$10° amplitude and a period of $\sim$0.12 Ma about a mean value, which, in turn, experiences wide variations at the $\sim$5 Ma time scale. These long-term variations are dynamically chaotic, hence they cannot be traced by calculation back in time farther than $\sim$10 Ma. Recent calculations [2] showed that it is probable that the typical obliquity in the Martian geological past was higher than the present. The same calculations showed that it is probable that the planet spent some part of its geological history at low obliquity (in comparison to the present $\sim$25°). It cannot be excluded that there were geologically long (10s or even 100s Ma) periods when obliquity oscillated about values as low as 10°-15°.

**Fig. 1** presents an example of such possible oscillations from [2]. Due to the dynamically chaotic nature of the obliquity variations, only the geological record can reveal if there were such periods in the geological history of Mars.

It has long been understood that at low obliquity, collapse of the atmosphere occurs [e.g., 3], because insolation of polar regions is very low, and the atmospheric pressure is buffered by permanent solid CO$_2$ deposits at the poles. The dependence of pressure on obliquity to first order can be found from a simple radiative balance model: year-average surface temperature at the pole is obtained from the equation relating year-average insolation and thermal radiation, and year-average pressure is obtained from the equation relating year-average surface temperature and CO$_2$ frost point [e.g., 3].

To understand better the nature and distribution of solid CO$_2$ deposits at low obliquity, we constructed a second-approach radiative-balance model for CO$_2$ deposition.

**Model:** Our second-approach model resolves latitudes and seasons. Very similar models were used long ago to model seasonal CO$_2$ caps on Mars. We calculate day-average insolation for a given latitude and time of year. For temperature equal to the CO$_2$ frost point, we calculate the rate of CO$_2$ condensation or sublimation from the energy balance, which involves insolation, thermal radiation and release or consumption of sublimation latent heat. To derive the rates from this energy balance, we need to assume thermal infrared emissivity of solid CO$_2$ deposits, $E$, and visible albedo, $A$. We assume that $E = 1$ is sufficiently accurate. Following [3] we took a "nominal" value of $A = 0.65$ and studied the effect of its variations. Of course, no sublimation occurs in the model calculations when there is no solid CO$_2$ on the surface. We trace the amount of CO$_2$ at the surface at each latitude through the year and recalculate CO$_2$ frost point depending on the atmospheric pressure, which is simply taken from the amount of gaseous CO$_2$ on the planet. We start from the total amount of CO$_2$ approximately equal to the present and no solid CO$_2$ at the surface. Our model reproduces reasonably well the deposition and removal of seasonal CO$_2$ caps and seasonal pressure variations at the present epoch.

We ran our model through an example of spin-orbit variations with persistently low mean obliquity (Fig. 1) for "real" time series for $\theta$, $e$, and $L_p$ from [2]. Our results confirmed that (1) the first-order model (for the same $A$) gives quantitatively accurate estimates of mean atmospheric pressure for a given obliquity; (2) $e$ and $L_p$ have little influence on atmospheric pressure at low obliquity; (3) the onset of both perennial CO$_2$ deposit formation and total atmosphere collapse strongly depends on $A$; (4) the spatial extent of perennial CO$_2$ deposits is smaller at lower obliquity.

**The role of surface topography:** We included surface topography in our model. This provides a principal advance in comparison to previous studies. In addition to the latitude and season, we consider surface slopes of different steepness and orientation in our insolation calculations. We took the actual latitude-dependent frequency distribution of slope steepness and orientation from high-resolution polar gridded MOLA topography. Slopes were calculated at $\sim$200 m baseline (slopes at lower latitudes, where MOLA data are insufficient for slope calculation at this baseline, play no role in our calculations, because perennial CO$_2$ deposits never form at lower latitudes). Poleward from 86° latitude, where there is no dense MOLA coverage, we assumed the surface to be horizontal. With this
modification, we ran the model in the same way, as described above.

Results: The results differ drastically from the case of a perfectly spherical planet. Fig. 2 shows the example of the calculated evolution in the obliquity - pressure domain. The total year-average mass of solid CO₂ deposits is proportional to the deviation of the pressure from 800 Pa. The pressure shows wide hysteresis: it increases with the obliquity increase much slower than it decreases with the obliquity decrease. The periods of relatively high obliquity are too short to evaporate all surface CO₂ deposits, and the pressure does not return to its maximal value even at high (24°) obliquity peaks. This behavior is caused by deep cold traps that are provided by surface topography. Steep (>~10°) pole-facing slopes at the 70-80° latitude zone are the coldest places on the planet for obliquity below 20°. Massive CO₂ deposits are accumulated in these cold traps. The total area of these places is small, and the integral rate of sublimation, being proportional to the area, is also small. This leads to survival of much solid CO₂ on the surface during a geologically long time. There is not enough time to bring the system to the equilibrium atmospheric pressure during obliquity maxima.

Discussion and conclusions: Results of model calculations (such as shown in Fig. 2) should not be used as firm predictions of the pressure evolution and deposit distribution, because there are too many uncertainties in the model parameters. Albedo A strongly depends on the microphysics of the CO₂ deposit surface and can change with time, as it is actually observed for present-day seasonal deposits. At relatively higher obliquity, when polar areas become warmer and some H₂O vapor appears in the atmosphere, the patchy residual CO₂ deposit can act as traps for H₂O frost. This would increase A and protect the deposits from sublimation. All these effects are not included in the model. In addition, the model does not include the feedback between CO₂ deposition and topography: the topographic cold traps are modeled as having infinite capacity, while in reality most of them are related to small-scale topographic features and will be quickly filled with solid CO₂.

Despite the incompleteness of this model, our main conclusions are robust. First, the distribution of the perennial CO₂ deposits and atmospheric pressure is not a simple function of obliquity, but crucially depends on the geologically long pre-history of climate variations and solid CO₂ deposits.

The second conclusion is that massive CO₂ deposits at pole-facing slopes at high latitudes are formed during low obliquity periods. Geologically long oscillations of obliquity around low values should lead to formation of associated H₂O ice deposits in these places, which can produce morphologically observed traces at the surface. Identification of morphological evidence of such deposits could potentially point to actual geological epochs of low obliquity in Mars history.

Models of the internal structure of Europa and Callisto based on Galileo gravity measurements and geochemical constraints on the composition of silicate fractions of ordinary and carbonaceous chondrites are constructed. A permissible thickness of Europa’s outer water-ice shell lies between 100 and 160 km for these classes of chondritic matter. The total thickness of an outer water-ice shell of Callisto is up to 270-320 km. The results of modelling support the hypothesis that Callisto may have an internal liquid-water ocean.

**Introduction.** The purpose of this study is to reproduce characteristic features of the internal structure of Europa and Callisto (thickness of an outer water-ice shell, mantle composition and density, and core sizes and masses) on the basis of Galileo gravity measurements (the mass and moment-of-inertia factor) and geochemical constraints on the composition of silicate fractions of ordinary and carbonaceous chondrites, which are taken as representatives of nebula matter. The general methodology is to combine the geophysical and geochemical constraints and thermodynamic approach, and to develop, on this joint basis, the self-consistent models of Europa and Callisto, accounting for their composition and internal structure.

**Approach.** The mass and mean moment of inertia [1,2] are used as input data for determination of (1) the thickness and phase state of an outer water-ice shell, (2) the density distribution with depth, and (3) the core sizes and masses. Two compositional models are considered for a core: an Fe-10 wt.%S core for ordinary chondrites and an FeS core for CM chondrites. The phase compositions and mantle densities are modelled within the system Na$_2$O-TiO$_2$-CaO-FeO-MgO-Al$_2$O$_3$-SiO$_2$ including the solid solutions. The equilibrium phase assemblages were calculated using the technique of free energy minimization and thermodynamic data for minerals summarized in the THERMOSEISM database [3]. The density variations in the mantle and core radii are found by the Monte-Carlo method.

**Europa’s composition and internal structure.** The results show that Europa is differentiated into a water-ice shell, anhydrous mantle and iron-sulfide core [3]. 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). Within these models, the permissible thickness of Europa’s water-ice shell lies between 105 and 160 km (6.2-9.2% of total mass) for any model of differentiated or undifferentiated chondritic matter, Fig. 1. The amounts of iron and iron sulfide, and the (Fe$_{core}$/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, Fig. 2. 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. Core radii are estimated to be 470-640 km for the L/LL chondritic models with a central Fe-10 wt.%S core (5.3-12.5% of total mass). The allowed thickness of Europa’s H$_2$O 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.

**Callisto’s composition and internal structure.** The problem of modeling the internal structure of Callisto is described by a system of equations specifying the conditions of thermodynamic and hydrostatic equilibrium, equations of state and heat conduction, and mass and moment conservation. We consider a six-layer model of Callisto consisting of an ice layer, a water ocean, a three-layer rock-ice mantle, and a rock-iron core. In this study we proposed that rheological behaviour of the ice-I layer was non-newtonian. In this case the outer ice shell becomes stable against convection [4]. In agreement with [4], the distribution of temperature can be calculated from the conditions of conductive heat transfer in an ice-I layer, and the temperature profile in the fields of stability of water and high-pressure ices goes along the adiabat. Internal structure of Callisto with an internal ocean is shown in Fig. 3. The maximum thickness of the outer water-ice shell is 315-320 km, Fig. 4. The radii of the rock-iron core for a model of Callisto with an internal ocean are in the range of 0-950 km. Concentration of ice (40 wt%) is constant in the rock-ice mantle if the total thickness of an outer ice shell is maximal (~320 km). The content of water and ice in Callisto is between 47 and 54% for a mantle model composed of dry silicates + hydrous silicates and 49-54% for a mantle model composed of dry rock. The total thickness of an outer water-ice shell of Callisto is up to 270-315 km. The permissible thickness of an icy crust and internal ocean are estimated to be 135-150 km and 120-180 km respectively, Fig. 4. The results support the hypothesis that Callisto may have an internal liquid-water ocean.

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Kuskov and Kronrod, CHEMICAL DIFFERENTIATION OF EUROPA AND CALLISTO

Fig. 1
Fig. 1. The effect of the mantle density on the thickness of an outer water-ice shell and core sizes for chondritic models of Europa’s differentiated (dashed lines) and undifferentiated (solid lines) mantle. For the L/LL material, the allowed thickness of H$_2$O layer ranges from 105 to 145 km; for the CM material, H(H$_2$O) = 125-160 km.

Fig. 2
Fig. 2. Element ratios for Europa (empty boxes) and chondrites (shaded boxes). The empty boxes outline an allowed iron to silicon ratio and content of pure iron, and show that Europa’s bulk composition appears to be within the composition of the L/LL chondrites. Geophysical and geochemical constraints show that H and C chondritic materials may be excluded for the bulk compositions of the satellite.

Fig. 3
Fig. 3. Internal structure of Callisto with a subsurface ocean. Because Callisto is only partially differentiated, 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 maximum radius of the rock-iron core is 950 km.

Fig. 4
Fig. 4. The permissible thickness of Callisto’s water-ice shell. If heat flow from Callisto is in the range of 3.3-3.7 mW/m$^2$, the permissible thickness of the water-ice shell would be 270-320 km. The thickness of an icy crust is estimated to be 135-150 km. The thickness of an internal ocean is found to be 120-180 km.

EVIDENCE OF THE SEASONAL REDISTRIBUTION OF WATER IN THE SURFICIAL MARTIAN REGOLITH BASED ON ANALYSIS OF THE HEND MAPPING DATA.

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Introduction. The global mapping of the neutrons emission from the Mars, conducted recently by HEND instrument from “Mars Odyssey” spacecraft, have shown that the surface layer (1-2 m) on the high latitudes of the planet (up to 50°) is very reached by water ice with abundance more 50% by mass [1,2,3]. It was also shown that water ice distribution in surficial layer of the northern and the southern sub-polar regions is notably different [4]. Until today the existing HEND data already covers the period more then one the Martian year. This let to study the seasonal effects of volatiles redistribution associated with processes of sublimation and condensation of the seasonal polar caps and water exchange between the surface regolith and atmosphere. The goal of our work was to analyse the dynamic of the globally mapped neutrons flux as key to understanding of the seasonal redistribution of the water ice in the surface layer. For this we analyzed the globally mapped flux of the neutrons with different energy and corresponding effective layer of their emission.

Observations. The global mapping of the neutrons emission from Mars has been realized at different energy ranges of the neutrons: by two ranges for both epithermal (100eV-10keV and 10keV-1MeV) and fast neutrons (1Mev-2.5MeV and 2.5MeV-10MeV). The corresponding effective layers from where the neutrons emitted are equal to ~1.5-2 m, ~1 m, 20-30 cm and ~10 cm respectively for indicated energy ranges. We analyzed the data as function of areocentric longitudes (Ls) and the latitude. The data were averaged in 10° range of latitude and in 15° range of Ls and have been normalized to the neutrons flux emitted from Solis Planum as the driest region on Mars. Using the data, the maps of the normalized neutrons flux dynamic during one the Martian year for different neutrons energy ranges. For epithermal neutrons (100eV-10keV (a) and 10 keV-1MeV (b)) and fast neutrons (1Mev-2.5MeV (c) and 2.5MeV-10MeV (d)).

Figure 1. The maps of the normalized neutrons flux dynamic during one the Martian year for different neutrons energy ranges. For epithermal neutrons (100eV-10keV (a) and 10 keV-1MeV (b)) and fast neutrons (1Mev-2.5MeV (c) and 2.5MeV-10MeV (d)).
sublimation of the caps. At that, on the latitudes less of 60° the neutrons flux is mostly constant during all year. Other situation is found for the dynamic of higher energy epithermal and fast neutrons flux (with thinner effective layers of the neutrons emission): in the northern hemisphere it remarkably different than in southern hemisphere. As it well seen from fig.1b,c,d, two distinctive “hollows” of neutrons flux reduction have been appeared in the northern hemisphere during northern summer at Ls=130°-170° and in first half of northern winter at Ls=270°-330°, being extended from high to low latitudes. At that, later “hollow” (Ls=270°-330°) is characterized by much stronger reduction of the neutrons flux and it traces from northern polar region up to low latitudes in the southern hemisphere. The first “hollow” is related with periods of the northern middle summer, while the second one – with of the southern middle summer. In both case the residual polar caps serve as main source of the water in the Martian atmosphere. It is remarkable that during period with Ls=270°-330°, when the seasonal cover of CO₂- ice must be formed on the latitudes >60°N, the noticeable decreasing of the neutron flux is observing in both the sub-polar and middle-latitude regions. Observing reduction of the neutrons flux in the northern sub-polar region represents the sagging on background of monotonic increasing of the neutrons flux in the autumn/winter period (Ls 200°-360°) associated with growing of seasonal cover from CO₂-ice. At that, there is the next distinct tendency: the higher energy of the neutron (or thinner the effective layer), the much stronger reduction of the neutrons flux is observing (fig.2). As it seen from fig.2, the percentage of the neutrons flux reduction (in each energy range) increases very slowly in the latitude range 90°N-70°N and much intensively in 70°N-60°N, being constantly higher on the higher latitude and in most shallow effective layer.

**Discussion.** Because the value of the neutrons flux are sensitive to water abundance in the surface layer [5], the observing effect of the neutrons flux reduction may to be considered as indicator of some temporal increasing of the water content in the surface layer in the time range of the year. It is known [5] that even increasing of the water abundance in the surface material on 1% of mass may result to reduction of the emitted fast neutrons approximately on 10%. The observing maximum value of the reduction of the higher energy fast neutrons flux (in the period with Ls=270°-330°) approaches ~ 20%. That is the abundance of the water (in form of water ice or clathrate CO₂₆H₂O) in ~ 10 cm thickness surface layer could be increased notably in the period on. As water source for this apparently serves the water vapor mass transferred meridionally to here from residual southern polar cap due to the significantly lower partial pressure of H₂O over the cold surface of the northern seasonal cap and its surrounding region. We suggest that the visible reduction of the neutrons flux outside of the seasonal cover of the CO₂-ice (up to the equatorial regions), may to be associated with both condensation of the H₂O frost on the surface and hydration of a salts minerals (mostly sulfates and chlorides) contained in the surface layer of the regolith. The neutrons flux reduction observing in the period Ls=130°-170° is rather associated with hydration process in the surface regolith due to high atmospheric humidity in the period.


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UNUSUAL FEATURES OF WIND-RELATED EROSION WITHIN A SMALL IMPACT CRATERS IN CHRYSE PLANITIA ON MARS. R.O.Kuzmin¹, I.V.Kuznetsov¹ and R.Greeley², ¹Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, 19 Kosygin str., Moscow 119991, Russia, e-mail: rok@geokhi.ru, ²Department of Geological Sciences, Arizona State University, Tempe, AZ 85287

Introduction. High resolution Mars Orbiter Camera (MOC) images [1] show wide variety of the aeolian features within the impact craters attributed to wind erosion and deposition. The features include different types of bright and dark wind streaks behind the craters, duneforms, bright transverse dunes, interacrat deposits, and rim scouring forms. The orientation of these aeolian features is consistent with the direction of current strong winds [2,3], while origin of some of them could be related with paleowind regime[4]. Here we present the results of study of unknown before phenomenon of the wind-related modification of the impact craters on Mars in the form of blowout hollows which have been found only in two places on Mars: mostly in southern part of Chryse Planitia and in rare case in south-western part of Elysium Planitia.

Observation. The studded features represent the circular and elliptical relatively shallow depressions (20-200 m in diameter), settled predominantly on the interior craters slopes of SW-SE exposition in quantity of one, two, several and numerous forms (fig.1). Average size of the depressions is 105 m along the long axis and 70 along short one. The range size of the craters with such features is 100-2000 m.

Often the depressions have distinct rim. In rare cases the features also are placed on a hills and escarments slopes with the SW-SE exposition (see fig.1c,d and fig.2). In the cases with several or more depressions within crater, the features are organized in chain of overlapping each other forms.

Fig.1. Examples of the blowout hollows within the impact craters (a,b), on the slopes of the hills (c) and escarpments (d)

Fig.2. Frequency distribution of the blowout depressions within impact craters, on hill’s and scarp’s slopes.

As seen from the map, the area of the features distribution is elongated in NE-SW direction and is located in transitive zone between the lowland surface of the Chryse Planitia and the highland terrain of western Arabia Terra.
Our study show that disposition of the blowout depressions within the craters directly correlates with both the leeward facing interior slope of the craters and orientation of the bright craters streaks (fig.4a).

The craters streaks are predominantly associated with the strongest winds (blowing from NE to SW), which are predicted by Mars GCM [5] for winter season. Similar correlation has been found also for the depressions located on the leeward slopes of the hills and escarpments (see fig.4b). Apparently exactly those winds were responsible for observing blowout depressions within impact craters.

**Discussion.** We interpret the features as blowout depression formed due to localized and high shear-stress of vortical air movements within the craters. Such air movement may to be resulted by leeward air flow separation in the form of vortex with vertical axis, which have been formed on (and below) the edge of the crater rims, the escarpments edges and the hill’s leeward slopes. It is not excluded that much more higher radiation heating of the craters interior slopes with SW-SE exposition during winter time could to increase the effect of flow separation on the leeward crater slope. The blowout depressions located on leeward slope of Bruneau star dune in western Idaho may to serve as terrestrial analogy of observing features on Mars. The experimental modeling of the wind flow streamline of craters models (filled by fine sand), conducted in Wind Tunnel Facility of Arizona State University, has demonstrated appearance of similar blowout hollow on leeward internal slope of the crater model due to formation in the place of vortical air movements below the edge of the crater rims. The experimental results well support wind-related origin of the studded blowout depression within the Martian small impact craters.


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ANALYSIS OF CRATER LAKES AND LAKE CHAINS IN EASTERN RIM REGION OF HELLAS BASIN, MARS: PRELIMINARY RESULTS FROM THE MEX HRSC DATA. H. Lahtela¹, V.-P. Kostama¹, J. Korteniemi¹, J. Raitala¹, M. Aittola¹ and G. Neukum² and the HRSC Co-Investigator Team. ¹Planetology Group, Astronomy, Department of Physical Sciences, P.O. Box 3000, FIN-90014, University of Oulu, Finland, <hlahtela@paju.oulu.fi>. ²Institute of Geosciences, Freie Universität Berlin, Germany.

Introduction: The southern hemisphere of Mars has one of the largest impact basins in the solar system: the Hellas Basin. This basin has now a depth of ~9 km and a diameter of ~2000 km [1]. Hellas region is rich in details of quite recent fluvial formations. In addition to the dominating outflow channels of the eastern rim, we also have a very high number of smaller channels on both sides of Hellas. In many cases these channels are associated with craters as well as troughs and depressions of the region, creating a good case for suggesting the existence of water reservoirs in some point of regional geological history [2, 3].

In this still ongoing study we examine these probable, but now dry reservoirs in the greater Hellas region. Among others, Hellas basin and its complimentary regions are one of the "sites" that the HRSC-camera has been viewing several times already during the European Mars Express mission revealing new details of Martian features and offering new tools to study Martian geology. The stereo and color capabilities are particularly usable in studying the fluvial formations (channels, depressions, sediments etc.), although also the high resolution of the camera (~10 m/pixel with HRSC and 2.3 m/pixel with SRC), combined with larger areal coverage than that of the MOC narrow angle and THEMIS-VIS gives us much needed information of the targets studied.

Morphology of the studied area: Late-stage effusive volcanism of the Tyrrhena Patera probably triggered events forming the outflow channels (Dao, Nijger, Harmakhis and Reull Vallis) and possibly even larger scale flooding of the region [4]. This, along with other possible processes (i.e. glacial activity [5]) has probably contributed to the fact that the Eastern rim of Hellas has been practically eroded away.

Within the region there are also several craters and depressions which show interaction with post-impact fluvial activity. The crater lakes are in fact one of the most prominent features of lacustric processes in the Hellas region. In this study we use a morphological classification by Cabrol and Grin [3]. The crater lakes are either closed systems with only a distinct inlet channel or open systems with the additional outlet channel. If there are at least two craters connected by a same channel, a lake chain is formed.

![Figure 1: An open crater lake at the rim of Hellas basin with clear in- and outlet channels. It has also prominent large delta-like structure within. (A from HRSC image h0389_000.nd3.02, B from MOLA)](image)

A crater lake: In our earlier studies, the impact crater shown in Fig. 1 has been classified as an open crater lake (e.g. [6]). New research material supports this interpretation by allowing us to concentrate on smaller scale features. For example in Fig. 1a a feature resembling delta front and the channel within the crater are clearly visible. In addition, in Fig. 1b a channel is seen (arrows) on the topset of the formation, supporting the hypothesis of a delta.

Example of a lake chain: A very prominent lake chain in the eastern Hellas basin region is shown in Fig. 3a. Together with older datasets HRSC-images have given enough broad and accurate information for
a meaningful interpretation while studying the lakes’ morphology and thus also their formation.

The water in this lake chain system originate from eastern subsurface reservoirs, perhaps water or ice lenses. First water flowed as subsurface drainage and then bursted to the surface causing an oak leaf shaped depression (Fig. 3b).

From that depression water ran to the largest impact crater in the lake chain. The amount of water there is unknown. On one hand, the stream probably used the crater just as a route, possibly forming a small lake/lakes on the way. This idea is supported by the narrowness of the outlet from the crater. On the other hand, the channel within the crater can’t be recognized. This implies that the crater was filled with water. This is also supported by the fact that the crater floor has been shrunk at least at its outer limits, creating a distinct small slope facing the rim (Fig. 3a arrows) implying that the material was rich in volatiles.

After the first crater, water reached a depression which has since then filled partly with darker sediments (Fig. 3a). From this reservoir water has passed through this area mostly subsurficially indicated by a large scale sapping outlet at the southern part of the area (Fig. 3c). There could have also been a surface flow since there is a small channel visible in Fig. 3d. However, this might be a collapsed tunnel carved by subsurficial stream.

After being released back to surface, the water flowed to a smaller crater. From there it continued west merging with other flows partly creating the wast drainage network at the eastern Hellas basin region.

**Conclusions:** Fluvial as well as lacustrine processes are important factors in post-impact modifications on Mars [6]. The studies of Hellas show the region to be rich in formations of both fluvial and lacustrine origin (e.g. [2]). The MEX HRSC data has already proven to be valuable together with other datasets in analyses of fluvial and lacustrine features.

**Acknowledgements:** We gratefully acknowledge the efforts made by the MEX-HRSC Photogrammetry Team in processing the digital image data.


Fig. 2: A lake chain at the eastern rim of Hellas basin. (A, B and C from HRSC image h0248_0000.nd3.01, D from THEMIS mosaic 101882002, I8211002, 106376002)
AUTOMATIC COMPILING OF HYPsomETRIC MAP OF A PART OF THE VENUSIAN SURFACE. E.N. Lasarev 1, J. F. Rodionova 2, 1- Geographical faculty M.V. Lomonosov Moscow State University, 2- Sternbrg State Astronomical Institute, Universitetskij prospect 13, Moscow 119992, jeanna@sai.msu.ru

Altimetric data of “Magelan” spacecraft has been used for compiling the Hypsometric map for Ishtar Terra region [1, 2, 3]. The main principles applied in compiling the map such as the scale of the heights, types of prints and using of pictures of surface shall be apply in the compiling of the Hypsometric Map of Hemispheres of Venus in future. We consider that contours are the necessary features for hypsometric maps because allow to make measurement more accurate. The main tasks of compiling of the Hypsometric map are: treatment of original hypsometric scale, types for different features of relief and composition. We used the Magelan topographic data GTRD in the appearance of map in equal-square cylindrical projection, reformed in ESRI format Raster Grid. The solution of this map is 5x5 km on pixel. The images of the map on Ishtar Terra region (from 40º till 75º north latitude and from 325º till 15º longitude) have been reformed from equal-square cylindrical projection to equal-intermediate azimuthal projection with the use of Arc View and Arc/Info programs. Then a compiling of the hypsometric map was fulfilled in Arc View with the application Garticules and Measured Grids and 3D Analist: the coordinate grid, new hypsometric scale, contours and names of features in Russian were added. The contours have been passed through 500 m for the heights from -2 500 m till 4 000 m, but above 4 000 – through 1000 m. We selected multi-color and lightening to the summit scale. Such scale is more expedient for mountain relief (4). Different types of prints were used to show polygamy of relief forms. The list of names adopted by IAU (5) and Russian version [6] of the names were the main issue for mapping. Fig. 1 represents a fragment of Hypsometric map of Venus compiled by us. The color of equal-quadrangle projection have been changed in accordance with our height scale. We fulfilled the comparison of this fragment of the hypsometric map of Venus with the map [7] compiled in USGS at a scale 1:16 354 349 for latitudes 0º and at a scale 1:10 000 000 for near polar region. Contours on the map [7] are drawn trough the interval – 1000 m. The scale of the heights on our map (through 500m till the height 4000m) more detail underlines and reflects the different forms of relief of Venus. Moreover many forms of relief and contours on our map have signatures. It makes reading the map easy. The advantage of the map [7] is the available of the shading of relief and smoothed contours.

Acknowledgement: We thank Trent M. Hare from USGS and Dubinin M from the center of wild nature for the assistance.

Fig. 1. Hypsometric map of a part of Venus in equal-intermediate azimuthal projection and the map of surface of Venus in equal-quadrangle projection (color have been changed).
A comparative study of magnetic and nonmagnetic phases in Atlanta (EL6): representative of EL parent body. Z.A. Lavrentjeva, A.Yu. Lyul, N.A. Shubina, G.M. Kolesov. V.I. Vernadsky Institute of Geochemistry and Analytical Chemistry, RAS, Moscow, Russian Academy of Sciences, 117975, Moscow, Russia.

**Introduction.** Enstatite chondrites are an invaluable source of information concerning the chemical and physical processes that were active in the solar nebula and also for understanding metamorphism under reducing conditions during the early periods of the solar system. Temperature minima were determined from the enstatite-diopside solvus of Carlson (1) and indicate that the EL6 chondrites have been metamorphosed at the temperatures exceeding 900-1000°C. The temperature of the enstatite-plagioclase-quartz eutectic (1090°C), appropriate for EL6 plagioclase compositions, was used as the metamorphic maxima. Nitrogen fugacities determined for sinoite (Si₂N₂O) (2) bearing EL6 chondrites (fN₂ =10⁻¹ at 1050°C) indicate that EL6 meteorites either formed in a nebula of high pressure or more likely, are products of metamorphism.

**Samples and method.** In the present paper the results of elemental abundances in separated grain-sized magnetic and nonmagnetic fractions, enstatite from Atlanta are reported. The fractions were selected by handpicking under microscope and by particle-size analysis. Their elemental composition was determined by INAA using a technique for numerical substraction of the matrix element backgrounds (3). The tables show the average element enrichment factors relative to C1 (4).

**Results and discussion.** Of 12 grain-sized fractions of Atlanta EL6 analyzed for siderophile elements, 6 magnetic (metal, schreibersite) fractions have ratios [(Fe/Ni) A/ (Fe/Ni) C1] = 0.7 (mean) less than cosmic and nonmagnetic (sulfides, silicates) fractions – 2.4(mean) greater then cosmic. The pure (yellow) enstatite has ratio Fe/Ni = 9.4 (17.4 cosmic) less than cosmic. This fact supports the opinion that the main process controlling of the composition magnetic phase was sulfurization of metal in protoplanetary nebula. The Atlanta enstatite chondrite show a typically igneous siderophile element pattern with Ir more depleted than Au and Ni (magnetic fractions – Ir (1.9 – 6.0 xC1), Au (2.8 – 10.0 xC1), Ni (3.5 -6.7 xC1); nonmagnetic fractions – Ir (0.04 -0.3 xC1), Au (0.05 – 0.5 xC1), Ni (0.04 – 0.3 xC1). REE measurements in Atlanta show that all fractions with negative and positive Eu-anomalies are deficient in light REE [Lu (A) / Lu(C1)] / [La(A)/La(C1)] mean = 3 (magnetic) and 1.7 (nonmagnetic fractions). Neither the Eu anomaly nor the light REE depletion can readily explained by nebular condensation at least in solar gas (5). Atlanta was examined by Keil, Rubin (6,7,8) and were found to be free of oldhamite. Perhaps the positive and the negative Eu-anomalies in grain-sized fractions REE patterns are associated with plagioclase. Maximum metamorphic temperatures can be established by the enstatite – silica – albite (ESA) eutectic. If all three of these phases are present and if metamorphic temperatures exceeded the eutectic value, a silicate melt would be formed. Little evidence for such an igneous event - textural or otherwise – has been found in the enstatite chondrites that contain enstatite, silica and albite. While this does not preclude such an event it does suggest that most E-chondrites experienced temperatures less than ESA eutectic. However it is interesting to note that Atlanta (EL6) and Blithfield (EL6) do not contain free silica. An igneous event at the ESA eutectic or even slightly higher that fractionates the melt from the source region would leave an E-chondrite assemblage depleted in free silica. Therefore the absence of free silica in these two meteorites suggests that they may have undergone an igneous event (2).

**Conclusions.** From observed differences of compositions of magnetic and
nonmagnetic fractions it follows that our trace element data accord with idea that Atlanta EL6 reflect main process – sulfurization of metal in protoplanetary nebula and, perhaps, that it may have undergone an igneous event.

Table 1. The average element enrichment factors of magnetic fractions of Atlanta enstatite meteorite.

<table>
<thead>
<tr>
<th>Fractions (µm)</th>
<th>Na</th>
<th>Ca</th>
<th>Sc</th>
<th>Cr</th>
<th>Fe</th>
<th>Co</th>
<th>Ni</th>
<th>Zn</th>
<th>Se</th>
<th>Br</th>
<th>La</th>
<th>Sm</th>
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<th>Au</th>
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<tbody>
<tr>
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<td>0.09</td>
<td>0.1</td>
<td>2.3</td>
<td>2.9</td>
<td>6.1</td>
<td>&lt;0.03</td>
<td>&lt;0.3</td>
<td>0.06</td>
<td>&gt;0.4</td>
<td>0.5</td>
<td>&gt;0.6</td>
<td>&lt;0.8</td>
<td>1.5</td>
<td>&lt;1.0</td>
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<td>2.8</td>
</tr>
<tr>
<td>45&lt;d&lt;71</td>
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<td>&lt;0.2</td>
<td>0.002</td>
<td>0.1</td>
<td>4.3</td>
<td>5.8</td>
<td>6.6</td>
<td>&lt;0.3</td>
<td>&lt;0.3</td>
<td>0.3</td>
<td>&lt;0.4</td>
<td>&lt;0.8</td>
<td>0.8</td>
<td>&lt;1.2</td>
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<td>71&lt;d&lt;100</td>
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<td>0.06</td>
<td>0.06</td>
<td>4.6</td>
<td>7.0</td>
<td>6.7</td>
<td>0.08</td>
<td>&lt;0.1</td>
<td>0.5</td>
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<td>0.7</td>
<td>6.0</td>
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<td>0.05</td>
<td>4.2</td>
<td>6.3</td>
<td>4.8</td>
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<td>7.9</td>
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<tr>
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<td>0.3</td>
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<td>3.8</td>
<td>5.8</td>
<td>4.6</td>
<td>0.1</td>
<td>0.08</td>
<td>0.3</td>
<td>0.2</td>
<td>&lt;0.2</td>
<td>&lt;0.8</td>
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<td>1.2</td>
<td>4.4</td>
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<tr>
<td>260&lt;d&lt;360</td>
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<td>0.8</td>
<td>1.2</td>
<td>0.6</td>
<td>2.8</td>
<td>4.3</td>
<td>3.5</td>
<td>&lt;0.2</td>
<td>&lt;0.5</td>
<td>&lt;0.8</td>
<td>&lt;0.7</td>
<td>0.9</td>
<td>&lt;0.8</td>
<td>1.0</td>
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<td>3.0</td>
<td>5.2</td>
<td></td>
</tr>
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Table 2. The average element enrichment factors of nonmagnetic fractions of Atlanta enstatite meteorite.

<table>
<thead>
<tr>
<th>Fractions (µm)</th>
<th>Na</th>
<th>Ca</th>
<th>Sc</th>
<th>Cr</th>
<th>Fe</th>
<th>Co</th>
<th>Ni</th>
<th>Zn</th>
<th>Se</th>
<th>Br</th>
<th>La</th>
<th>Sm</th>
<th>Eu</th>
<th>Tb</th>
<th>Yb</th>
<th>Lu</th>
<th>Ir</th>
<th>Au</th>
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</thead>
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<tr>
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<td>0.2</td>
<td>0.3</td>
<td>0.03</td>
<td>2.9</td>
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<td>2.1</td>
<td>2.2</td>
<td>2.0</td>
<td>0.04</td>
<td>0.05</td>
<td></td>
</tr>
<tr>
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<td>2.3</td>
<td>1.9</td>
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<td>0.2</td>
<td>&lt;0.02</td>
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<td>0.6</td>
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<td>0.9</td>
<td>1.1</td>
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<td>0.2</td>
</tr>
<tr>
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<td>0.8</td>
<td>2.6</td>
<td>1.0</td>
<td>0.3</td>
<td>0.2</td>
<td>0.2</td>
<td>&lt;0.02</td>
<td>0.4</td>
<td>0.02</td>
<td>0.8</td>
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<td>1.0</td>
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<td>0.2</td>
</tr>
<tr>
<td>100&lt;d&lt;160</td>
<td>0.9</td>
<td>1.8</td>
<td>2.4</td>
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<td>0.3</td>
<td>0.2</td>
<td>0.2</td>
<td>0.03</td>
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<td>0.6</td>
<td>0.7</td>
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<td>1.3</td>
<td>0.2</td>
<td>0.2</td>
</tr>
<tr>
<td>160&lt;d&lt;260</td>
<td>0.8</td>
<td>1.4</td>
<td>2.7</td>
<td>2.7</td>
<td>0.7</td>
<td>0.3</td>
<td>0.4</td>
<td>0.1</td>
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<td>1.5</td>
<td>1.3</td>
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<td>1.5</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>260&lt;d&lt;360</td>
<td>1.4</td>
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<td>1.9</td>
<td>2.6</td>
<td>0.7</td>
<td>0.3</td>
<td>0.4</td>
<td>0.05</td>
<td>1.0</td>
<td>0.1</td>
<td>0.6</td>
<td>1.0</td>
<td>0.6</td>
<td>0.9</td>
<td>0.08</td>
<td>0.9</td>
<td>0.3</td>
<td>0.5</td>
</tr>
<tr>
<td>Enstatite (yellow)</td>
<td>0.6</td>
<td>0.2</td>
<td>2.4</td>
<td>0.06</td>
<td>0.05</td>
<td>0.1</td>
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<td>&lt;0.5</td>
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<td>0.1</td>
<td>&lt;0.2</td>
<td>&lt;0.6</td>
<td>&lt;0.7</td>
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<td>0.7</td>
<td>&lt;0.02</td>
<td>&lt;0.02</td>
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</tr>
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ON A TIME SPAN OF ASTEROID – RUBBLE PILE (ARP) CONSOLIDATION AND A REASON OF LOW DENSITY OF SUCH ASTEROIDS. G. A. Leikin and A.N. Sanovich. Sternberg State Astronomical Institute, Moscow, State University, 119992, Moscow, Universitetskij prosp. 13, Russia, E-mail: san@sai.msu.ru

Earlier [1, 2] we have shown, that ARP must lost the fragments having velocities exceeding \( \sim 10 \text{ m/s in } 10^7 - 10^8 \text{ s} \). Here we investigate with a simple model the time span of consolidation of asteroid fragments produced by dissipation of kinetic energy of fragments in collision. We estimate frequency of collision by the method of free path. The estimate is based on very simple assumptions about form and velocity of asteroid fragments. The dissipation of kinetic energy in double collisions is estimated in simplest dynamic assumption. The evolution of the fragments’ separation is followed, and the time span of ARP transformation to low density object, which seems to observer as consolidated one, is estimated.

Evidence for Remnants of Late Hesperian Ice-Rich Deposits in the Mangala Valles Outflow Channel.  
Joseph S. Levy 1, James W. Head 1, David R. Marchant 2, and Mikhail Kreslavsky 1; 1 Dept. Geological Sciences, Brown University, Providence RI, 02912, 2 Department of Earth Sciences, Boston University, Boston, MA 02215.

Introduction: New, high-resolution images from MGS and Odyssey reveal an unusual unit on the floor of the Mangala Valles outflow channel. In contrast to abundant terrain showing scour, intense erosion, and hydrodynamic shaping typical of the floors and margins of Mangala Valles and other outflow channels (1), this unit is smooth-surfaced, has arcuate and cuspat margins, and has a host of unusual surface features including round pits and ring structures often containing huge angular blocks. We assess several possible origins for this unit and the associated features, and conclude that the most plausible explanation is an ice-rich remnant created by a combination of ponding and ice-cover deflation during the waning stages of the outflow channel flood emplacement.

Description of the Smooth Unit: The smooth unit stretches from the head of Mangala Valles (-18.1 N, 210.6 E) to points at least 400 km north (-11.7 N, 208.8 E), varying in width from up to 10 km to less than 400 m. MOLA altimetry demonstrates that the smooth unit is found predominantly in the lowest regions of the channel, however, in several locations the unit drapes the walls of the channel. The edges of the unit are undulatory and rounded in some locations, and scalloped in others, forming arcuate serrations. Where the unit meets the adjacent walls of the channel, or where there is a contact between the smooth unit and the scoured unit it is superimposed upon, the relatively level and uniform surface of the smooth unit is beveled, yielding a minimum thickness of 10-15 m. Without speculating on the existence of exotic underlying topography, there is no evidence to suggest that the unit is substantially thicker. Towards the proximal end of the channel, the unit is composed of two layered sub-units, displaying similar morphology but different spatial extents. In several locations the width of the unit decreases towards the distal end of the channel, developing into a tightly braided system several hundred meters wide.

Description of Surface Features: The most striking features observed on the smooth unit are abundant, shallow, round pits. The pits range in diameter from 100-600 m, with a mean of 230 m and standard deviation of 60 m. Large, angular blocks of rock are found at the center of many pits near the proximal end of the channel. Most of the pits with blocks have a wide, marginal raised rim, however, none of the empty pits have raised rims. Pits with blocks tend to be larger than empty pits; however the largest pits are empty (Figure 1 and 3).

The pits are predominantly circular; however, clusters of pits show signs of coalescence, forming elongated rectangular depressions with rounded short sides and long axes sub-parallel with the channel. There is no spatial preference for pitting density.

Several channels < 100 m in width cut the smooth deposit at its margin. These meandering channels are all less than 6 km in length (Figure 2). Similarly, two incised features resembling small channels trace circuitous paths at the head of the channel (Figure 3).

Both the smooth unit and the surrounding scoured terrain are impact-cratered. Craters range in diameter from 20-560 m. Crater counts between the two surfaces are generally similar, with the scoured terrain appearing slightly older than the superimposed smooth unit. Both surfaces date to between 100My and 1Gy on the basis of crater-count [8].

Discussion and Interpretation: The smooth unit is an ice-rich residue created by a combination of ponding of flood water and ice-cover deflation during the waning stages of the outflow channel flood event. The presence of the unit in the lowest reaches of the channel supports a ponding hypothesis (2), while the draping of the unit over scoured terrain suggests deflation of an ice cover, as would have formed atop an outflow flood under cold, dry conditions (3). In the early stage, the ponded material and ice-cover

Figure 1. Smooth unit (left) encountering adjacent scoured wall unit (right). Singular and coalesced pits are both visible. From THEMIS image V01454001.
froze solid and sublimated, leaving the deposit as a sublimation residue. Locations showing two sub-units of smooth terrain may be areas where an ice-rich cap settled over a debris rich pond, leading to a more heavily degraded ice-rich upper unit, and a less degraded debris-rich lower unit (4).

The lack of elongated craters or streamlines argues against a flowing-ice origin for the unit.

The size-frequency distribution of the pits rules out the possibility that the pits are caused primarily by degradation of impact craters on an ice-rich surface. Rather, the pits in the smooth unit are interpreted as thermokarst-like features, similar to alases: terrestrial thermokarst pits (5). Whereas terrestrial alases form through a combination of melt, flow, and evaporative processes, resulting in the lowering of ice-rich terrain, the martian features are interpreted as sublimation pits. Both terrestrial alases and martian sublimation pits form when the thermal equilibrium of an ice-rich unit is disrupted, leading to enhanced, localized melting or sublimation (4).

The scalloped or serrated texture of the smooth unit margin is explained as localities where sublimation pits formed near the margin of the unit. Longer embayments may be locations where several sublimation pits coalesced, forming an elongated cavity.

In the distal ends of the channel, the presence of a raised rim around a depressed feature was used as the primary criterion for differentiating between impact craters and sublimation pits. If some craters in the smooth unit have relaxed such that the rim has been lost, they may have been counted as pits, rather than craters, resulting in an artificially young crater age compared to adjacent units (6).

**Figure 2.** Channel at the margin of the smooth unit (left). From THEMIS V044000003.

**Figure 3.** Pits with large blocks, and channel-like features at the head of Mangala Valles. From THEMIS V04762003.

**Conclusions:** The identification of ice-rich residue associated with an outflow flood event strongly suggests that this outflow event occurred on a cold and dry Mars. Further, the preservation of ice-rich debris beneath a sublimation till suggests that climate conditions have remained cold and dry on Mars for at least the past 0.1-1 Gy, and that flood residue can be maintained for previously unprecedented periods of time. The formation of extensive sublimation pits strongly suggests that parts of the remaining deposit are currently ice-rich, making this deposit a primary target in exobiological and ancient climate investigations.

Abstract. More than 2 years ago 2001 Mars Odyssey spacecraft was inserted into a circular orbit around Mars and global mapping of planet in neutrons (HEND, NS), gamma-rays (GRS), visible and infrared bands (THEMIS) was started. It was first attempt to use neutron spectroscopy of Mars to get new information about structure and composition of planet’s subsurface. In this study we pay attention to the distribution of water in polar and equatorial regions of Mars, estimation of mass and density of CO$_2$ frost, annual variations in Mars’s seasonal cycle. All these results are based on deconvolution of observational data of orbital neutron flux registered by Russian High Energy Neutron Spectrometer (HEND).

**Observations.** The High Energy Neutron Detector (HEND) is part of the Gamma-Ray Spectrometer (GRS) on the Mars Odyssey Mission. HEND has three $^3$He proportional counters for measuring epithermal neutrons in broad energy range from 0.4 eV up to 100 keV and organic scintillator for measuring fast neutrons with energy more then 1 MeV [1,2]. Basically the whole strategy of HEND observations may be divided into two independents ways: observation of summer surface (CO$_2$ frost free) and observation of seasonal changes caused by redistribution of atmospheric CO$_2$ between martian poles.

**Data analysis.** The analysis of the summer data allows to deconvolve distribution of water ice and chemically bounded water in martian subsurface. To do it we have model dependent technique based on numerical simulation of orbital flux and suggestion the model of martian regolith. In this analysis we applied two different structure models of martian soil: homogeneous model and double layered model. According the first one the water ice/chemically bounded water is homogeneously distributed through the subsurface. The single parameter of the model is a varied content of water. The second regolith model consists of the water ice layer which is placed beneath dry layer with small content of water. This model is described by two free parameters (thickness of dry layer and content of water in the bottom layer). The special fitting procedure was created and tested to find best correspondence between real data and model parameters. It was found that northern polar regions may be described by homogeneous model [3]. On Fig 1 we present distribution of water ice in this region. On the contrary, HEND data processing showed that southern polar regions are not compatible with homogeneous model and requires the double layered model of regolith [3]. On Fig 2 and Fig 3 we present distribution of water ice in these regions.

In addition to water ice rich polar territories HEND observations also revealed equatorial regions (Arabia Terra and Memnonia) with significant abundance of water. The estimation of water content in Arabia shows that its average percentage may be equal to ~10% by weight. But there are particular moisture spots where content of water may be found as high as 16% (see Fig 4). It is close or even higher than upper limit of chemically bounded water extracted from geochemistry. So the question what we see at Arabia Terra - chemically bounded water or remnants of water ice is still open.

It was also found that neutron flux depends on the seasonal changes of martian climate. The neutron flux (registered above the polar regions of planet) increases up to 10 times during the polar winter and drops back at the summer time. The model processing of HEND data allows to estimate column density ($g/cm^2$) of CO$_2$ frost for given seasonal interval and create 4-dimensional model of seasonal deposit. By help this model we may follow the dynamics of CO$_2$ thickness at different parts of seasonal caps [4,5]. The estimation of CO$_2$ frost column density is a direct way to get mass of seasonal deposit. It may be estimated as production of column density $C_d$ g/cm$^2$ to surface area $S$, cm$^2$ covered by CO$_2$ frost. On Fig 5 we present results of such estimation.
Fig 4. The estimations of water content (mass %, digits shown by black color) and surficial density of the upper dry layer (g/cm², digits shown by red color) for selected regions (black frames). These estimations are drawn above a map of epithermal neutron flux in Arabia Terra. The blue and cyan pixels corresponds to regions with significant abundance of subsurface water (>7-8%). The yellow and green pixels corresponds to relative dry regions (<6%).

Fig 5. The estimations of total condensed mass by General Circulation Model (red line) and by HEND data processing (blue line).

Fig 6. The estimations of volume density from comparative analysis between HEND and MOLA data.

Fig 7. Seasonal variations of neutron flux observed in northern hemisphere for two successive martian years.

Fig 8. Seasonal variations of neutron flux observed in northern hemisphere for two successive martian years.

Another important result which should be evaluated from 4-dimensional model of CO₂ frost concerns the estimation of volume density (g/cm³) of seasonal deposit. The comparative analysis between HEND CO₂ column density and linear thickness of CO₂ frost estimated by MOLA/MGS [6] give us possibility to calculate the volume density of CO₂ deposit. The first step in this direction has already done and preliminary estimations of CO₂ density at northern polar latitudes are found (see Fig 6).

The continuous observation of Mars in HEND experiment lasts more than 1 martian year that allows searching annual variations of martian seasonal cycle. The first results of such study presented at figures 7,8. It is seen that northern seasonal curves repeat the seasonal variations founded for previous martian year. On south the situation is a bit different and significant differences between seasonal curves obtained for two successive martian year are observed.

MASS WASTING EVENTS IN COPRATES CHASMA POSTDATING THE TECTONIC SCARP – EXAMPLES STUDIED FROM HRSC DATA. K. Luiro¹, J. Raitala¹, E. Hauber², G. Neukum² and the HRSC Co-Investigator Science Team. ¹Astronomy, Department of Physical Sciences, University of Oulu, PO BOX 3000, FIN-90014, Finland, kluiro@paju.oulu.fi; ²Institute of Planetary Research, DLR, Berlin, Germany; ³Institute of Geosciences, Department of Earth Sciences, Freie Universitaet Berlin, Germany.

Introduction: Coprates Chasma is part of the large Valles Marineris canyon system. A striking feature of Coprates is the linear scarp at the floor joint, cutting most spur and gully systems on the wall, thus leaving triangular facets. This scarp, prominent on the northern wall of Coprates Chasma, is usually categorized as a tectonic faulting feature [1]. While the scarp is mostly intact throughout the Coprates system, major mass wasting events stretching across the canyon floor were already found from earlier spacecraft images [2]. We looked at the new MEX-HRSC data to see evidence for smaller mass wasting events, too, in the 1 km range that overlap the tectonic scarp. Hence, we would arrange a three level age estimation: events pre-dating the scarp, events associated or contemporating with the scarp, and events postdating the scarp. Furthermore, the presence of a fan-shaped apron would indicate the landslide to be younger in age or more resistant to fluvial or aeolian erosional processes that took place on the canyon floor.

Events classified: We define a small event to be 0.1-2.0 km wide at the scarp level and to reach at least full scarp height downward – a terminal fan is not required, but some were found. The source can be direct upperwall rock debris following the gullies or secondary flows due to leveled terraces (Fig. 2). A large event is a complete wall collapse at least 10 km wide and with visible remnants at canyon floor.

Image resolution: The resolution in the images used varies from 12 to 65 m/pixel. The earlier HRSC orbits have a finer resolution, although various conditions (illumination, possible haze in valleys, map projection) affect the practical use a lot.

An event in detail: Fig. 3 shows a small landslide with a clear fan-shaped apron. The arrows mark the alcoves that do not fit the surrounding spur-gully morphology, thus proposing areas where wall rock met the conditions for collapse [3]. A facet much higher than others cuts the spur above. As often in Coprates, the upper wall gullies stretch way down, and material contribution from upper wall rock is implied, although it may have happened during an extended period after the fan-producing slide. A small crater decorates the

Figure 1. Wall collapses in Coprates Chasma.

Figure 2. Terraced scarp and small landslides.

Earlier data and HRSC Images: Massive landslides in Valles Marineris were found from Viking images [2]. Two such events, visible in HRSC at 292°E and 296°E in Coprates Chasma (Fig. 1), cut the north wall tens of km wide and distributed material across the canyon floor. These events carved deep arcuate re-entrants to the Coprates north wall thus postdating the scarp. With HRSC nadir images, details well below 100 m are observed while maintaining a good context, thus revealing many smaller mass wasting events, and providing a relevant dataset for studying both large and small-scale mass wasting events and for assessing the post tectonic activity.
MASS WASTING IN COPRATES CHASMA

apron, but the possibility to involve this event to nearby crater density countings and thus absolute age estimation [1] is unsettled.

Results: From the six HRSC datasets studied, all show marks of mass wasting overlapping the tectonic scarp. The size of the events varies greatly: massive collapses are visible in Orbit 515 covering the western Coprates and in Orbit 438 over the central Coprates, while most of the ten or so events detected are only 1 to 5 km wide. Only two smaller events in Orbits 438 and 471 show a clear fan-shaped apron. The mixed floor surface as viewed in Orbit 515 (Fig. 4) shows various grabens on the floor and a crater: hence the landslide flowed upon the scarp and the grabens, but was later deformed due to an impact.

Conclusions: Mass wasting studies provide a view to erosional events which have taken place in Coprates Chasma after the last tectonic step – the formation of the north wall scarp. From HRSC images plenty of mass wasting events postdating the scarp are observed – according to Schultz [3] these events could be triggered by seismic activity in the trough-bounding faults, and thus these events would be associated with the crustal stress in Valles Marineris area. Further erosional processes affecting the wasted material are evident – only a few aprons are seen in addition to the two massive wall collapses.

Acknowledgements: We gratefully acknowledge the efforts made by the MEX-HRSC Photogrammetry Team in processing the digital image data.

CONTENTS OF WATER IN TEKTITIC AND IMPACTITIC GLASSES (NEW DATA AND REVIEW)
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Introduction. Already the first determinations of water in tektitic glasses (Fridman, 1962; Gilchrist, 1967) showed that H₂O content in these glasses is very low (<0.03 wt.%) and significantly less, than one in volcanic glasses including obsidians (> 0.1 wt.%) that have a composition the most similar to tektites. The very low water concentration became one of important tektite signs and an argument for their extraterrestrial origin (O’Keafe, 1976). However at present time a majority of investigators arrived at a conclusion that tektites were formed as result of the terrestrial impact processes (see e.g. Koeber, 1986; Feldman, 1990; Heinen, 1998). But by contrast with impactites their spatial and genetic connection with impact craters in most cases is not evident. Essential progress in H₂O investigations in tektites was achieved due to use of local IR spectroscopy. This communication based on new FTIR measurements of H₂O concentration in tektitic and impactitic glasses and also on the review of available data is aimed to consider the following problems: 1) the variations of H₂O content in different types of tektites and impactites, 2) possible accuracy of H₂O measurements in tektitic glasses; 3) the influence of tektitic and impactitic glasses alteration during their stay in the earth crust for a long time on water content in them. Furthermore several features of H₂O behavior in impact processes using available physico-chemical and experimental data are discussed.

Results and discussion. Water contents in two moldavites (Koroseki, Chech Rep.), indoshinite (Vietnam) and irigizite - similar to tektitic glass from impact crater Zhamanschin (Kazakhstan) were determined by FTIR spectroscopy technique.

Water concentration in moldavite is 0.0095 - 0.0111 wt.% and agrees well with earlier determinations. The average value for all moldavite samples of 0.010±0.005 wt.% is close to our determinations (fig.1.). The measured concentration of H₂O in indoshinite sample 0.0038±0.0005 wt.% corresponds to the lowest concentrations in this tektite group 0.0047-0.031 wt.% and is significantly less than average value for indoshinites from various areas of Indo-China 0.0165±0.0009 wt.% (fig.2). The irigizite glass has the highest H₂O concentration of all investigated samples and comes to 0.0348±0.0008 wt.. This fact agrees with previous determinations that show the variation within range of 0.017-0.051 wt.% and the average value of 0.0322±0.0020 wt.% (fig.1). In general available data demonstrate a wide variations of water contents in tektitic and impactitic glasses. Impactites have significantly higher water contents as a rule in comparison with tektites. The average H₂O content in impactites (0.054 wt.%) in three times more then the average H₂O content in tektities (0.015 wt.%). However there is wide enough range of overlapping of water concentrations in tektitic and impactitic glasses (fig. 2).
density. The errors of water determination in tektite glass is usually < 10 relative %. However error value may reach 20-25 % in case of very low H₂O concentrations in tektitic glasses. One of unsolved problems is the calibration of the intensity of ~ 3550 cm⁻¹ absorption band within water concentrations range typical for tektites. The errors of H₂O determinations in various works are similar because the authors use a like techniques that gives an opportunity to compare the data presented in various works.

The solubility rate of glasses in hydrous solution is faster than the diffusion rate of H₂O in these glasses at relatively low temperatures characteristic for nearsurface layers of the earth crust. This fact allows to conclude that in the most cases tektitic glasses were not undergone the hydration during their stay in the earth crust for a long time (Glass, 1984; La Marche et al., 1984; Barkatta et al., 1984; Mazer et al., 1992; Glass et al., 1997; and others). At the same time the glass interaction with hydrous solutions at elevated temperatures can cause a significant hydration and their essential alteration. These alterations are frequently observed in impactites that originally consisted completely or partially of glass. The reverse zonal distribution of water in tektitic samples (decrease of H₂O content to sample edges of the sample) apparently evidence the absence of tektitic glass hydration as result of their interaction with atmosphere on the fly stage after impact events.

Experiments conducted in order to understand the behavior of water and other volatile components in the powerful impact events (fast heating up to very high temperatures 2000-2500°C, melting and evaporation of silicates in vacuum and others) appears to be not adequate model of impact processes. Evaporation of the matter and its subsequent condensation are undoubtedly the most effective processes of differentiation concerning volatile and other components. At so high temperatures that are necessary to provide complete melting and evaporation of silicate substance (>2500-3000 °C), the water in vapor phase should exist mainly in the form of H, H₂, O₂, O, OH, particles, but not in molecular form of H₂O [1]. It is possible that part of these particles altogether with petrogenetic elements may belong to more complicated groups (clusters) formed at evaporation of silicate substances. In the process of vapor cooling and condensation volatile components connected with large-scale clusters are included in silicate liquid first of all. Water and other volatile components in the form of more simple particles may be also partially captured from vapor phase during its condensation. It is possible that the formation of small silicate glass spheres (beards) with such a low concentration of water as for tektites (~ 0.01 wt.%) after test nuclear explosions [2] is more adequate model of volatiles behavior during formation of tektites as result of powerful impact events.

Conclusions. The most part of tektites are natural glasses of terrestrial origin with the lowest concentrations of water that basically reflect initial water contents in melts forming during impact explosions. The impactitic glasses preserved without alteration have higher water content as a rule in comparison with tektites. The H₂O concentration in impactitic glass varies within considerably wider range than that in tektites. At the same time there is wide enough range of overlapping of water concentrations in tektitic and impactitic glasses. The processes of evaporation and condensation of silicate liquid has the most influence on behavior of water and other components during impact events at extreme energetic effect on target rocks.

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Martian volcanic area Olimp and big crater field on different planet’s sides.
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Scheme-map Fig 1(60 N-60 S) displays geosyncline (GS) fold zones and ocean ridges (solid lines) of the Earth. These ridges are GS fold zones with collided fronts with basalt final covers in there rears. Earthen structural lines have axial global symmetry (author’s books 1978,1983, 1986, 1993 in [1].
www.gpi.ru/~mkrn/lpsr). Different ages and modern state of Earthen seams is evident [1].

On Fig.1 dotted lines denote structural seams and young volcanic areas from Martian map. Earthen 0°=Martian 60° W [2, 3].
Border of Noctis labirint and seams Mariner near equator (90W Mars) have their images on the other hemisphere, as S border of Libia mounts. Solis Planum (~ 90W) with borders SW- Fossae Claritas and SE – Mellas, Felis Dorsa seams has as its twin Terra Tyrrhena (~270W) with its borders: SE - seam Harmahys and SW- margin of Hellas. Ring Argyr (~40W) has the twin in Cimmeria Terra with its crater Kepler and SE border Eridania et al.

For Martian volcanic areas one can see: Alba Patera (immersed plateau) has the place on Earthen volcanic I/Cr area, rear to Atlantic rise Corner-Miln. On the other hemisphere its place is on the Pg/N volcanic field, rear of laramian Japan arc. Vlc.Olimp with neighbour rings has its place on Earthen Mexican Bay and Appalachian rear with their volcanic post-variscan P/T fields. The place of vlc.Olimp on other Earthen hemisphere (immersed rings on Martian map) is the rear of variscan (hercynian) GS Sikkan-Junnan fold zone, it is P/T Eymeshan basalt field. Earthen images of the three grand Martian volcanoes near equator (small rings with central point) are absent. At their places on the other side of Mars we can see imersed rings only. If to turn the Mars 180 grade, grand three Martian volcanoes coincide with Earthen rises: isl. Sumba, isl. Halmahera, the place of Palau ridge. At the same time Elisium rise on the Earth’s African hemisphere coincides with Near East region. Youngest Earthen volcanic Syrian-Jordan basalt plateau (Ng-Q) is placed nearby, in the rear of young (Pg/Ng) Palmyra fold arc.

The place of vlc.Olimp and the craters on other Martian hemisphere (see grade net, [4]) one can see in other scale and in different projection on Fig.2

Volcanic grand ring-plato (H:L = 1:20), margins of adjacent peripheral basalt plateoes, cracks, big craters and small craters along many cracks are shown by solid lines. All craters (big and small) of the other Mars hemisphere, just as cracks with chains of small craters along them, are shown by dots.

Numbers on Fig.2 denote big craters names on the Mars hemisphere, opposite to the hemisphere with vlc.Olimp. 1- _assini, 2- Tikhonravov, 3- Antoniadi, 4- Flammarion.

It is evident: these big craters circle the region, where there is vlc.Olimp on the other planet side.
It might be that megavolcanic complex of one hemisphere and the system of aforementioned craters of the other hemisphere have developed in different ways. Either from margins to the center (vlc.Olimp planet side), or from the center to the periphery - on the other planet side.

The crater 5- Shoner is also shown on Fig.2. It nearly coincides with the center of vlc.Olimp on the other planet side.

To the South, craters: 6- Janssen, 7- Shroeter, 8- Huygens are shown also. It can be seen, that the 7-th and 8-th craters belong by their place in coordinates nets to N and S periphery of Ares, on the planet other hemisphere.

Mutual matching of circular forms of both hemispheres in axial symmetric Martian regions is evident. This phenomenon directly contradicts the interpretation of Martian craters (at least for abovementioned) as results of random impacts. Without doubt, they are endogenic.

Now pay attention to the following. Craters denoted by numbers, just as other ring and arc forms of the corresponding Martian side, look similarly. Their margins almost every where or on large parts of these arc margins are accompanied by small rings-craters.

The late cross marginal board arcs by their forms, and thus they are relatively younger.

Thus, the craters of this hemisphere in the simplest way may be represented as former volcanoes, which have immersed slowly into hot substratum above deep volcanic zones of Martian entrails. Perhaps, the lavas have never reached the surface (?). Small crater rings most probably fix peripherical eruptions from the same deep zones, i.e. the eruptions of the youngest volcanic products.

**Resume.** The Mars, as the Earth, has the axial structural symmetry of its outer shell. The structures on it are endogenic.

**Ref:**
VENUS GEOLOGIC MAPPING: INSIGHTS INTO CRUSTAL EVOLUTION ON LOCAL, REGIONAL, AND GLOBAL SCALES. S. M. McColley and J. W. Head, III, Dept of Geological Sciences, Brown University, Providence, RI 02912. Shawn_McColley@brown.edu

Introduction and Background: Over the last several years, two views have emerged regarding the large scale evolution of the crust of Venus. A global stratigraphic model is supported by [1] and [2]. This model, defined as directional by [3], finds evidence for the time-dependent evolution of surface feature formation, as well as a traceable progression in the type and style of volcanism that has occurred over the visible geologic past of Venus [1, 2]. Adopting these principles provides the means of cataloging and organizing geologic units into epochs of Venustian history. In response to the global stratigraphic model proposed by [1] and [2], [3] suggested that coronae, rifts, wrinkle ridges, small and large edifices, and large flow fields have formed throughout the geologic history that is recorded on the surface of Venus. They further assert that both tectonic and volcanic features show evidence for protracted histories and possibly reactivation following periods of dormancy, implying a nondirectional evolution for the observable history of Venus.

The surface of Venus is geologically young with a crater retention age of 300-500 Ma [4, 5, 6]. A broad range of geodynamic models have been proposed to explain the apparent youth of the surface. Models include: heat pipe mechanisms [7], hot/cold spot tectonics [8], stagnant lid convection [9], mantle overturn [10], a thin-to-thick lid transition [11], shift from plate tectonics to a one-plate planet [12], and several other derivations. Geodynamic models carry implications for the way in which a body will evolve both from an internal and a surface point of view. These implications are governed by assumptions regarding heat loss. Heat loss predominantly dictates the type of features (tectonic, volcanic, etc.) that are observed on the surface. Thus, in terms of the global stratigraphic model (directional) and the nondirectional model for the observable history of Venus, what type of features might we observe if Venus has lost heat monotonically, or in contrast, in a more step-like fashion?

Methods: In an attempt to constrain better the evolution (at a variety of scales) of Venus, we have undertaken several mapping projects ranging from the regional-scale (the Lada Terra (V-56) quadrangle) down to the detailed mapping of individual volcanic deposits (festoons). Lada Terra quadrangle (Figure 1a) lies between 50-75°S and 0-60°E. The quadrangle contains a sufficient representation of currently identified Venustian geologic units, such that we feel it and surrounding quadrangles may hold significant clues to the overall evolution of Venus. We are utilizing all available Magellan data (C1MIDR, C2MIDR, FMIDR, GXDR, and gravity) to identify local and regional variations in geology and structure, as well as styles of volcanism.
Figure 1. a) The Lada Terra (V-56) quadrangle (50-75°S, 0-60°E). The white box indicates the location of the sketch map shown at the right. b) Sketch map showing inferred cross-cutting and embayment relationships (white = crater material, intermediate grey = lobate plains, light grey = densely fractured plains, black = plains with wrinkle ridges, dark grey = tessera, shield fields lie just to the south of the sketch map).


Figure 2. a) Festoon deposit superposed on the tessera of Ovda Regio (6.5°S, 95.5°E). b) Sketch-map showing the seven subunits that comprise the festoon (map scale is equal to that of c). c) Sketchmap showing a distribution (about every fifth ridge) of compressional ridges used to infer flow direction and possible source areas.
EVIDENCE FOR INTERNAL DEFORMATION AND FLOW IN THE NORTHERN POLAR CAP OF MARS: PART 1, BACKGROUND.  S. M. Milkovich and J. W. Head, III, Dept of Geological Sciences, Brown University, Providence, RI 02912.  Sarah_Milkovich@brown.edu

Introduction: Within the northern residual polar cap of Mars are dark lanes or troughs; on the walls of these exposures are layered deposits. These deposits consist of laterally extensive layers of ice and dust and are found throughout the polar cap. They were first identified in Mariner 9 images [1, 2] and later studied in detail with Viking orbiter data [e.g. 3, 4, 5, 6, 7] and with the Mars Orbiter Camera (MOC) data from the Mars Global Surveyor (MGS) [e.g., 8, 9, 10, 11]. The polar layered deposits are thought to contain the record of recent climate change and polar history [e.g., 5, 7, 10, 11]. An understanding of the processes at work shaping the polar cap can help to interpret this climate record.

It is well-documented that terrestrial ice sheets of sufficient thickness can flow [e.g., 12, 13, 14]; if the martian northern polar cap is thick enough and the geothermal gradient at the base of the cap is high enough, it should flow as well [15, 16, 17, 18, 19, 20]. However, the strength of the cap material is not known due to the unknown proportion of dust in the polar layered deposits internal to the cap; the presence of dust can strengthen or weaken the cap material, depending on the amount of dust [21, 22]. The presence of layers deformed by flow would provide insight into the strength and behavior of the northern cap as well as the conditions and style of deformation.

In this abstract, we review the current efforts to model flow of the northern cap and outline predictions of the resulting geomorphology of the polar layered deposits. In a companion abstract, we examine the evidence for deformation and flow of the polar layered deposits.

Ice Flow Models: In a simple symmetric terrestrial ice cap in temporal steady state, accumulation of ice in the center of the cap and ablation of ice at the edges are constant. Mass balance requires that ice flows from the accumulation zone to the ablation zone. Ice and other material deposited in the accumulation zone moves downward and is thinned by compression and shear flow (Figure 1). Thus, layers internal to the ice cap (for instance, dust or tephra) also get thinner and experience more shear as they get closer to the ice bed. In addition, the ice towards the very base of an ice sheet must flow over the ground surface that underlies the ice sheet; the topographic relief of this surface may distort the layers [summarized in 16].

A major difference between the simple ice cap described above and the martian north polar cap is the presence of spiraling troughs cutting into the martian cap, exposing layers of ice and dust. The walls of these troughs are much darker than the ice-covered polar flats between troughs, which has led to the theory that ablation in the form of sublimation occurs on the equator-facing slopes while deposition occurs on the bright flat areas [4, 5, 7, 16, 17]. This process may cause the troughs to migrate poleward in a conveyor-like fashion [7, 16, 17, 18]. Thus, the troughs may be moving inward on top of the ice which is moving outward [16].

Figure 1. Behavior of material in a simple domed ice cap. Developed from an actual cross section of Greenland. From [16].

Figure 2. Fisher’s model of cap flow. A) Proposed particle paths through migrating scarps. As the scarp migrates, some sublimated water vapor moves to the polar flats between troughs. Some layers have originated near troughs farther north and moved down from flow. B) Schematic cross section through the cap. A given trough wall is fed by ice laid down on the polar flat immediately above it and by ice from higher up the cap. This results in a temporal discontinuity in the layers exposed in any given trough wall. The troughs lower in elevation tend to have more layers originating from the interior of the cap. From [16].

The behavior of internal layers in such a cap is quite complex and open to several interpretations. Fisher [16] calculated the behavior of internal layers in an ice sheet with idealized trough topography assuming that the ice thickness was much thicker than the trough depth, in effect, that the trough system only involved the very uppermost section of ice cap (Figure 2). In this model, the uppermost layers in the trough walls are young layers formed from material recycled from the retreat of the trough wall; as material is removed from the trough wall, a portion is redeposited on the nearby polar flat and a new layer is formed. The troughs at the lower elevation and outer margin of the cap contain older layers which are
from material deposited towards the center of the cap which has flowed out, similar to the behavior of a simple ice sheet. In effect, there is a simple ice sheet with internal layers underlying a complex ice sheet with topographic layering; layers from the simple sheet are exposed at the edges of the cap and at depth. A discontinuity is expected between layers from different source regions. Fisher [17] expanded on this theory, adding the prediction that the upper layers formed by local recycling of polar volatiles will be more disorganized and non-homogenous than the older, deeper layers. Waves in the layers are expected at depth due to the presence of the troughs; the wavelength and amplitude of these waves are a function of the rate of trough migration compared to the rate of outward ice flow.

Figure 3. Hvidberg’s model of cap flow. Particle paths along a topographic profile of the cap. A) An enlargement of a part of the flow line, and B) the full flow line. The presence of troughs on the surface affects the motion of the cap material all the way to the base; thus, the troughs divide the cap into discrete dynamic units. From [19].

Modeling work by Hvidberg [19] leads to another model of flow in the cap (Figure 3). Hvidberg [19] used a finite element ice flow model to calculate flow rates and trajectories assuming pure ice in the northern cap using the topographic profile of the cap measured by MOLA. She found that rather than allowing ice to flow below the trough system, the troughs separate the cap into distinct units. In this model, there is no flow between the major troughs. This is due to the fact that ice exposed in the pole-facing slopes of the troughs will either not flow at all or flow slowly towards the pole. Indeed, flow alone should smooth the troughs in less than 10^5 years; this implies that other forces such as sublimation or erosion by katabatic winds are actively keeping the troughs open.

Predictions: What kind of evidence should we be looking for to determine where, when, and how much the ice flows, in order to test these different models? Ice flows by creep, or movement within or between individual ice crystals. The rate of ice creep is a function of shear stress and is affected by impurities such as the presence of dust within the ice [20, 21, 22, 23, 24, 25]. Most creep occurs in the lower region of an ice cap or glacier where the shear stress is greatest. When the ice cannot creep fast enough to allow the ice cap or glacier to adjust its shape under stress, the ice can undergo brittle failure or fracture [13, 14, 26, 27].

Extensional zones within the ice can cause individual layers to thin (in the case of a ductile layer) or fracture (in the case of a brittle layer). A brittle layer surrounded by ductile layers will break apart, and the surrounding ductile material will flow around and into the gaps to form boudinage structures.

Compression zones within the ice can cause fold-and-thrust features similar to those found at compressive plate boundaries on the Earth [27]. Folded ductile ice can also be detached from the underlying substrate along a bedding plane known as a décollement. The folded ice can be further deformed by overthrusting reverse faults.

The state of stress (compressional vs extensional) at the base of an ice sheet depends on whether the ice is accelerating or decelerating. This in turn depends on the thickness of the ice, the basal temperature of the ice, and whether the ice is located beneath an accumulation zone or an ablation zone. For a simple domical ice sheet, the interior regions experience extension while the marginal regions undergo compression (Figure 4) [23].

This simple model of ice cap dynamics (Figure 4) offers a framework for assessment of evidence for flow. We should look for evidence of flow in the northern polar cap by searching for deformation in layers near the base of the cap. The troughs which cut the deepest into the cap are located towards the cap margins; based on terrestrial experience [23], we expect that any deformation present will be due to compression.

EVIDENCE FOR INTERNAL DEFORMATION AND FLOW IN THE NORTHERN POLAR CAP OF MARS: PART 2, OBSERVATIONS. S. M. Milkovich and J. W. Head, III, Dept of Geological Sciences, Brown University, Providence, RI 02912. Sarah_Milkovich@brown.edu

Introduction: Within the northern residual polar cap of Mars are dark lines or troughs; on the walls of these exposures are layered deposits. These deposits consist of extensive lateral layers of ice and dust and are found throughout the polar cap. They were first identified in Mariner 9 images [1, 2] and later studied in detail with the Viking orbiters [e. g., 3, 4, 5, 6, 7] and with the Mars Orbiter Camera (MOC) onboard the Mars Global Surveyor (MGS) [e. g., 8, 9, 10, 11]. The polar layered deposits are thought to contain the record of recent climate change and polar history [e. g., 5, 7, 10, 11]. An understanding of the processes at work shaping the polar cap can help to interpret this climate record.

It is well-documented that terrestrial ice sheets of sufficient thickness can flow [e. g., 12, 13, 14]; if the martian northern polar cap is thick enough and the geothermal gradient at the base of the cap is high enough, it should flow as well [15, 16, 17, 18, 19, 20]. However, the strength of the cap material is unknown due to the unknown proportion of dust in the polar layered deposits internal to the cap; the presence of dust can strengthen or weaken the cap material, depending on the amount of dust [21, 22]. The presence of layers deformed by flow would provide insight into the strength and behavior of the northern cap.

In this abstract, we examine the evidence for deformation and flow of the polar layered deposits. In a companion abstract [23], we review the current efforts to model flow of the northern cap and outline predictions of the resulting geomorphology of the polar layered deposits. Based on that analysis, we now look for possible examples of deformation of layers towards the base of troughs at the margin of the polar cap. We predict that we will see features characteristic of compression, such as folds, reverse thrusts and décollement, and evidence for extension, such as boudinage [16, 24, 25, 26], depending on position in the cap (See Figure 4 in [23]).

Evidence for deformation and flow in the Polar Layered Deposits: Previous examinations of high resolution images located around the cap have not found any definite evidence of layer deformation. In fact, the upper 300 m of layered deposits appears to be continuous and undeformed around the cap [27]; this may imply that the upper sequence of layers has not experienced stress conditions required for brittle fracture or has not experienced flow at all. We therefore turn our attention to layers present towards the base of the polar cap where the maximum shear stress should be felt and deformation is likely to occur [25].

An examination of the highest-resolution MOC narrow angle images around the polar cap reveals several layers with characteristics consistent with brittle deformation. Three images located in the same trough near 83.3° N and 94° E contain a layer that appears to have been folded and locally faulted and overthrust (Figure 1a, b, c, d, e, f). The layer, located towards the base of the trough at an elevation of about – 3750 m, is ~ 1 m thick (the shallow slope of the trough makes it appear thicker in the images). The layer does not occur at the very bottom of the trough, and so is unlikely to be sand dunes. Individual segments of the layer range from ~ 90 m to ~ 260 m in length, averaging ~160 m. The edges of most segments are rotated away from the neighboring segments, oriented at an angle to the layer plane; this is consistent with motion along a fault due to shear stress. An additional image located at 83.3° N and 98.3° E (Figure 1g, h) in a neighboring trough also has a layer with brittle deformation features. The segments in this layer average ~120 m length and are also rotated away from the layer plane. An unusual layer is observed in the bottom of a trough located at 84.8° N, 124.1° E (Figure 2); this layer may contain a fault. Unfortunately, only a small portion of this layer is exposed in the image. Nevertheless, there is evidence for thinning of the upper layer, topography on the basal contact, stratigraphic differences across the structure, ad an apparent fault-like ramping. Initial examination of the geometry of these fault-like features indicates that they are oriented so the hanging wall is moving towards the margin of the cap; this is consistent with compression experienced at the margin of a flowing sheet of ice [25].

It is important to determine whether these features are consistent with compression or extension. The deformed layer in Figure 1b in particular appears to contain several segments that have not quite broken apart but are connected and may be folded. This is similar to layer deformation in a compressive stress regime where the layer is folded to such an extreme that portions of the folds undergo faulting and low-angle overfaulting. Therefore, we interpret this layer as a compressive feature. The presence of this layer combined with the probable faulted layer in Figure 2 indicates that compressive forces were at one time present in these troughs.

Discussion: The presence of features consistent with deformation such as faults in the layered deposits is evidence for flow of the ice cap at some point in the cap history. The scarcity of such layers may indicate that most of the layers undergo ductile flow and therefore do not form features such as boudinage or reverse thrusts and folds. Alternatively, lack of common flow features may indicate that ice flow is an infrequent occurrence in the history of the polar cap. The layers identified as having features consistent with brittle deformation are located in troughs near the edge of the cap; this indicates that the layers near the cap margin have experienced some compressive shear stress. Such stress is consistent with our understanding of the behavior of ice flow in terrestrial ice sheets and glaciers; in regions where the ice is thinner, the rate of flow decreases and thus the ice is compressed [25]. Further analysis of the geometry and distribution of these features is underway and will provide more insight into the stress regimes and history of the polar cap.

Figure 2: Deformation in the north polar layered deposits. A) Subframe of M00-00163. B) Sketch of M00-00163. Image has been rotated so the top of the trough is up.

NORTH POLAR CAP OF MARS: MILANKOVITCH CYCLES AND RECENT CLIMATE HISTORY.
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Introduction: The northern polar cap of Mars is characterized by spiraling troughs cutting through the cap surface. Horizontal and subhorizontal layers exposed on the walls of these troughs are thought to contain varying ratios of water ice and dust. These polar layered deposits (PLD), first observed in Mariner 9 images, extend throughout the cap. In Viking images, a thick sequence of PLD is present at each pole and includes exposed sequences of up to 20 regular layers of alternating dark and light material, each layer pair between 10 and 30 meters thick; there are apparent unconformities between some sets of layers. North and south individual polar layers are similar but have some morphologic differences [e.g., 1]. Layers include dust and water ice but at depth within the cap, CO$_2$ clathrate hydrate ice may be present as well [e.g., 2].

Layer formation models based on these images called for obliquity cycles to drive climate change to produce the layers [e.g.3-6]. Such approaches are similar to models in which orbital Milankovich cycles drive ice ages and climate change on the Earth [e.g., 7-10]. Changes in insolation patterns due to quasi-periodic orbital cycles will affect the brightness of an individual layer; the duration and intensity of southern summer influences the occurrence of global dust storms [11] which affects the dust to ice ratio, and in turn the brightness, of the layer deposited at the northern cap. For example, on Mars a light-dark PLD layer pair might be deposited as the planet moved from low obliquity (favoring ice deposition) to high obliquity (favoring dust deposition) [e.g., 3].

Although extensive analysis and discussion resulted from the Viking data, no consensus was reached on the interpretation of many important aspects of the layered terrain and the individual layers. In a key synthesis paper, Thomas et al. [1] summarized what was known about the polar deposits (outlined above) and what was not known. Outstanding issues included the exact composition and ratio of dust to ice of the PLD as a whole, as well as the vertical sequence of individual layers, their correlation, the physical characteristics that cause them, whether they are compositionally distinct from residual frosts, and their relationship with expected climate cycles. The advent of much higher-resolution Mars Global Surveyor (MGS) Mars Orbiting Camera (MOC) images of the polar layered terrain, combined with the Mars Orbiting Laser Altimeter (MOLA) altimetry measurements, has produced a dramatic new data set, equivalent to individual vertical "cores" through the sequence of layers exposed in troughs and cliffs across the polar caps [12]. In these individual "cores", the PLD is shown to contain even more layers than those seen in Viking images, revealing layers with a variety of brightnesses and thicknesses down to the limit of resolution (a few meters) (Figure 1c). Furthermore, comparison of individual images suggested possible correlations of layers between "cores" [12-15]. Thus, the new MOC data set is analogous to an initial oceanographic expedition to a sedimentary basin to undertake a comprehensive collection of deep-sea sediment cores to study the paleoceanography and paleoclimate record on Earth.

The higher-resolution data set raises additional questions concerning the PLD. For individual vertical sections of layers and comparisons between sections, the list of outstanding questions is large and includes the following: What is the scale of individual layers within a single vertical section? Are they laterally continuous within a section? Are they vertically repetitive? Are there any cyclic patterns and if so, what is their nature and origin? Do any such patterns change with depth? What is the temporal meaning of layers and units? Do they correlate with any predicted climate variation, and what does this mean for the polar history of Mars? How far laterally can individual layers and groups of layers be traced? Is there evidence for discontinuities and unconformities? If so, how extensive are they and what do they mean? What is the implication of the dip of the layers for the polar history of Mars? How are regional correlations related to orbital parameters, climate cycles, the origin of troughs, and the general volatile history of Mars?

Characterization of Vertical Sections: As a first step in unraveling the relationship between the polar layered deposits and climate, it is necessary to understand and quantify the characteristics of the layers themselves in individual "cores" or sections. Then it is necessary to establish whether there are any correlations between adjacent sections as well as any regional or cap-wide correlations. Finally, it is important to establish if and how such correlation is consistent with depth in the vertical sections representing changes with geological time. Once the signals encoded in the layers and their lateral and vertical correlations are known, then their relationships to climate change can be assessed. The results reported here are part of an ongoing effort to characterize quantitatively the layers both vertically, by looking for patterns in vertical stratigraphy, and horizontally, by examining variations in layer continuity on local (trough-wide) and regional (cap-wide) scales. The results are then used to assess models of polar history and climate. Similar questions are commonly asked in the study of layered sedimentary sequences on Earth and in the analysis of their relationship to pa-
leoclimate conditions. Thus, we first examined techniques used in the analysis of individual sequences (e.g., sediment cores or rock sections) and their lateral relationships and correlation, and we then applied appropriate techniques to the Mars polar deposits.

**Results:** We use two techniques commonly employed in paleoceanography for the study of deep-sea sediment cores on Earth to establish the characteristics of layers in individual cores (Fourier analysis) and to determine the correlation between cores (curve-shape matching algorithms). Application to "cores" (vertical sections) of the north polar layered terrain on Mars reveals several fundamental properties of north polar cap stratigraphy: 1) Fourier analysis of the layer vertical sequences reveals a characteristic and repetitive wavelength of ~30 m thickness throughout the upper part (Zone 1) of all sequences analyzed. 2) Application of curve-shape matching algorithms demonstrates that layers correlate across at least three-quarters of the cap (~6x10^5 km^2) in the 13 images analyzed to date. 3) Assessment of geometric relationships shows that layers are not horizontal, but rather have an apparent dip of approximately 0.5 degrees. We interpret these results (Fig. 1) as follows: 1) The fundamental ~30 m signal is interpreted as a climate signal that may correspond to a 51 kyr insolation cycle. 2) The lateral correlation and broad distribution of these layer sequences strongly imply that layer accumulation processes are widespread across the cap, rather than confined within a single trough or region. 3) Local to regional variability in individual layer thicknesses (and thus accumulation and sublimation rates) is typically less than a factor of 2.5, providing the ability to study regional trends, but often making it difficult to correlate visually the vertical sequences in individual cores. Finally, initial examination of layers located deeper in the stratigraphic sequence within the north polar cap than the ~300 m thick Zone 1 provides evidence for a unit less than 100 m thick (Zone 2) in which the fundamental ~30 m sequence is not detected. We interpret this as a deposit having formed during a recent high-obliquity phase of Mars, during which time polar volatiles underwent mobilization and were transported equatorward, leaving a polar lag of dust-rich material. The most recent "ice age" (~0.5-2 Ma) offers a plausible candidate for this period of ice cap removal and lag deposit formation. An underlying Zone 3 (~200 m) contains a dominant 35 m signal, and a lowermost Zone 4 (~200 m) contains multiple signals but no dominant one. Together these four zones represent ~800 m of vertical stratigraphic section, about one-fourth of the total thickness of the cap. These findings support earlier interpretations that orbital parameter variations could cause significant erosion and possibly complete removal of the polar caps. The interpreted crater retention ages of

the layered terrain are consistent with the correlations and vertical sequences described here, suggesting that the polar caps wax and wane throughout geological history, depending on the evolution of orbital parameters. Definition of the ~30 m unit signal holds promise for determining 1) the detailed origin of individual layer types, 2) the nature of deposition and sublimation processes and their relation to insolation geometry across the polar cap, and 3) correlation with and comparison to the south polar layered terrain record.


![Figure 1. Composite stratigraphic column for the north polar cap showing the four zones recognized.](image-url)
MARS: RECENT AND EPISODIC VOLCANIC, HYDROTHERMAL, AND GLACIAL ACTIVITY REVEALED BY THE MARS EXPRESS HIGH RESOLUTION STEREO CAMERA (HRSC) EXPERIMENT.  G. Neukum, R. Jaumann, H. Hoffmann, E. Hauber, J. W. Head, A. T. Basilevsky, B. A. Ivanov, S. C. Werner, S. van Gasselt, J. B. Murray, T. McCord, and the HRSC Co-Investigator Team, 1 Institut fuer Geologische Wissenschaften, Freie Universitaet Berlin, Malteserstr. 74-100, Bldg. D, 12249 Berlin, Germany. 2 DLR-Berlin, Germany (swerner@zedat.fu-berlin.de). 3 Brown University, Providence, RI 02912, USA. 4 Vernadsky Institute of Geochemistry and Analytical Chemistry, RAS, 119991 Moscow, Russia. 5 IDG-RAS Moscow, Russia. 6 Open University, Milton Keynes MK7 6AA, U.K. 7 University of Hawaii, Honolulu, Hawaii 96822, USA.

Introduction: On board the ESA Mars Express Orbiter, the High Resolution Stereo Camera, a multiple line scanner instrument, is acquiring high-resolution colour and stereo images of the surface of Mars[1]. Resolution down to 10 meters per pixel coupled with large areal extent (swaths typically 65-100 km wide and thousands of km long) means that small details can be placed in a much broader context than was previously possible. Among the major objectives of the experiment is an assessment of the level of recent geological activity on Mars, particularly the type of volcanic and climate-related deposits that might indicate areas of hydrothermal activity and recent water exchange conducive to exobiological activity.

Data: We have made use of the new HRSC images and their particular qualities in mapping out terrain-types for the interpretation of morphological features and topographic relationships from the 3-D data and high-resolution imagery including the Super Resolution Channel (SRC) data (resolution down to 2.5 metres per pixel) [1]. The high-resolution colour data was very useful for distinguishing different materials. The combined use of the HRSC data and nested MOC [2] or SRC imagery has proven to be extraordinarily helpful in the interpretation of morphologies and processes which shaped the landforms. The main emphasis in this paper was put on understanding the time-stratigraphic relationships and the sequence of events in order to understand the geologic evolution of the Martian areas investigated. Time sequences were obtained by determining the number of superimposed impact craters and deriving absolute ages.

Results: The HRSC Experiment on the ESA Mars Express Mission has obtained new evidence for recent and episodic volcanic resurfacing activity on Mars, revealing an unusually geologically robust and recently active planet. Calderas on five major volcanoes in the Tharsis and Elysium regions show repeated activation and resurfacing during the last 20% of Martian history, with caldera floors as young as 100 Ma, and flank eruptions as young as 2 Ma. These results confirm that the edifices are constructed over billions of years [3] and are characterised by episodically repeated phases of activity[4] continuing almost to the present and suggesting the volcanoes are potentially still active today. It appears that the more recent volcanic activity on both the Tharsis and Elysium volcanoes clustered around 100-200 Ma ago, practically coinciding with radiometric ages of several Martian meteorites[5]. Glacial deposits at the base of the Olympus Mons escarpment[6,7,8] show evidence for repeated phases of activity over the last 5% of Martian history, with the latest phase occurring as recently as ~4 Ma ago. Bright deposits on the flanks of Olympus Mons, on the top of the scarp and on high-standing plateaus at the edge of the western scarp are interpreted to be remnants of ice and dust accumulations dating from these times and even earlier periods as old as 3.8 Ga ago. Morphological evidence is found that snow/ice deposition on the Olympus construct at elevations more than 7000 m high led to episode(s) of glacial activity at this height. The data suggest that water ice protected by an insulating layer of dust may now be present at high altitudes at the edge of the Olympus Mons shield. The presence of the young glacial deposits in the tropics of Mars at the base of the Olympus Mons escarpment supports the hypothesis that the obliquity of Mars[9]may have recently been in excess of 30°, during which times snow and ice accumulations are predicted in these regions[10,11]. These new results confirm that Mars is fundamentally important exploration target, combining a preserved ancient record with accessible deposits representing recent volcanism and the geological record of recent glaciation.

THE REINER GAMMA SWIRL AS CHARACTERIZED WITH EARTH-BASED CCD PHOTOMETRY.
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Introduction: The Reinerγ Formation (RGF) is the best example of swirls on the lunar surface. Swirls are albedo structures that are not manifested themselves in topography. The RGF is located in the western portion of the nearside. Swirls are considered to be results of cometary or meteoroid swarm encounters. Infrared data showed that the RGF has somewhat lower thermal inertia, indicating presence of fine-grain regolith. Radar measurements reveal no significant anomaly, exhibiting that the near-surface population of stones in the area resembles that for the average mare regolith layer. The RGF shows a polarimetric anomaly at large phase angles, which indicates either presence of coarse-grained regolith or that the regolith is comparatively dense. The formation is generally considered to have strong forward scatter, as it shows up at large phase angles near terminator, whereas craters with bright halos disappear. Thus the RGF material is characterized by lower slope of the phase function as compared to surrounding mare regions. The purpose of this study is advancing CCD-imaging photometry of the RGF at phase angles when the shadow-hiding effect is the main contributor to the phase angle behavior of lunar surface brightness [1].

Source data: New observations of the Moon were carried out with the telescope Zeiss-600 of Crimean Observatory (Simeiz). A CCD LineScan Camera SONY ILX707 was exploited. The western portion of the lunar disk was scanned with the 2048 pixels line at the wide spectral bands with λeff = 0.65 μm and 0.45 μm. Several scans of the Moon were done at the phase angles near 18°, 39°, 90°, 122°, and 134°. Note that the phase angle 134° is almost the highest one that can be reached in Earth-based observations. The scan data were brought to images using complicated geometric transformations. Then for each image we compensate the brightness trends from the limb to terminator and from the poles to equator. All images were coregistered with an original heuristic algorithm and transformed to the direct orthographic projection. Finally we mapped phase-angle ratios (39°/18°), (90°/39°), (122°/90°), and (134°/122°) that are presented in Figure 1.

Results: The mapped area includes the RGF that is clearly seen as a bright diffuse spot near the frame center. The crater Reiner is located on the right hand. As one can see the RGF (or its portions) shows up almost in all the phase ratio images. The brighter the details on the phase ratio images, the smaller the slope of the phase angle curve of lunar surface brightness. The phase ratio patterns differ from the albedo one. The phase ratio patterns are similar to each other in red and blue light. Different portions of the RGF are characterized with slightly prominent forward scattering in the phase angle range 18° – 90°, which consists with results obtained earlier [1]. This feature result from higher albedo of the RGF material, as high albedo implies multiple scattering that decreases the slope. On the other hand the phase ratios (122°/90°) and (134°/122°) demonstrate higher slopes of phase curves of the RGF than those of surrounding regions. In particular, the external bright area of the RGF demonstrates steeper slope of phase angle curve (and hence more rough surface) than mare vicinity (the same effect is observed for highlands). The described slope differences are relatively small, < 5 – 10 %, but they are reliably detected. Unlike the phase ratios (39°/18°) and (90°/39°), the ratio (122°/90°) and especially the ratio (134°/122°) can be significantly effected by meso-topography on the scale ~ 1 m, that is relatively low surface slopes.

Conclusion: Our results clearly show that on average the RGF in the phase angle range 90° - 134° reveals smaller forward scattering than surroundings. The found effect is due to surface roughness, and it does not compensated with the albedo influence. This indicates that the RGF surface is perhaps more rough on the scale ~ 1 m, than the surrounding mare surface. This conclusion is in agreement with model of meteoroid swarm encounters [2,3].

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Figure 1. Brightness and phase-angle ratios images.
**Introduction:** Ascraeus Mons is the northernmost of the three Tharsis Montes shield volcanoes (Arsia Mons, Pavonis Mons, and Ascraeus Mons), which are aligned along a N40°E trend on the crest of the broad Tharsis Rise. These three volcanoes, by virtue of their great size and apparent similarity to terrestrial shield volcanoes such as Mauna Loa, Hawaii, have attracted considerable attention in the study of the geologic history of Mars. Previous studies have included investigations into their caldera morphologies as an indicator of late-stage summit activity [1,2,3], investigations into the evolutionary history of the volcanoes based on analyses of structural and morphologic features on and around them [4,5,6], and combined morphologic studies and crater counting to improve understanding of the overall stratigraphy of the volcanoes and the surrounding terrains [5,7]. Several of these studies provide evidence to show that although Arsia, Pavonis, and Ascraeus Mons have had similar structural histories; differences between the three imply that each has a unique variation on the theme.

One particularly interesting commonality between the three Tharsis Montes shield volcanoes is the occurrence of a distinctive and unusual fan-shaped feature, extending approximately northwest, on their western flank. Based on crater counts, these deposits are thought to be among the youngest in the region, forming during the Upper Amazonian concurrent with late-stage, minor volcanism, most likely in the form of fissure eruptions on the flanks of the volcanoes [7]. Three major facies, a ridged, a knobby and a smooth facies, are generally contained within the fan-shaped deposits [6-12].

For some years there has been a debate over the emplacement of these fan-shaped features. The problem is important because it concerns nothing less than the environmental history of the Tharsis region, Mars’ most prominent tectonic and volcanic region, which extends over ~25% of the planet’s surface [13] and records some of its oldest and youngest events [14]. Based on Mariner 9 and Viking Orbiter images various interpretations of the origin of these features have been proposed including gravity-driven sliding [4,7,15,18], glaciation [16, 17], a combination of a catastrophic sliding and subsequent pyroclastic activity [6,18], and a combination of sliding, volcanism and ground-ice activity [10].

More recently, various authors [9,12,19] have used newly available MOLA data combined with MOC and THEMIS images to build on early comparisons of the fan-shaped features with terrestrial glacial deposits, such as the recessional moraines of Malaspina glacier in southeastern Alaska [17]. These later studies have used depositional frameworks of polar glaciers in the Antarctic Dry Valleys to demonstrate the consistency of the fan-shaped features with cold-based glacial deposits [9,12,19]. In this scenario, the ridged facies are interpreted as drop moraines formed from lateral retreat of cold-based ice, the knobby facies forms from in-situ downwasting of the ice in a process analogous to sublimation till formation, and the smooth terrain represents relict features of debris-covered ice or rock glaciers.

This study focuses on Ascraeus Mons (Figure 1), which is the tallest of the three Tharsis Montes, rising 17 km above the surrounding plains to attain an elevation of 18.5 km above the Mars mean datum. Our purpose is a re-examination of the Ascraeus Mons fan-shaped deposit using higher resolution data than has previously been available, namely MOC and THEMIS images coupled with MOLA data, in order to assess the plausibility of a cold-based glacial origin and other possible processes of origin for the deposit.

**Characterisation of the fan-shaped deposit:** The Ascraeus fan-shaped deposit (Figure 2) is ~180 km at its widest point, and extends ~100 km from the shield base along a N82°W trend. The deposit is emplaced on the Tharsis Plains unit (Atcb in [7,10]), which consists of Amazonian sheet-like flows from Ascraeus Mons. Overall, the deposit covers a surface area of 14,000 km², which is significantly smaller than its counterparts at Arsia and Pavonis Mons (Table 1). The Ascraeus Mons deposit covers an elevation range of 1.5-2.5 km above the Mars mean datum, with a large portion being located within a topographic depression adjacent to the western flank of the volcano. This lower section is separated from the distal margins of the deposit by an arcuate scarp with a relief of ~180 to 300 m. The southern end of this scarp appears to have been disrupted by a lobate lava flow that can be traced from a source near the SW flank of Ascraeus Mons (Figure 2). Interestingly, this flow appears to trace the rim of the depression that contains a large portion of the deposit, rather than flowing into it like earlier lava flows.

![Figure 1: MOLA topography of Ascraeus Mons superposed over a shaded relief map.](image)

| Table 1: Morphometric data for the Tharsis Montes fan-shaped features. |
|-------------------|---------------------|---------------------|
| Arista           | Pavonis            | Ascraeus            |
| Surface area (km²)| 180,000            | 75,000              | 14,000              |
| Trend            | N62°E              | N27°W               | N82°W               |
| Elevation range (km) | 2.7-5.5           | 2.9-4.5             | 1.5-2.5             |
| Length (from shield) | 450               | 235                 | 100                 |

Previous studies of the Ascraeus Mons fan-shaped deposit have reported fewer interior deposits compared to the fan-shaped deposits of Arsia and Pavonis Mons [6,20]. Zimbelman and Edgett [6] reported observations of ridged and knobby facies within this deposit, but noted an apparent lack of any unit equivalent to the smooth facies.
observed at the other Tharsis Montes. Current geomorphic mapping based on a wider suite of image data has enabled us to identify three morphologically distinct units (also described in [21]):

**Ridged Terrain.** This consists of at least seven near-continuous, concentric ridges that trace the distal edge of the deposit over a distance of ~180 km. Ridges have a mean spacing of ~800 m, but tend to converge towards the northern and southern portions of the deposit. MOLA data reveal general ridge heights of 10-40 m except for the outermost ridge, which is the most prominent reaching heights of up to 80 m. Additional smaller ridges (<10 m high) can be observed ~20 km east of the outer ridge. Ridges appear to cross-cut each other without disruption, which is consistent with the “blanket-like nature” [9] of ridges of the Arisia and Pavonis fan-shaped deposits.

**Hummocky Terrain.** This forms an almost continuous deposit covering an area of ~1800 km². Hummocks contained within this unit can be subdivided into two distinct types: (1) numerous rounded to sub-rounded hummocks, several km in diameter and tens of km in height (‘knobby facies’ in Figure 2), and (2) arcuate, ridge-like hummocks ~800 m long and ~60 m wide aligned approximately N60°E (‘complex hummocks’ in Figure 2). The former hummock type is consistent with the knobby terrains observed in the Arisia and Pavonis fan-shaped deposits, which have been interpreted as analogous to sublimation till in terrestrial glacial environments. The latter hummock type appears consistent with thumbnail terrain observed elsewhere on Mars, and interpreted as glacial in origin [e.g. 17,22].

**Flow-like Feature.** A relatively flat-topped, elevated plateau can be observed ~7 km west of the shield base in the central portion of the deposit. This mesa-like feature is 34 km long, 19 km wide, and is elongated north-south. Morphologically similar features have been described from the Pavonis fan-shaped deposit [12] and from central Acidalia Planitia [23], where they have been compared to terrestrial subglacial lava flows and table mountains, respectively. Other interpretations from studies of the Tharsis fan-shaped deposits include eskers formed by sedimentary deposition beneath or within a wasting ice sheet [10], unique lava flows [10], and shield base remnants [6].

**Interpretations and Discussion:** Based on stratigraphic relations and our interpretations of each facies, we envision two end-member for the glacial history of the area: (1) the area represents a relict landscape that has subsequently been covered by cold-based ice resulting in the passive emplacement of glacial deposits, and (2) ice has been more instrumental in carving the landscape with intimate volcano-ice interactions.

1. **Relict landscape and subsequent glacial deposition.** In this scenario, the landscape was predominantly modified prior to the accumulation of ice. The scarp formed from faulting accompanying the tectonic development of the Ascraeus shield volcano, perhaps becoming more pronounced due to the affects of aeolian erosion. The degraded western flank, the mesa-like remnant of the shield base, and exposure of a dike at the surface in the northern part of the deposit could all be used to suggest that long-term aeolian erosion has exhumed a relatively old surface. Similar to the Antarctic Dry Valleys environment [e.g. 24], this relict surface would have been preserved beneath cold-based glacial ice. Hence, ice had only a minor impact on the landscape with the deposition of ridged and hummocky terrains during its recession.

2. **Intimate volcano-ice interactions.** In this alternative scenario, glacial deposition occurred concurrent with volcanic activity. The position of the scarp represents a stable margin of a relict glacier, and formed when one or more lava flow(s) cooled and solidified against its margin. As the lava accumulates against the ice, it will rapidly form a chilled margin that insulates the hot flow interior and prevents extensive melting of ice [25]. Following scarp formation, ice advanced to the position marked by the distal edge of the ridged facies. The formation of the ridged facies is a result of step-wise retreat of the glacier. A lava flow that can be traced within the depression adjacent to the western flank was then emplaced after the ice retreated completely. Subsequent re-advance and down-wasting of the ice resulted in the superposition of knobby terrain over the underlying landscape.

**Conclusions:** Geomorphic mapping of the Ascraeus Mons fan-shaped deposit implies a more complex history for this volcano than has previously been thought. The fan-shaped deposit on the western flank of Ascraeus Mons does provide significant evidence in support of a cold-based glacial origin. However, volcanism also appears to have been a factor in the evolution of these deposits.

**References:**

THE CASTALIA MACULA REGION: NUMERICAL ANALYSIS OF A PROPOSED CONVERGENT BOUNDARY ON EUROPA.  G. W. Patterson and J. W. Head, Department of Geological Sciences, Brown University, Providence, RI, 02912 (Gerald_Patterson@brown.edu).

Introduction.  Voyager and Galileo images of Europa have shown its surface to be highly deformed by tectonic features.  Extensional and strike-slip structures are ubiquitous [1,2,3] with examples of extension on the order of tens of percent regionally [1,4].  Conspicuous however, is a general lack of evidence for compressive features on the scales necessary to compensate the observed extension.

The reconstruction of offset features along tectonic plate margins is one method of locating such features and two have been proposed, based on this method, in a region near the prominent dark spot Castalia Macula [5,6].  Reconstructions of this sort implicitly assume that the plates involved act rigidly and the validity of the reconstruction performed, as well as the assumption of plate rigidity, can be confirmed in a quantitative manner by the determination of an Euler pole of rotation.  An Euler pole is a pivot point that can describe the motion of one plate relative to another on a sphere by a rotation about the pole [7].  Such poles were not determined for either of the reconstructions that have suggested possible convergent boundaries.

Here we show the results of a numerical technique we have developed [8] to test the validity of the proposed convergent boundaries located in the Castalia Macula region.  This is accomplished by the determination of a finite pole of rotation about which the plate boundaries identified in the region can be reconstructed.  We will show, using this technique, that the assumption that the plates involved in these reconstruction behaved rigidly is incorrect and, as a result, previous reconstructions by Sarid et al. (2002) and Patterson and Pappalardo (2002) need to be reassessed.  Our analysis will instead demonstrate that the deformation inferred by the rotation of the tectonic plates defined in this region with respect to one another was most likely accommodated by internal deformation.  Furthermore, this result casts doubt on the hypothesis that some Europan bands display a morphology that is characteristic of convergent boundaries [5,9].

Background.  The Castalia Macula region is located on the trailing hemisphere of Europa’s equator and extends from -11° to 10° latitude and 208° to 232° longitude (Fig. 1).  The prominent dark spot Castalia Macula is located within this region at -2° lat. and 226° lon.  The resolution coverage of this area ranges from ~220 m/pix in the western portion of the image to ~1.5 km/pix in the eastern portion.  A diverse assemblage of structural features can be found in this region (Fig. 2).  They include bands and cycloidal ridges trending predominately east-west across the image, ridges and complex ridges with no preferred orientation, isolated lenticulae found throughout the image, and a region of chaos in the northeastern portion of the image.  Our analysis will focus on a band-like complex and set of cycloidal ridges in this region (Fig. 1 – dashed lines).

Cross-cutting relationships indicate these features divide the region into six plates that have rotated with respect to each other (Fig. 1).  Examination of 60 offset features along the margins of the plates outlined by the cycloidal ridge set suggest, to first order, that they are dextral transform boundaries (Fig. 2).  However, the offset of features along the boundaries varies from ~1.5 to 6.0 km and thirteen of the features appear to have a sinistral offset indicating that one or more of the dextral transform boundaries may be transpressive.  In order to determine quantitatively if this is the case, we need to identify Euler poles for the various plates in this region.

The inverse technique we have developed [8] determines an Euler pole of rotation using the trend and offset of preexisting features as inputs.  This method employs an iterative grid-search technique that allows us to test all combinations of Euler pole location and rotation within a specified grid.  The result is a map of the rotation for each possible Euler pole location that yields a least-squares minimum for the offset of all the features input.  The pole location with the minimum value of offset for this map is designated as the best-fit Euler pole for the features input.

Results.  The location and magnitude of five finite rotations (Euler poles) representing the best-fit reconstructions of four plates in the Castalia Macula region are shown in Table 1 (Analyses for plates 5 and 6 were not performed).
because the modeling technique we employ requires at least four offset features in order to determine a pole of rotation). The relationships between these plates offer us a unique opportunity to test the assumption by Sarid et al. (2002) and Patterson and Pappalardo (2002) that they behaved rigidly while they deformed. If we consider that the matrix addition of the Euler poles \( \mathbf{E}_2 \) and \( \mathbf{E}_4 \) as well as \( \mathbf{E}_3 \) and \( \mathbf{E}_4 \) will both yield a pole location and magnitude for \( \mathbf{E}_4 \), then we will have three independent determinations of a single Euler pole. If these plates behaved in a rigid manner during deformation then the three determinations of \( \mathbf{E}_4 \) should be equivalent. If they are not equivalent then deformation must be distributed within one or more of the plates. The determination of \( \mathbf{E}_4 \) by addition \( \mathbf{E}_2 \) and \( \mathbf{E}_4 \) indicates a pole located at \(-2.6^\circ\) lat., \(211^\circ\) lon. with a rotation of 0.99’. Determination of \( \mathbf{E}_4 \) by addition of \( \mathbf{E}_3 \) and \( \mathbf{E}_4 \) indicates a location of \(-4.2^\circ\) lat., \(50^\circ\) lon. with a rotation of 5.2’. The discrepancy between these two determinations of \( \mathbf{E}_4 \), as well as that determined by direct application of our numerical technique (Table 1), indicates that the plates in this system did not behave as if they were perfectly rigid.

An approximation of the degree to which this system deviated from perfect rigidity can be determined by calculating a difference pole between any two of the three values of \( \mathbf{E}_4 \). An arbitrary boundary can then be placed between the plates that constitute those two values and a point on that boundary can be rotated about the difference pole. The distance between the original location of the point and its location after rotation about the difference pole will then yield an approximate value for the amount of deformation that has been distributed within one or more of the plates involved.

As an example, the difference pole between the determinations of \( \mathbf{E}_4 \) by the matrix additions of \( \mathbf{E}_2 \) and \( \mathbf{E}_4 \) and \( \mathbf{E}_3 \) and \( \mathbf{E}_4 \) is located at \(-3^\circ\) lat., \(227^\circ\) lon. with 6’ of rotation. We then define an arbitrary boundary at \(-1.85^\circ\) lat. extending east and west across the image and rotate a point on the boundary at \(-227.5^\circ\) lon. about the determined difference pole (Fig. 2). Using that pole, the point is rotated \(-2.6\) km to the northeast. This is equivalent to the average offset of features along the ridges that constitute the plate boundaries in this region. The rotation also indicates that the deformation that produced the non-rigidity in this system was compressive.

<table>
<thead>
<tr>
<th>Euler pole</th>
<th>Location (º)</th>
<th>Rotation (º)</th>
<th>( V_i ) (km²)</th>
<th>( V_f ) (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \mathbf{E}_2 )</td>
<td>8 217</td>
<td>0.32</td>
<td>3.70</td>
<td>0.0904</td>
</tr>
<tr>
<td>( \mathbf{E}_4 )</td>
<td>0 28</td>
<td>-0.67</td>
<td>14.54</td>
<td>0.691</td>
</tr>
<tr>
<td>( \mathbf{E}_3 )</td>
<td>5 53</td>
<td>0.93</td>
<td>3.49</td>
<td>0.236</td>
</tr>
<tr>
<td>( \mathbf{E}_4 )</td>
<td>4 49</td>
<td>4.31</td>
<td>11.12</td>
<td>0.331</td>
</tr>
<tr>
<td>( \mathbf{E}_4 )</td>
<td>9 47</td>
<td>1.9</td>
<td>1.92</td>
<td>0.0363</td>
</tr>
</tbody>
</table>

Table 1. Determined Euler poles for the finite rotations involving the plates identified in the Castalia Macula region. The terminology \( \mathbf{E}_i \), from Table 1 indicates that plate \( y \) has been rotated with respect to plate \( x \). Rotations are counterclockwise when positive. \( V_i \) and \( V_f \) indicate the pre- and post-reconstruction variance of offset features about zero respectively.

Conclusions. Previous reconstructions of a set of ridges in the region surrounding Castalia Macula (which did not include a pole of rotation) suggested the presence of a convergent boundary along the southern [5] or eastern [6] margin of plate 4 (Fig. 1). However, determination and analysis of five finite rotation poles for four of the tectonic plates in this region indicates that, while deformation of the region was compressive, the plates did not behave rigidly. Furthermore, the amount of deformation that was distributed within one or more of the plates in this region is equivalent to the average offset of features along the plate boundaries. This indicates that distributed deformation plays a significant role in accommodating compressive stress and that previous reconstructions [5,6] suggesting a portion of the boundary that constitutes plate 4 was convergent need to be reevaluated with this in mind. Furthermore, this analysis suggests the assertion that the morphology of one of the boundaries in this region is characteristic of convergence on Europa [5,9] needs to be reassessed.

MODELING POLARIZATION PROPERTIES OF COSMIC DUST. D. V. Petrov, E. S. Zubko, Yu. G. Shkuratov. Astronomical Institute of Kharkov National University. 35 Sumskaya St. Kharkov. 61022. Ukraine. E-mail: petrov@astron.kharkov.ua

Introduction: Solid particles in space may form clusters (aggregates) due to particle collisions and adhesion. Since many light scattering effects cannot be explained within the framework of model of aggregates of compact particles [1], for more adequate simulation of scattering properties of cometary and interplanetary dust some hierarchical (fractal-like) properties of aggregates should be taken into account. Structures that are formed at association of solid grains, in the case of a diffusive character of their movement, are called fractal clusters or fractal aggregates. The fractal clusters have characteristic branchy structure. For the last years intensive investigations of such objects by computational methods were carried out [e.g., 2]. Computer experiments allowed us to gain some insight into such aggregates and character of their formation.

We note that processes of fractal cluster formation at association of solid grains are related to a number of many other physical systems and processes. These are solidification of colloidal solutions, coagulation and percolation processes, formation of polymers, dielectric break-down, etc. This also includes formation of aggregated particles in space and on the planetary surfaces.

Used model of aggregates: To construct pre-fractal clusters, we use the Whitten-Sander model, generalizing it for the 3-D case. According to this model we divide a limited 3-D spherical volume on a set of cubic cells. Then we locate in the volume one particle (grain) adding another particle one by one. Each new particle moves in an adjacent cell in random mode – its path is selected by the Monte Carlo method. If the particle reached the border of the volume, it is reflected from the border proceeding its movement till it appears in the neighborhood of one of particles of the cluster. Then it is stopped in a given cell, and next particle begins its moving in the volume. This process forms a fractal cluster.

Let us consider one more model of generating fractal clusters – so-called “random rain” model. As before we divide restricted three-dimensional space on a set of cubic cells, one particle seats in the center of range, and further particles – candidates for addition – drop on growing cluster, as rain drops, along some random directions from the volume border. Each particle starts the motion from randomly selected cell and move along some random chord [2]. Fig. 1 and 2 show examples of Whitten-Sander (a) and “random rain” (b) fractal clusters for 2-D and 3-D cases, respectively. As can be seen the “random rain” model generates more ramified structures.

Results and discussion. To investigate polarization properties of random inhomogeneous chaotically oriented dielectric scatterers we used the DDA method. Figure 3 shows the phase curves of the intensity and degree of linear polarization of scattered light for Whitten-Sander’s (a) and “random rain” fractal aggregate (b) with the size parameter $x = 8$ (solid lines), 10 (dashed lines) and 12 (dot lines), where $x = 2\pi r/\lambda$, $r$ and $\lambda$ being the radius of spherical volume and wavelength, respectively.
Calculations were done at the refractive index $m = 1.5 + 0.1i$ which corresponds to “organic” particle. As can be seen in Fig. 3, the degree of polarization originating at light scattering on fractal aggregate increasing the size parameter makes the negative polarization more prominent.

**Conclusion.** We have made calculations for various size parameters and different types of fractal particles. We find that the negative polarization of aggregated particles depends weakly on the fractal cluster type. Increasing the size parameter makes the negative polarization more prominent.


Fig. 2. Examples of three-dimensional fractal clusters from the Whitten-Sander (a) and “random rain” (b) models.

Fig. 3. Phase curves of intensity and degree of linear polarization for the Whitten-Sander (a) and “random rain” (b) fractal aggregate.
**Moonrise: Sample Return from the Enormous South Pole-Aitken Basin**

Carle Pieters¹, Mike Duke², and the Moonrise Team;  
¹Brown University, RI USA; ²Colorado School of Mines, CO USA

The Moonrise mission, which will land in the South Pole-Aitken (SPA) Basin and return samples to Earth, has been selected for Phase A studies by NASA as part of the New Frontiers program. Analysis of Moonrise samples will provide critical new insights into the first half-billion years of Solar System history and early planetary evolution. Specifically, the returned samples and their geologic context will a) be used to date the oldest and largest of the preserved impact basins and test (establish) early basin impact chronology for the Moon and in the inner solar system, b) provide materials derived from deep in the crust or upper mantle of the Moon to better understand the Moon's structure and differentiation, and c) afford unique information to better understand giant-basin-forming processes.

Dr. Mike Duke is Moonrise Principal Investigator with Dr. Brad Jolliff of Washington University acting as Deputy Principal Investigator. Science Team Co-Investigators participate from ten institutions. The Jet Propulsion Laboratory and Lockheed Martin are the principal partners for mission design, management, systems engineering, and operations for both the spacecraft and the landers. Malin Space Science Systems provides the imaging systems. During the summer of 2005, one of two missions currently in Phase A will be selected to proceed to Phase B. Moonrise is scheduled for launch early 2010 and samples will be returned a few months later. Moonrise will deploy two landers that will target two independent sites within SPA and land one month apart. The landers are relatively simple; their primary goal is to collect materials that represent the geologic and lithologic diversity of SPA basin. This will be accomplished by collecting large numbers of representative rock fragments of the regolith and returning them to Earth for detailed analysis.

<table>
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<tr>
<th>Science Team Co-I</th>
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<tr>
<td>D. Bogard</td>
<td>Johnson Space Center</td>
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<td>G. Lofgren</td>
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<td>M. Drake</td>
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<td>P. Warren</td>
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The Moonrise team invites the international community to participate in site selection. Two open workshops will be held during Phase C/D for identifying and prioritizing pairs of targets within the basin for landing sites. The first is likely to be held January 2008 and the second a year later. Sites proposed using additional new data from SMART-1, SELENE, Chandrayaan-1, and Chang'E would be welcome. Science and engineering requirements for potential targets will be made available well in advance of the workshops. The current science requirements emphasize sites that a) have well developed soil that contain representative fragments of surrounding and regional geologic formations, b) have a high probability of sampling SPA impact melt breccia or deeply derived material, and c) contain (but are not dominated by) SPA basalts and/or cryptomaria. More detailed information about the site selection process and requirements will be announced after Moonrise enters Phase B.
PALEOSTRUCTURES OF THE MIDDLE CLARITAS FOSSAE REGION, MARS, AS STUDIED FROM THE MOLA AND MEX HRSC DATA. J. Raitala1, M. Aittola1, J. Korteniemi1, V.-P. Kostama1, E. Hauber2, P. Kronberg3, G. Neukum4 and the HRSC Co-Investigator Science Team. 1Astronomy Division, Univ. of Oulu, Finland, (jouko.raitala@oulu.fi), 2Inst. of Planet. Res., DLR, Berlin, Germany, 3Inst. of Geology, TU at Clausthal-Zellerfeld, Germany, 4Inst. of Geosci., Dept. of Earth Sciences, Freie Universität, Berlin, Germany.

Introduction: Claritas Fossae (CF) deforms the southern reach of the Tharsis uplift. Volcanic, tectonic, and water-/ice-related events have alternated in its complex geologic history. Long tectonic faults cut the surface which consists of old uplands, lava plains, and fluvial and colian formations [1].

Fig. 1. MOLA topography of Claritas Fossae.

The N-S to NNW-SSE structures of CF dominate the southern MC-17 and the northwestern MC-25, and cut through remnants of ancient highlands and Tharsis-related lava plains. The ruggedness of the highland outcrops reveals them to be of Hesperian to Noachian age [1]. In the south, CF crosses NEE-SWW-trending Thaumasia Fossae. The Claritas-Thaumasia tectonics may reflect plume-related dike intrusions [2]. These extensional features were used to compute the extensional strain of Tharsis [3]. The wide intersection area has numerous channels, including Warrego Valles, indicating the previous existence of flowing water. An intrusive body may have caused heating and hydrological activity forming the channel networks [4,5].

Claritas Rupes (CR). One of the youngest important fault, best visible in the MOLA topography (Fig. 1), is the CR scarp. It extends from the high reaches of CF close to Noctis Labyrinthus to Thaumasia in the south. In the north, CR has a higher W side and a lower E side. Just to the south of this, it has a graben-like character while in the middle and south it has a higher eastern side and a lower western side (Fig. 2). Most of CF consists of a wide graben-horst set which cut through the units, thus indicating the relative youth of these faults. This gives a relative marker to find older geological events.

Fig. 2. MOLA profiles (cf. Fig. 1) of Claritas Rupes.

Fig. 3. Lava-flooded, tectonized graben of Tantalus as seen in the HRSC red channel image.
**Tantalus Highland.** The lava plains, originating from the flanks of Arsia Mons and Syria Planum, cover most of the surface leaving only the highest ancient uplands to peak out of them [cf. 6]. Tantalus Highland, being slightly larger than the rest of the peaks, is located on the middle eastern side of CR. Together with the smaller highland peaks, Tantalus provides clues to the paleostuctures in the Claritas area. A closer look of the area reveals several sets of older NEE-SWW to E-W throughs. Tantalus Fluctus itself is a graben depression which, within the Tantalus Highland, has a flat lava-flooded bottom, later cut by the crossing faults of the Claritas set. To the west of CR, the graben has fluvial characteristics. Five Tantalus phases were identified [cf. 6]: The old upland was cut (a) by E-W graben, which provided (b) channels for water, and (c) flood lavas, which smoothed the higher reaches of the graben bottoms. Later (d) the graben-channel system was cut by Claritas faults resulting in CF (Fig. 3), and finally, (e) the lowest reaches were covered by Arsia lavas.

**Tantalus Paleolake.** Tantalus Highland provided water to downhill directions. The best preserved system is found to the south of it. A deep depression (Figs. 4,5), closed by Tantalus Highland in the north and by smaller uplands on the other sides, acted as a sink for water flowing in from the surrounding high areas [cf. 6,7]. The paleolake finally drained through an outflow channel which broke the lowest western upland side of the depression and led water through a crater and further into the northern Icaria Planum.

**Conclusion.** The region is ideal in finding causes for valley formation tied to climate and geology, as volcanic features, craters, rifts and other topographic features can be observed here. The flow features are tied to topography and regional slopes, but they may also relate to hydrothermally driven outflows due to volcanism and tectonics. Melting of snow and ice due to areothermal heat might have been responsible for the formation of the paleolake and channel.

![Fig. 4. Mola topography over the Tantalus paleolake.](image4.png)

![Fig. 5. MOLA profiles of the paleolake (cf. Fig. 4).](image5.png)

![Fig. 6. The HRSC red image shows the channel (centre) which broke the Tantalus paleolake (right) rim draining it through an impact crater into northern Icaria Planum (left). Additional sapping added water from the larger and olded crater (top).](image6.png)

**References:**

The software for image analyses, based on Delphi 6 is presented. By using this program, analysis of some celestial bodies images, with the target to detect photometric contrast areas on their surfaces, is done. All considered images can be classify by degree of contrast. An example of low contrast is the image of Annefrank asteroid. In this case the contour map of isophotos, which reflect main details of Annefrank shape, is constructed. In case high contrast image, the increase of small details contrast is possible. So, at the developed image of Phobos, the inhomogeneous structure of the inner wall of the largest satellite’s crater is clear, and famous furrows are continued to satellite’s limb.

At present time, program requireq bmp-format of input image file. The program allows to build brightcurve along the profile, change thickness and number of lines. In some, the simplest cases, aside from writing output information in bmp-file, the conservation of the result in the vector format is possible.

In addition to astronomical application, the program can be used for the test of images of the terrestrial surface for the purpose to study varied ecological and cartographic problems.
In the 40s of the last century some new geomorphologic forms were discovered in North America (J. Pardee) and in the 80s – in Eurasia (V.V. Butvilovsky, A.N. Rudoy), they were called “giant current ripples”. Their development was associated with the processes which had occurred within the outflow channels during cataclysmic glacial superfloods – floodstreams (diluvial floods) resulted from the outthrows of great Late-Pleistocene ice-dammed lakes. Giant current ripple marks – are active channel relief forms which developed on the near-talweg areas of the pre-core parts of main valleys of cataclysmic outflows out of the outburst water bodies, the latter being of the ice-dammed origin, as a rule (A.N. Rudoy). These marks are morphologic and genetic macroanalogues of small sand ripples in the contemporary streams and rivers, but they are by two-three orders bigger in dimensions and are formed with boulders and pebbles with minor participation of smaller fractions.

The relief of giant current ripples serves very often as the basis for hydraulic reconstructions (work by V.R. Baker and P.A. Carling), which enable calculation of discharges, velocities and depths of the water streams across the ripple fields. The outcomes of these calculations do not reflect the maximal values of chief parameters of superfloods (these values are achieved at the channel lines within narrower parts of the channels, i.e. – within gorges where no ripples are formed, but the destructive flood activity predominates). These outcomes, however, give a notion of the grandiosity of water-glacial cataclysms (discharges of the floodstreams in the Altai over the fields of giant current ripples, according to V.R. Baker and P.A. Carling, would exceed a half million cubic meters per second).

At the boundary of the 20th-21st centuries the publication activity of the opponent of the diluvial origin of giant current ripples became very active. Reports and papers which appeared at the time inherited to a certain extent the opinion of B.A. Borisov who had always written that the ridged relief in the mountains of the southern framing of the USSR was ribbed moraines. But practically at the same time these works suggested some different, frequently polar with the same authors, hypotheses. In particular, some opinions were expressed, that the current ripples in the Kuray basin are: 1) inversional marginal intraglacial water-ice relief (Okishev, the 70s); 2) small-ridged, polyridged ribbed moraines (Okishev, the 80s); 3) tectogenic relief (Okishev, beginning of the 21st century, together with A.V. Pozniakov); 4) cryogenic relief (Okishev, beginning of the 21st century, together with A.V. Pozniakov); 5) no ripples but something from the things mentioned above (Baryshnikov and others, beginning of the 21st century). Giant current ripples at the site of Platovo-Podgornoye settlements at the Altai foothills are: 1) “megarifels” (Okishev, 1980s); 2) giant current ripples (G. Ja. Baryshnikov, the 90s); 3) ravines (D.A. Timofeev, beginning of the 21st century, oral report). These and some other authors, who published all these intercontradicting hypotheses, do not, nevertheless, polemize with each other, and, moreover, do not argue with themselves. For that, they all together appose the diluvial origin of giant current ripple marks polemizing furiously and originally with the authors of the diluvial theory.
Meanwhile, both in America and Russia the structure and morphology of this relief have been studied rather well. The very fact of the presence of the giant current ripple relief in the river valleys, on the one hand, enables reconstruction of the water-glacial events at the upper reaches of the rivers and their dating to some extent. On the other hand, it gives the opportunity assembled with other types and forms of the diluvial morpholithocomplex to restore the hydraulic parameters of the water streams and to estimate the geologic role of cataclysmic processes, which had been caused by climatic alterations, glacial dynamics, transgressions of the lakes and outbursts of the latter, in the formation of the contemporary surface of the dry land.

The report presents diagnostics of the giant current ripples according to their morphology and structure on the basis of the well-studied locations of the relief in different regions of the world.
LAKE SEDIMENTS OF THE NORTHERN ALTAI AS AN INFORMATION SOURCE ABOUT CLIMATE ALTERATIONS DURING PLEISTOCENE AND HOLOCENE IN THE SOUTH OF WESTERN SIBERIA

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Detailed field works on the territory of the northern area of medium and low mountains in the Altai discovered over 50 lakes of various genezes with their total area of 530 km² and water volume of 17 km³, which are dated back to Pleistocene and Holocene. Most of the lake depressions are dry at present. These lakes have been classified, among them evorsional lakes and dammed lakes group are of the greatest palaeoglaciologic interest. Formation of the evorsional lake depressions is associated with evorsional and cavitation activity of Late-Würm floodstreams in the Katun River valley (the well-known examples are the survived lakes within the evorsional-cavitational depressions of Manzherok and Aja, which were formed behind the side ridges and within valley extensions in the counter-stream zones with vast stationary whirlpools). Among the lakes belonging to the dammed group, diluvial-, ice- and collapse-dammed ones as well as flood-blocked lakes are of much information.

Diluvial-dammed lakes developed at the mouths of all tributaries of the main valleys of cataclysmic outthrows of giant ice-dammed lakes out of the intermontainous basins in the Central and South-eastern Altai (Chuya, Kuray, Ujmon and other basins) due to blocking of the tributary outflows with diluvial masses thrown down by the superfloods.

In the depressions of most diluvial-dammed lakes there are sediments which have been bio- and rhythmstrographically analyzed. Some samples for radiocarbon dating have been selected according to most of them. The structure of the lake diluvial-dammed units reflects the sequence of the lake-glacial events in the Altai highlands and alterations in the climate conditions in the north of the Altai and in the south of Western Siberia for the dated chronological intervals. And the absolute datings proper of various bunches of these sediments point at the time of cataclysmic outbursts out of the basinal ice-dammed lakes and at their number. The correctness of the reconstructions gained is checked according to the data received by the analysis of bottom sediments from other lake territories.

On the whole, about 17 thousand years ago the air temperature of the northern part of the Altai hardly differed from the contemporary ones in the Chuya basin (-5°C and lower degrees – the average annual and the average perennial marks, and +3.8°C – the average one in July) with, however, much higher humidity which possibly exceeds the contemporary marks. The landscape-climatic zone limits went down apparently not less than by 1000 m. The average annual temperatures in Late Drias were by 4°C lower than modern ones, and the snow-line depression was on the average not less than 600-650 m. The stage glacier expansions on the background of the general degradation were of a surge character. The basinal ice-dammed lakes burst out into the highlands. There was not less than 5 floodstreams in the Bia River valley, and in the Katun River valley – not less than 6 ones, four of the superfloods which occurred in the valleys being synchronous. The earliest floodstream in the Katun River valley is dated back to the initial stages of the Würm glacier, the next one (the second one for the Katun and the first one for the Bia) occurred about 45 thousand years ago, the third floodstream for the Katun was about 20 thousand years ago, the fourth one for the Katun and the second one for the Bia – about 17 thousand
years ago, the fifth for the Katun and the third one for the Bia 16190-13220 years ago and the last (the sixth and the fourth ones accordingly) – at the end of the Middle Drias. The maximal floodstream discharges gained by means of the HEC-2 programme were about $18^6 \text{ m}^3\text{ps}$ (Baker, Benito, Rudoy, 1993). The maximal floodstream discharges gained by means of the HEC-RAS 3.0 for the pre-mouth area of the Chuya River valley were $8^6 - 12^6 \text{ m}^3\text{ps}$ (Herget, Agatz, 2003).
A TREATMENT OF DATA BANK OF MORPHOLOGIC CATALOGUE OF MERCURIAN CRATERS. B. D. Sitnikov., E.A. Kozlova, J.F. Rodionova. Sternberg State Astronomical Institute, Moscow, jeanna@sai.msu.ru.

The bank of data of Morphological catalogue of Mercurian craters in diameters 10 km and more including 6 500 craters have been compiled for 45% of the surface of Mercury in Sternberg State Astronomical Institute on the base of Mariner 10 data [1]. Coordinates of craters in the catalogue are changed in accordance with [2] new data of Mercury’s North Pole.

A treatment of data bank is fulfilled with the use of software in regime of “client-server” on the choice of ensemble of the objects, compiling maps of density distribution of craters. The maps of density distribution of craters with different degree of rim degradation permit to determine the regions of concentration a “young” and “old” craters. The average density distribution of mercurian craters is 193 on 1 million km². Table 1 show the quantity of craters in % with different morphological features on the investigated part of Mercury and on the Moon. The craters of class 3 (smooth rim) are prevailed on the investigated part of Mercury in comparison with the Moon [3, 4].

48% of the mercurian craters has diameters 10-20 km, about 30% - diameters 20-40 km. 51% of mercurian craters has terraces and faults on inner slopes instead of 7.5% on the Moon. The reason of a great number of terraces and faults of mercurian craters may be connected with stress deformation of the surface of Mercury. Hills on the floor of mercurian craters are twice more than on lunar craters. Flat floor of craters on Mercury is met often than on the Moon. There are three times it is more craters on the plains of Mercury than on the lunar maria. It is necessary note that it was difficult to determine the morphology features of many craters of Mercury because of pure quality of mages in some regions.

Fig.1 show the density distribution of craters in diameters 10 km and more on the investigated part of Mercury. We turned the Molveide projection on -20º along meridians. The scale of the map show the density of craters - the quantity of craters referenced to 1 million km².

<table>
<thead>
<tr>
<th>Morphological features</th>
<th>Mercury</th>
<th>Moon</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rim degradation</td>
<td>1</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>46</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td>5</td>
<td>10</td>
</tr>
<tr>
<td>Terrace</td>
<td>38</td>
<td>6</td>
</tr>
<tr>
<td>Faults</td>
<td>13</td>
<td>1,5</td>
</tr>
<tr>
<td>Hills</td>
<td>27,5</td>
<td>12,9</td>
</tr>
<tr>
<td>Central peaks</td>
<td>7,3</td>
<td>5,5</td>
</tr>
<tr>
<td>Ridges</td>
<td>0,9</td>
<td>3</td>
</tr>
<tr>
<td>Chains</td>
<td>10,5</td>
<td>12</td>
</tr>
<tr>
<td>Fissures</td>
<td>0,3</td>
<td>2</td>
</tr>
<tr>
<td>Flat floor</td>
<td>26,2</td>
<td>7</td>
</tr>
<tr>
<td>Rough floor</td>
<td>35,3</td>
<td>71,5</td>
</tr>
<tr>
<td>Dark material on a part of floor</td>
<td>8,5</td>
<td>11</td>
</tr>
<tr>
<td>Dark material on total floor</td>
<td>3,5</td>
<td>0,1</td>
</tr>
<tr>
<td>Local terrain: highland</td>
<td>82</td>
<td>94</td>
</tr>
<tr>
<td>Local terrain: plains</td>
<td>11</td>
<td>3</td>
</tr>
<tr>
<td>Local terrain: transitional zone</td>
<td>6,5</td>
<td>3</td>
</tr>
</tbody>
</table>

Table 1 Quantity of craters with different morphological features at Mercury and the Moon in %.

Fig.1 The density distribution of craters in diameter 10 km and more on Mercury referenced to 1 million km².
Introduction. Tharsis is one of the most prominent features on Mars. Volcanism and tectonism associated with the plateau by far exceed the levels of activity in other areas of the planet. This monopolar distribution of tectonism and volcanism led to the suggestion that the planform of mantle convection on Mars is dominated by a single, long-lived, thermal plume originating at the core-mantle boundary similar to a terrestrial plume but much larger [1].

Although the plume model explains some features of Tharsis, there are both observational and theoretical reasons to consider alternatives. First, the plume model has not reproduced Tharsis development on timescales consistent with observations [2]. Second, constructional volcanism seems to be a major contributor to Tharsis elevation [3]. A thermal plume contributes only a fraction to the present-day topography and areoid [4]. Third, the plume hypothesis implies an actively convecting mantle and a sufficiently large heat flux from the Martian core. However, in the absence of plate tectonics, convection in the Martian mantle is likely to be sluggish or absent [5] and mantle heating shuts off any core heat flux and associated plume activity [6]. Finally, large variations in Sm/Nd and Lu/Hf ratios among shergottites suggest a heterogeneous mantle which retains an isotopic signal of initial differentiation [7]. This also argues against vigorous convective mixing.

An alternative hypothesis is that Tharsis could be associated with a large impact early in Martian history. Geodynamical consequences of this hypothesis were investigated in [5]. It was shown that impact-induced plumes can survive for the entirety of planetary evolution and can contribute to the present-day areoid. Production of Tharsis by such a long-lived, impact-related plume requires neither globally occurring convection nor generation of plumes at the core-mantle boundary. In the present study, we explore if this hypothesis can also explain the magmatic evolution of Tharsis.

Model. To model evolution of impact-induced thermal and compositional heterogeneity we use the spherical shell code TERRA in which compositional information is carried by particles. The region heated by the impact shock wave, isostatic adjustment, and core formation is assumed to be hemispherical with radius $R$ left as a model parameter. Heating is assumed to be uniform with $\Delta T = 300$ K while upper mantle temperatures quickly drop to the solidus temperature. Because any core heat flux was probably short lived the bottom boundary is assumed to be insulating throughout evolution. Melting results in a decrease of residual mantle density and an associated compositional buoyancy. Exponential temperature-dependent viscosity is considered.

Results and discussion. Initial, localized upwelling results in decompression melting and additional depletion buoyancy producing an extended period of magmatism the duration of which depends on interior mantle viscosity (Fig. 1). Decompressed material spreads out at the bottom of the viscous lid. For all cases, melt production decays with time from an initial maximum to very low levels. As interior viscosity decreases, the decay rate and total melt volume decrease and increase, respectively. In all cases, widespread volcanism is suppressed by lid thickening, and volcanism is concentrated in the thermochemical plume region. For low interior viscosities, there are two spatial scales in the distribution. The outer scale is that of the impact plume. The inner scale is associated with localized, small scale convection within the plume.

The quantity of melt generated by large impacts can be sufficient to obliterate evidence of the initial crater. Scaling laws suggest that the ratio of crater volume to retained melt volume produced by the initial impact shock wave increases with transient crater radius for Mars, the ratio reaches $\sim 1$ when the transient crater radius is $\sim 2000$ km. If it is assumed that the initial thermochemical anomaly region corresponds to shock wave pressure $\geq 50$ GPa (corresponding to $\sim 20$% partial melting at the boundary), then for initial anomaly radius on the order of half the mantle thickness, $R \sim \Delta r/2$, the initial retained melt volume is $\sim 1/2$ the crater volume. Additional melt volume produced during the extended period of magmatism can result in a total melt volume (initial shock melting plus extended decompression melting) equal to the crater volume if the interior viscosity is sufficiently low. For $R \sim \Delta r$, the initial retained melt volume $\sim$ crater volume. In this case, additional decompression melting associated with the impact plume can result in volcanic construction.

The topography and material contained within a depression due to Tharsis loading and lithospheric flexure correspond to $\sim 3 \times 10^8$ km$^3$ of igneous material [8]. While the $R \sim \Delta r/2$ case is only capable of producing a total melt volume $\sim$ initial crater volume, for $R \sim \Delta r$, all of the melt produced subsequent to initial shock melting is available for igneous construction. Decompression melt volumes for interior viscosities $\eta \leq 10^{22}$ are on the order of the observed volume of Tharsis igneous material. The duration of large scale melting for all cases is $< 1$ Gyr which is approximately the time by which Tharsis was emplaced [8].

Figure 1: Magmatic evolution of impact-induced thermochemical anomalies for the case $R = 1500$ km. Total magmatic rate as a function of time (left). Cumulative spatial distribution of magmatism (center). Final distribution of depleted mantle material along a cross section passing through the hemispherical anomaly axis (right). Color (white through red) indicates degree of depletion and the solid line indicates the initial radius of the hemispherical anomaly.
**MAPPING OF COMPOSITIONAL ANOMALIES IN AGGLOMITATES ON THE LUNAR SURFACE.**

L.V. Starukhina\(^1\), Yu. G. Shkuratov\(^1\), V.V. Omelchenko\(^1\), C.M. Pieters\(^2\), \(^1\)Astronomical Institute of Kharkov National University, Sumskaya 35, Kharkov, 61022, Ukraine; \(^2\)Brown University, Providence, RI 02912 USA, starukhina@astron.kharkov.ua

**Introduction:** Composition of mare and highland agglutinates was shown to be slightly shifted to the average between that of mare and highland [1], similar to the shift observed for regolith samples as compared to the local bedrocks [2]. Mare agglutinitic glasses are Al-enriched and depleted of Fe, Ti, and Cr, whereas highland ones are depleted of Al and enriched in Fe, Ti, and Cr. This suggests global mixing processes that affect predominantly the particles most exposed to the space environment, i.e., agglutinates [1]. Such processes, namely, transport of dust and atoms via the lunar exosphere, were considered in [3,4]. Here we continue the simulation of the regolith compositional evolution mapping agglutinate compositional anomalies on the lunar surface.

**Maps of agglutinate composition anomalies:** Detailed compositional and spectral analyses of lunar soils carried out by the Lunar Soil Characterization Consortium (LSCC) [e.g., 5] give a unique opportunity to develop a technique that might be successfully used in remote compositional analysis of the lunar surface. The analysis supplies us with chemical data not only for bulk samples, but also for mineral components including agglutinates. We used the LSCC data to find relationships that provide the highest correlation coefficients between various linear combinations, \(\log(P) = aA_{415} + bA_{430} + cA_{480} + dA_{500}\), of albedo \(A\) (in %) at Clementine UVVIS spectral bands and the parameter \(P\) that presents the abundance of a mineral or chemical element [6]. We have found the coefficients of the linear combinations (Table 1) for agglutinate content (in %), bulk (total) abundance of FeO and Al\(_2\)O\(_3\), and abundance of FeO and Al\(_2\)O\(_3\) in agglutinates. The corresponding correlation coefficients are also presented in Table 1.

<table>
<thead>
<tr>
<th>Table 1.</th>
<th>(a)</th>
<th>(b)</th>
<th>(c)</th>
<th>(d)</th>
<th>(e)</th>
<th>(k)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Agg. cont</td>
<td>-0.062</td>
<td>0.04</td>
<td>-0.063</td>
<td>0.052</td>
<td>1.776</td>
<td>0.84</td>
</tr>
<tr>
<td>FeO</td>
<td>-0.063</td>
<td>0.102</td>
<td>0.029</td>
<td>-0.112</td>
<td>1.376</td>
<td>0.89</td>
</tr>
<tr>
<td>FeO agg.</td>
<td>-0.116</td>
<td>0.107</td>
<td>-0.008</td>
<td>-0.04</td>
<td>0.888</td>
<td>0.83</td>
</tr>
<tr>
<td>Al(_2)O(_3)</td>
<td>0.052</td>
<td>-0.082</td>
<td>0.046</td>
<td>0.107</td>
<td>0.991</td>
<td>0.86</td>
</tr>
<tr>
<td>Al(_2)O(_3) agg.</td>
<td>-0.006</td>
<td>-0.029</td>
<td>0.169</td>
<td>0.191</td>
<td>0.961</td>
<td>0.74</td>
</tr>
</tbody>
</table>

In Fig.1 the map of the ratio of Al\(_2\)O\(_3\) in agglutinates to total Al\(_2\)O\(_3\) percentage is presented, in Fig.2 the same for FeO is shown; both maps are normalized by agglutinate content. As one can see in Figs.1 and 2, mare agglutinates are enriched in Al\(_2\)O\(_3\) and depleted in FeO as compared to the bulks, whereas the agglutinates on highlands the opposite relation is observed. As can be anticipated in both these cases the highest anomalies are observed for young craters and their ray systems.

**Chemical composition evolution of regolith particles:** Change in chemical composition of fine particles (<10 \(\mu\)m) occurs in two processes. (1) Loss of surface atoms in sputtering by solar wind protons and evaporation of target material in micro-meteorite impacts and deposition of atoms that were sputtered or evaporated at distant sites. (2) Contamination of the upper regolith layer by high-velocity dust ejecta from distant craters. As dust particles that cover mare-highland distances have speed \(\approx 1\) km/s, they penetrate in the regolith to some depth. Thus so only a part of high-velocity dust ejecta contribute to agglutinate particles. Following [3,4], global deposition flux of sputtered and evaporated atoms can be estimated as \(8\times10^7\) cm\(^{-2}\) s\(^{-1}\), or \(3\times10^{12}\) g cm\(^{-2}\) s\(^{-1}\). Among the sputtered atoms, heavier elements are preferentially deposited [3,4], so in atomic mixing, the abundance of Fe increases as compared to its abundance in the regolith composition averaged over the Moon. This average composition (21.8 wt.% Al\(_2\)O\(_3\) and 7.3 wt.% FeO) is the limit point for mixing by the dust mechanism. Total flux of globally distributed dust was estimated with the mass spectra for small impactors [7], taking crater-to-projectile diameter ratios from [8,9]. The fraction of dust that takes part in global transport was calculated using the ejecta velocity distributions from [10,11]. The resulting flux of dust amounts equals \(1.1\times10^{16}\) g cm\(^{-2}\) s\(^{-1}\) (or \(3\times10^{16}\) cm\(^2\) s\(^{-1}\)), \(\approx 40\%\) of it being due to the ejecta from microcraters on particle surfaces.

The best fit with the compositional trends on \(\text{Al}_2\text{O}_3\)-FeO plane from [1] is obtained, if \(\approx 60\%\) of globally distributed dust enter into agglutinates, i.e., for atoms to dust mass ratio \(\approx 1:2\). In Fig. 3 the simulated trends are shown. The total time scale is of the order of that found in [12] for total exposure time for a fine particle on the regolith surface (1.5·10\(^7\) years). The modeling show that the ratios of the final Al\(_2\)O\(_3\) and FeO contents to their initial contents for mare and highland particles are about the same, as is obtained for mare and highland areas in Figs.1 and 2.

**Conclusions:** Maps of prognosis of bulk and agglutinate composition of the lunar regolith were obtained using the technique proposed in [6]. This technique uses LSCC and Clementine data. Simulation of the chemical evolution of a regolith particle exposed to lunar exosphere and to fine dust particles of ejecta have shown that this mechanisms of global mixing can account for the compositional anomalies observed for agglutinitic glasses [1] and obtained for the entire lunar surface.
MAPPING OF COMPOSITIONAL ANOMALIES IN AGGLUTINATES: L. V. Starukhina et al.

Fig. 1. $\text{Al}_2\text{O}_3(\text{agg.})/(\text{Al}_2\text{O}_3(\text{total})\cdot \text{agg})$

Fig. 2. $\text{FeO}(\text{agg.})/(\text{FeO(total)}\cdot \text{agg})$

Fig. 3. Simulation of the evolution of composition of 10 µm regolith particles; initial and final points of the trends correspond to those in (Pieters, Taylor, 2003), total time intervals are 330 thousand years.

THE MORPHOMETRIC ANALYSIS OF THE FEATURES OF MARTIAN CRATERS (10 – 20 km). I.A. Ushkin¹, G. G. Michael². 1. Moscow State University, Vorobjovy Gory, 119899, Moscow, Russia, gray_pigeon@mail.ru. 2. ESA, Noordwijk, the Netherlands, greg.michael@rssd.esa.int.

In our work [1] we got data for 87 large martian craters with diameters from 110 up to 411 km (Fig.1). In the present work the morphometric parameters for 180 small martian craters (with diameters from 10 up to 20 km) [2] have been determined: depth of a crater $\Delta H$, height of a rim $h$, with the help of profiles constructed on the basis of supervision of space vehicle MGS [3]. The comparison with similar morphometric parameters of small lunar craters [5] also is fulfilled. An attempt of an estimation of thickness of layer of regolith of the planet as result of distribution of approach from the work of Melosh [4].

The following extreme parameters of sizes 10-20 km are received: the maximal values of them are those: $<\Delta H> = 1668$ m, $<h> = 612$ m. Their minimal values the following: $<\Delta H> = 54$ m, $<h> = 0$ m. Figure 2 show one of characteristic half profile. Comparison of the received results with morphometry of lunar craters [5] has been lead. As result of generalization of calculations we obtained the following dependence (Fig.3): A degree of degradation – crater’s depth.

$\Delta H(RD)=-139,8*RD+878,2$ for the Mars,

$\Delta H(RD)=-620*RD+3300$ for the Moon.

And a degree of degradation - height of a rim (Fig.4):

$h(RD)=-17,4*RD+168,6$ for the Mars,

$h(RD)=-67*RD+545$ for the Moon.

The height of lunar crater rim [5] for the same degree of degradation is more than for the height of martian crater rim.

Graphic generalization of results in the following conclusion: for the same degree of degradation such morphometric characteristics of lunar craters as depth and height of a rim are expressed more strongly, than at martian craters. It is a result of stronger gravitation on Mars (as speech here goes about large craters for which its role is especially important), and also active atmospheric processes.

**Estimation of thickness of layer of regolith.**

Geological targets are not homogeneous and isotropic and have no ideally flat surface. In real situations we deal or with layered targets, or with the targets consisting from casual of rocks with various mechanical properties, but influence of these roughnesses of a relief on process of formation of a crater till now is badly investigated. The most investigated case - a layered target: the soft layer lays on strong material (it is investigated at the end of 60-th [4]). It has been found, that the morphology of a resulting crater strongly depends on the relation of diameter of a crater on a crest of a rim (D) and thickness of a layer. Process of an estimation of thickness a layer of regolith of a planet on this method (more detailed description of it can be found in [4]) is reduced first of all to correlation of a crater with one of four characteristic morphological attributes – presence of the central hill, a flat bottom, a concentric crater and normal morphology.

In our work we got the senses for layer of regolith such as: maximum thickness of layer of regolith is more 3 km and minimum thickness of layer of regolith is less than 1.2 km.

**References:**

Figure 1. Dependence “rim degradation – crater’s depth” for diameters 110-411 km (brown line – for Mars, blue line – for the Moon)

Figure 2. Characteristic half profile (coordinates: latitude 60.9º, longitude 157.2º; diameter – 12 km)

Figure 3. Dependence “rim of degradation – crater’s depth” (red line – for Mars, gray line – for the Moon)

Figure 4. Dependence “rim of degradation – height of a rim” (red line – for Mars, gray line – for the Moon)
THE DECISION OF PROBLEMS MLS IN CONDITIONS OF INTERDEPENDENCE OF ASTRONOMICAL PARAMETERS. S.G. Valeev, T.E. Rodionova, V.E. Zharov Ulyanovsk State Technical University, st. Severny Venez, 32, Ulyanovsk, Russia

In so-called parametrical models of data processing in which the basic result is the vector of MLS-estimations of parameters, the basic problem is decrease in influence of interdependence (multicollineation - MC) variables which amplifies because of presence at model statistically insignificant, but correlating with significant, composed. Unaccount MC results in significant casual and regular mistakes, and even at all to discouraging results. Influence MC can be lowered at use of the corresponding computing circuit of data processing.

Existing computing methods share on two classes: 1) methods parametrical estimations, applied to models of constant structure; 2) methods of structural - parametrical identification (SPI), changing structure of an initial model - hypothesis. The most perspective represent methods SPI, including at a stage estimations the orthogonal circuit Hausholder. An offered new method SPI - a method step ortogonalizing (MSO), possessing properties of a method of step-by-step regress (SR) to remove from structure one of duplicating variables on one of stages of processing, possesses an additional opportunity to keep a lot of estimated parameters, than in method SR. The developed system parametrical estimations (SPE) contains besides MSO methods of plural and step-by-step regress, a characteristic root both other methods and the procedures deciding problem MLS.

Research of application of a method SPI for descriptive models it was carried out by the example of processing radiointerferometer observations of outside of galactic sources. For processing have been taken 10 samples by volume from 9 up to 20 observations. It was supposed to determine values of 5 factors at regressors. It has been executed estimations the parameters of model by a method of step-by-step regress which now is the most widespread method of definition of structure of model in problems of definition of differential amendments. The analysis of correlation matrixes specified samples allows to make a conclusion about presence of effect of interdependence of parameters, and also presence statistically insignificant composed in models.

With the help of a method of step-by-step regress was the models containing from 2 up to 3 estimations of significant parameters from 5 possible are received. With the help of a method SPI it was possible to receive up to 5 estimations for different samples. For comparison the general estimations of parameters the models received by various methods of processing, including plural and step-by-step regress have been considered. Estimations received by various methods for noncorrelation, significant composed (1 of 5 possible) practically coincide.

Further the analysis of methods step estimations with the help of external measures of quality of model has been executed. The applied external measures are based on the analysis of divergences between the forecast and known value of parameter of model. Value of the regular mistake caused in a general sense by a method of processing, and a random error for the considered methods of processing was compared. The received values of mistakes for researched methods are sizes of one order. It is necessary to note, that various strategy SPI have allowed to estimate the greater number перфессоров and, opportunities having lowered effect of multicollination.

Thus, results of processing the radiointerferometer observations by competing methods allow to draw a conclusion on high efficiency of a method SPI.
ABOUT DEVELOPMENT OF INTELLECTUAL SYSTEM OF DATA PROCESSING OF INDIRECT EXPERIMENTS. S.G. Valeev, D.M. Yastrebov.
The Ulyanovsk State Technical University, st. Severny venez, 32, Ulyanovsk, russia

One of the most important computing procedures at processing indirect supervision (the data of passive experiment) is the stage of definition (estimation) parameters of mathematical models.

Unfortunately, the traditional approach in the specified areas to estimate the parameters, providing rigidly fixed model and application of a method of the least squares, does not correspond to growing requirements of practice. Separate attempts to leave from frameworks standard principle of the least squares (PLS) are aimed at the decision of private problems and do not provide the system approach to a problem.

As alternative S.G.Valeev's to traditional approach the methodology regression the modeling, providing in problems estimate regression the analysis, check of assumptions, adaptation has been offered in case of their infringements in this or that sequence and supposing presence of the special software - systems of processing of the information, automate process of calculation and the analysis. Regression modeling (R-) is an adaptive system approach, at which correctness of application of any element of system (sample, model, a method estimate parameters, a method estimate structures, the measure of quality, a set of assumptions) can be subjected to doubt and check with corresponding adaptation under the set script at infringement of the set conditions. The basic problem thus is creation of the corresponding software.

Regression analysis (R-) is one of the most productive methods of the mathematical statistics, allowing to describe multivariate results of supervision. In regression the analysis exists three the basic precisely allocated a stage:

1) The initial mathematical description of results of supervision,
2) Estimating parameters of mathematical model on the basis of a principle of the least squares,
3) Search and a choice of optimum model by criteria of quality.

For today there is a significant set of computing receptions and the means allowing successfully to work in conditions of infringement of separate assumptions. However it is clear, that the problem cannot be resolved by partial updating. Strategy or methodology of the approach is necessary for its decision to the account whenever possible all consequences of non-observance of conditions of the basic assumptions regression the analysis.

Now there are two approaches to realization RA. Both approaches admits, that those or other assumptions are broken and in the pure state the normal circuit to apply it is necessary with care. The methodology of the first approach provides check of assumptions and in case of their infringements use of corresponding adjusting procedures. In the second case (without check of preconditions) or other method of statistical processing (dispersive, covariation analysis, etc.) is applied, or algorithm RA which is taking into account infringements at once of several preconditions is designed. Thus it is considered, that other assumptions are observed.

Within the framework of development and program realization of the first approach in 2001 the program complex system of searching for optimum regression (SSOR) of version 1.0 has been created.

In connection with a wide circulation recently personal computers and platform Windows there is a question on creation of the specialized, in detail focused package SSOR on this platform. The software package available now the SSOR 1.0 realizing the approach regression of modeling (R-
approach) to the decision of problems PLS, demands the serious updating including: 1) translation into platform Windows; 2) development of base of functions and knowledge; 3) processing and expansions of functional filling; 4) development of language of tasks for managements and formations of scripts of processing; 5) improvement of the interface and means of development according to last achievements in sphere of information technologies.

As the adaptive \( R_- \)-approach assumes performance of procedures parametrical estimation and structurally-parametrical identification, diagnostics of essential infringement of conditions regression analysis PLS (RA-PLS) and application of adaptive procedures at their detection, the functional block includes programs: plural regress, libraries of criteria of quality of model and diagnostics of infringements, ridge regression, methods of the characteristic root, the compressed estimations, the step-by-step regress, casual search, etc.

The package SSOR of version 2.0 is intended first of all for reception of optimum models of the data processing used for the forecast and wide application can find at the decision of problems PLS (problems of restoration of dependences) on superfluous indirect supervision in astronomy, heavenly mechanics, geo-and planet- physics and other areas.

In the near future SSOR of version 2.0 is planned to include the managing program in a package on adaptation to infringements \( R_- \). The given program will allow to automate process of adaptation to infringements of assumptions regression the analysis, reducing corresponding time and material inputs.

Background: A variety of landforms indicates the possible existence of past or present ice in the near subsurface of Mars [1-3]. Among the most spectacular ice-related features are lobate debris aprons (LDA). They have been interpreted to be a mixture of rock particles and interstitial ice [4-5] analogous to terrestrial rock glaciers (debris transport systems comprising a creeping mixture of rock fragments and segregated or interstitial ice [6]). Rock glaciers are sensitive indicators for the climatic environment during their formation and - if present on Mars - are thought to be possible, large and easily accessible water reservoirs. The analogy between terrestrial rock glaciers and Martian LDA is predominantly based upon the shape of debris aprons, their surface texture, their relationship to adjacent regions with permafrost-related morphologies, and the correlation of their global distribution with the predicted stability of Martian ground ice. LDA are known from the dichotomy boundary (i.e., Tempe Terra (TT), Deuteronilus/Protonilus Mensae (DM/PM), southern Elysium Planitia (EP)) and from regions surrounding the large southern hemispheric impact basins. So far, detailed classification and morphometric work has been carried out in the DM/PM [7-9] and the TT area [7-8,10] and the Hellas region [11,12].

This work focuses on the distribution of LDA in the circum-Hellas Planitia region and compares the morphometry derived from earlier measurements on the basis of MOLA topography in combination with Viking image data with image and topographic data obtained through the High Resolution Stereo Camera (HRSC).

HRSC Data: The HRSC onboard the ESA spacecraft Mars Express has obtained a dense mapping coverage of the Argyre and Hellas Planitiae regions with an image resolution of at least 25 metres per pixel in the panchromatic channel. In 35 orbits between 01/09/2004 (orbit 8) and 06/28/2004 (orbit 561) the HRSC instrument covered an area of approximately $4 \times 10^6$ km$^2$ between 55°E - 120°E and -60°N - -20°N (s. fig. 2). Although southern-hemispheric dust obscures a considerable fraction of the image data obtained, the majority of images has such a good quality, that textural details can be examined and photogrammetric processing of the data allows generation of high-resolution digital terrain models.

The large image strips of the HRSC instrument not only allow us to analyse small-scaled features but it also enables us to nest highest-resolution MOC and THEMIS image data into a much broader context than before. Because of that the quality of mapping of morphologic/geologic and/or textural units has become much more efficient. Furthermore, the colour channels of the HRSC instrument help us to distinguish between surficial units of different compositions. This enables us to differentiate between various textures related to debris material and bedrock across several remnant units.

With the help of the newly available data, we have re-complied our measurements and show first results focussed on morphometry, surface texture mapping, and comparison of the combination of multiple data sets.

Morphometry and First Results

As described in detail in [7] the morphometry of LDA and isolated remnants and knobs as their source regions has been obtained by measurements on the basis of the MOLA derived digital terrain models with a scale of 463 metres per pixel and individual MOLA profiles. For 25 objects in the Hellas Planitia region we derived the planimetric area of aprons and remnants, volume of aprons and remnants (based upon the assumption that the underlying bedrock is not inclined), and thickness of aprons. In order to minimize systematic errors
in our measurements we preferred to compare volume and area ratios instead of using concrete figures. The results of previous MOLA based measurements have been confirmed by HRSC derived topographic data and additional remnant–apron features have been identified and measured in order to close the gap in the small–scaled feature range.

The area ratios between debris aprons and remnants obtained so far from the global population of lobate debris aprons are in very good accordance with data obtained by [6] for terrestrial rock glaciers, which are in a range of 1.8 to 4.5 (s. fig. 1). Other workers in this field proposed values in the range of 1.36 [13] to 4.4 [14].

As suggested by [6] the ratio is primarily controlled by bedrock resistance, relief and climate. As values for the relief have been taken into account through topographic data analyses and due to the fact that climatic variations take place on a global scale, differences in bedrock material and ages for formation are most probably the main reasons for variations in the morphometric ratios obtained. Apart from that the results obtained so far are comparable and although different mechanisms for formation can be proposed (dichotomy-related/impact related), individual regions have obviously undergone similar stages of development.

A MODEL FOR POLARIMETRIC AND PHOTOMETRIC CHARACTERISTICS OF THE MOON AT MODERATE PHASE ANGLES BASED ON REALISTIC ASSUMPTIONS ON REGOLITH MICROSTRUCTURE. Yu. I. Velikodsky, V. V. Korokhin, and L. A. Akimov. Astronomical Institute of Kharkov University. Sumskaya Ul., 35, Kharkov, 61022, Ukraine. E-mail: velikodsky@astron.kharkov.ua

There are known two potential mechanisms for explaining positive polarization that is observed for atmosphereless celestial bodies at large phase angles [1]. The first one, Fresnel's reflection, may take place due to existing of large (in comparison with wavelength) smooth surfaces in regolith (glasses). The second one, Rayleigh's scattering, supposes the presence of particles with sizes smaller than wavelength, i.e. with submicron sizes. It is known that typical size of regolith particles is about tens of micrometers. The particles are mainly aggregates of smaller grains of micron- submicron sizes. Such particles cannot produce a pure Rayleigh polarization. On the other hand, Fresnel's reflection yields too large phase angles of the positive polarization maximum and cannot explain this effect by oneself.

Taking into account that the principal contribution to scattering (and, hence, to positive polarization) may produce small particles with subwavelength sizes we propose a combinative heuristic model using the Rayleigh-Gans approximation [2], Fresnel's reflection on large surfaces, shadow-hiding effect, and multiple scattering on micro- and macroscales. The model shows a good agreement with observational data for the Moon for both brightness and polarization degree phase dependences (Figs. 1 and 2) with the same set of parameters. Also the model explains Umov's law and decreasing of the phase angle of polarization maximum with albedo increasing, which is observed for the Moon and asteroids.

Phase dependence of brightness was approximated at moderate phase angles (Fig. 2), where a good approximation with exponential function exists [3]. At large phase angles the model yields higher values of brightness than the exponent. It is known that phase dependence at large phase angles differs from exponent [3], and our model can explain this difference. To check this we plan to study phase dependence at large phase angles by absolute observations and investigation of mare-highlands contrast. These and related results are available at our site: http://www.univer.kharkov.ua/astron/dslpp/moon/polar/.


Fig.1. Approximation of lunar observational polarization data (A - diffuse albedo). Negative polarization effect at small phase angles is not considered.

Fig.2. Two approximation curves of phase dependence of lunar brightness (at area with mirror reflectance geometry): a model curve and exponential function (proposed in [3]). Approximation was performed at phase angles 35-45°. Opposition effect at small phase angles is not considered.
Introduction: From HRSC observations of image data of orbit 18 near eastern Valles Marineris, a wide range of high- and lowland morphologies have been observed and analyzed in detail. The rugged morphology of the features resembles that of terrestrial landforms dissected and eroded by water. Characteristic morphologies are defined by small buttes, angular mesas and debris aprons at varying elevation levels. The valley floors are situated at an elevation level of approximately -5100 m, whereas the relief of mesas varies between several hundred to a maximum value of more than 2500 m. The smooth surface texture and colour properties of the flat-topped mesas and surrounding highlands suggest similar ages. It has to be verified whether the different elevation levels of the mesas are (1) due to surface removal and subsequent abrasion, or whether (2) they are due to removal of subsurface material, block tilting or lowering of the remnant rocks. We carried out measurements to obtain crater size-frequency distributions in order to determine crater retention ages for corresponding areas. Furthermore, small scale height measurements were carried out by the use of a stereo comparator. The colour characteristics of varying geological and geomorphological units will be analyzed.

In particular, the bright material on the debris aprons of the mountains, on the rims of several mesas, and on the valley floors will be analyzed in terms of spectral characteristics and lateral distribution. The major goal of this procedure is to constrain the main erosional processes and the evolution of the characteristic terrains, and to understand the subsurface structure of the area.

Method: Heights of distinct mesas have been obtained from measuring parallaxes between HRSC’s high resolution nadir and a stereo channel with a stereo comparator. This method allows to find height values for small features which are below the spatial resolution limit of the digital terrain model derived from HRSC processing. In order to derive height information, parallax shifts have been calibrated using the MOLA DTM. Heights range from -5100 m to 2500 m. Crater size-frequency distributions have been measured at three different height levels in Hydraotes Chaos. Then, the cratering chronology model by Hartmann & Neukum [1] was applied in order to obtain absolute cratering model ages.

Results: In top level mesa and highland plains of Xanthe Terra the cratering measurements reveal a long-lasting process of erosion which may have ceased 1.7 Ga ago or has slowed down considerably but continued maybe thereafter at a much lower level. Nevertheless it is not clear how long this second episode of erosion was going on due to lack of image resolution.

In what is now the Hydraotes Chaos area, the erosional process was much more effective and eroded the top level Xanthe Terra basaltic plains unit down to different levels in several episodes, thus creating mesa tops at several different intermediate levels.

The process of degradation slowed down considerably 400-200 Ma ago, and most of the former mesas and highland units had by then been eroded more or less down to the valley floor level.

These findings most likely mean that over a long period starting 3.7 Ga ago and ending sometime between 400 and 200 Ma ago, the valley area of Hydraotes Chaos first was formed by the action of flowing water because the morphology as now seen in the images at high resolution, colour and 3D, cannot be explained by any other process. Then, subsequent erosional processes do not seem to have been absolutely continuous, but the different mesa levels and erosional epochs showing up in the age measurements suggest that we deal with at least three extended episodes of erosional activity.

NEW INSIGHTS INTO THE EVOLUTIONARY HISTORY OF THE MAJOR VOLCANIC CONSTRUCTS FROM MARS EXPRESS HRSC DATA. S. C. Werner¹, G. Neukum¹, and the HRSC Co-Investigator Team, ¹ Institut fuer Geologische Wissenschaften, Freie Universitaet Berlin, Malteserstr. 74-100, Bldg. D, 12249 Berlin, Germany. (swerner@zedat.fu-berlin.de).

Introduction: During the first half year of the ESA Mars Express mission in orbit, the High Resolution Stereo Camera, a multiple line scanner instrument, is acquiring high-resolution colour and stereo images of the surface of Mars[1]. Resolution down to 10 meters per pixel coupled with large areal extent (swaths typically 65-100 km wide and thousands of km long) means that small details can be placed in a much broader context than was previously possible. Most of the major volcanic constructs have been covered in the first half year of the mission. The ability to image in colour and stereo simultaneously gives us new opportunity to better characterize most of the volcanoes in the Tharsis and Elysium region and some highland volcanoes geomorphologically and chronostatigraphically. We have remapped major parts of the volcanic shields and calderas on the basis of the high-resolution (as good as 10 m/pixel) HRSC imagery in colour and stereo and in combination with nested MOC imagery[2] and the Super Resolution Channel (SRC) (as good as 2.5 m/pixel) of the HRSC.

Method: To determine absolute ages on Mars we measure the crater size frequency distribution for a geomorphologically mapped unit and fit the crater production function [3,4] to the data set, extract a size-frequency value for craters of one kilometer and larger, and apply the Hartmann/Neukum chronology model[5] for the derivation of an absolute age.

Figure 1 a) and b) show Hecates and Albor Tholus observed in Orbit 32. Both belong to the Elysium region, one of the smaller volcanic bulges in the northern lowlands. The caldera morphologies indicate step-wise volcanic activity and the ages derived from crater size-frequency measurements yield a period of activity over 2 billion years for Albor Tholus. For Hecates Tholus the recorded ages range between 1 billion and 100 million years.

Figure 1 c) and d) show two of the three Tharsis Montes, belonging to the largest volcanic region of Mars: Arsia Mons (c), covered in Orbit 263, is the southernmost construct of that unit. It is characterized by a single caldera floor of an age of about 130 Ma. Small volcanic domes in the Arsia Mons caldera, following the major fault direction of the Tharsis Montes group, hint to possibly still active volcanism. Ascræus Mons (d) has been covered in Orbit 68 (and Orbit 16) and is the northernmost of the volcano triplet. Its caldera is represented by a number of floor levels indicating again repeated eruption activity in the last one billion years. Possibly the caldera V has been formed much earlier in the life time of the volcano.

In figure 1 e) the caldera of the largest Martian volcano Olympus Mons is shown. The caldera unit has been covered in Orbit 37. Again the caldera floor level and morphology indicate repeated eruption activity. The age measurements yield an activity phase between 100 and 200 million years ago. Crater frequencies measured on the different caldera floors indicate ages, which slightly deviate from the morphologically expected time-stratigraphic derived volcanic sequence for the formation of the calderas. A detailed study of the caldera morphology show for the larger caldera floors tectonic and volcanic resurfacing affecting the age results[6].

Results: Crater size-frequency measurements confirm that the edifices have been constructed over billions of years[7] and are characterized by episodically repeated phases of activity[8] continuing almost to the present. The youngest ages determined by the crater size-frequency measurements are about 2 Ma suggesting that the volcanoes are potentially still active today. A number of caldera floor ages cluster around 150 Ma indicating a relatively recent peak activity period and practically coinciding in age with radiometrically measured crystallization ages of a group of basaltic meteorites from Mars (SNC meteorites)[9].

GAMMA RAYS FROM MAJOR ELEMENTS BY THERMAL NEUTRON CAPTURE REACTIONS: EXPERIMENT AND SIMULATION FOR PLANETARY GAMMA-RAY SPECTROSCOPY.

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Introduction: The elemental composition of planetary surface can be determined by remotely measuring the energies and intensities of gamma rays that leak out of the surface. Gamma rays are emitted from the surface nuclei which are excited either naturally or by the exposure to cosmic rays and the subsequent interaction of the secondary neutrons [1].

In order to calibrate the gamma-ray intensities to elemental abundance on the surface, one needs to know the production rate of line gamma rays, which is a function of the abundance of element of gamma-ray origin as well as neutron fluxes. The continuum in the energy spectrum is also important to simulate for the estimation of the noise level, counting rate, and detection threshold. Only computer simulation can provide such information and the precision of an observation rely on it. Therefore, it is essential to verify the computer simulation with ground experiments for high-precision missions such as SELENE [2], MESSENGER and BepiColombo [3]. In this work, production and detection of gamma rays from major elements by thermal neutron capture are simulated experimentally and numerically, and the results are discussed.

Experiments: Thermal neutron beams were irradiated to samples at Japan Research Reactor No.3 (JRR-3) at Japan Atomic Energy Research Institute (JAERI) [4]. With nominal flux of ~10^7 /cm^2 sec and radiated to samples at Japan Research Reactor No.3 through thermal neutron transport to energy rays [5]. The first calculation is conducted thorough from thermal neutron transport to energy.

Calculations: The computer simulations need to reproduce variety of interaction and complex geometry of a detector system in a unified manner. For this reason a Monte Carlo computer simulation code Geant4 release 6.2 was chosen to calculate production, transport, and reaction of neutrons and gamma rays [5]. The first calculation is conducted thoroughly from thermal neutron transport to energy deposition of gamma rays in the Ge crystal using the hadronic physics list “QGSP_HP”. Fig. 1 shows a calculated energy spectrum of gamma rays emitted from the Fe target, together with the one by the Ge detector in the experiment. The spectrum by the calculation has few discrete gamma-ray peaks, and it has a continuum while the Fe target is ~0.11 cm thick. The same tendency has been found in other targets and in calculations by MCNPX. Note that in Fig.1, the emission spectrum from the target is shown as a calculation result in order to emphasize a serious difference from the experiment, even though the Geant4 code can provide a spectrum based on energy deposit of gamma rays in a Ge crystal. It is clear that there is a discrepancy in calculating gamma-ray production rates.

Evason of the problem: Counting rate of a certain gamma-ray peak observed by a spectrometer can be expressed as

\[ C = \left( \frac{M}{A} \right) \cdot N_A \cdot b \cdot \phi \cdot \sigma \cdot \eta(E_p) \cdot \epsilon(E_p) \]

where M is the mass of the target material exposed to neutron flux, A is the atomic number of the element of interest, N_A is the Avogadro number, b is the isotope ratio, \( \phi \) is the thermal neutron flux, \( \sigma \) is the cross section of thermal neutron capture reaction, \( \eta(E_p) \) is the branching ratio to a channel of \( E_p \), and \( \epsilon \) is the absolute peak detection efficiency. In the computer simulation, capture reaction rates can be calculated properly. By applying the branching ratios to the capture reaction rate outside of the simulation code, it is possible to retrieve the correct gamma-ray production rates. Then the calculation can be restarted by producing gamma rays and let them trans-
port and react, with initial energies determined by the outside multiplication and initial locations where neutron capture reaction took place. The initial directions of produced gamma rays are assumed to be isotropic.

Results and Discussions: Calculated peak gamma-ray intensities observed by the Ge detector are compared with experimental results after employing the evasion method described above. The ratios of experimental results to calculations are shown in Fig. 2, where the major peaks from Fe and Al targets are represented. The ratios are normalized at 352 keV for Fe and 1779 keV for Al because of a neutron flux ambiguity. The error bars are attributed to the experiments. As can be seen from the figure, peak gamma-ray intensities observed in the beam experiments are well reproduced by the calculation method. The ratios of the experimental results to the calculation are 100.0% at 1260 keV, 99.7% at 1613 keV, and 98.3% at 4218 keV for the Fe target, and 100.0% at 4734 keV, 99.4% at 2590 keV, and 99.1% at 983 keV for the Al target. Most of other strong peaks are consistent within the error of ~8%, except for the higher energy doublets from Fe at 7631 and 7646 keV, where more extensive analyses of experimental data are required due to the interference with each other. Even though this method is an indirect way and does not satisfy the requirement of simulating gamma-ray spectroscopy in a unified manner, the method is confirmed to be reliable to reproduce gamma-ray peaks from thermal neutron capture reactions. Comparison of other major elements should be made by using other targets, including stone targets, whose elemental abundances are more resemble with those on the planetary surfaces.

**Introduction.** All Galilean satellites are in synchronous rotation; their orbits are nearly circular and lie in the equatorial plane of Jupiter. Io is the large satellite closest to Jupiter. Therefore, the influence of the Jupiter’s tidal potential on the equilibrium figure and gravitational field of Io is appreciably stronger than it is on the remaining large satellites. For the theory of Io’s figure to be consistent with currently available observational data, it must include effects of the second order of the small parameter [1].

\[
\alpha = \frac{3\pi}{G \rho_0 \alpha^2} = \frac{\omega^2 s_1^3}{G m_0} = \left( \frac{M}{m_0} \frac{R}{s_1} \right)^3 = 171.37 \times 10^{-5}, \tag{1}
\]

where \( \rho_0 \), \( \tau \), \( m_0 \), and \( s_1 \) are the mean density, rotation period, mass, and mean radius of Io, respectively, \( G \) is the gravitational constant, \( M \) is the mass of Jupiter, \( R \) is the radius of the satellite orbit. The numerical value in (1) was obtained using data from Table 1.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Io</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orbital radius</td>
<td>421.6</td>
</tr>
<tr>
<td>Period, days</td>
<td>1.769</td>
</tr>
<tr>
<td>( s_1 ), km</td>
<td>1821.6 ± 0.5</td>
</tr>
<tr>
<td>( m_0 ), ( 10^{23} ) g</td>
<td>893.2</td>
</tr>
<tr>
<td>( \rho_0 ), ( g ) cm(^{-3} )</td>
<td>3.5278 ± 0.0029</td>
</tr>
<tr>
<td>( g_0 ), cm s(^{-2} )</td>
<td>179</td>
</tr>
<tr>
<td>( \alpha = \frac{3\pi}{G \rho_0 \alpha^2} ), ( 10^{-5} )</td>
<td>171.37</td>
</tr>
<tr>
<td>( \frac{C}{m_0 s_1^2} )</td>
<td>0.37685 ± 0.00035</td>
</tr>
<tr>
<td>( J_2 ), ( 10^{-6} )</td>
<td>1845.9 ± 4.2</td>
</tr>
<tr>
<td>( C_{22} ), ( 10^{-6} )</td>
<td>553.7 ± 1.2</td>
</tr>
<tr>
<td>Jupiter’s mass ( M ), ( 10^{30} ) g</td>
<td>1.897</td>
</tr>
</tbody>
</table>

**Table 1.** Observational data and the model parameters for Io

**Principal formulas.** A remarkable achievement of the successful Galileo mission was the determination of the first coefficients in the expansion of the gravitational field in terms of spherical functions for the Galilean satellites and the proof that their figures are actually in hydrostatic equilibrium [2]. The Love number \( k_2 \) in [2] was found by using the observed value of gravitational moment \( C_{22} = (553.7 ± 1.2) \times 10^{-6} \):

\[
k_2 = 4(C_2/\alpha) = 1.2924 ± 0.0027 \tag{2}
\]

To determine the dimensionless moment of inertia Anderson et al., [2] used a formula valid in the Radau approximation for a satellite in hydrostatic equilibrium from the book by Jeffreys [3]

\[
\frac{C}{M a_1^2} = \frac{2}{3} \left[ 1 - \frac{2}{5} \sqrt{\frac{4 - k_2}{1 + k_2}} \right] = \frac{0.37685 ± 0.00035}{3}. \tag{3}
\]

Let us now explain the meaning of formula (3), which was used by Anderson et al., (2001) to constrain the moment of inertia of Io. Recall how this formula was derived. For the equilibrium figure of a rotating planet or satellite in the Rado approximation, the Radau-Darwin formula is valid (see, e.g., Zharkov and Trubitsyn 1980; formula (32.20))

\[
\frac{C}{M a_1^2} = \frac{2}{3} \left[ 1 - \frac{2}{5} \sqrt{\frac{5 \alpha}{3 J_2 + \alpha - 1}} \right], \tag{4}
\]

where \( C \) is the polar moment of inertia, \( M \) is the mass, \( a_1 \) is the equatorial radius, \( J_2 \) is the quadrupole moment, and \( \alpha \) (1) is the small parameter of the rotating equilibrium planet. Zharkov et al. (1985) showed that for a synchronously rotating equilibrium satellite in the field of the tidal potential in the first approximation, the equilibrium quadrupole moment \( J_2 \) of the body under consideration is the sum of the part attributable to the tidal potential \( J_{2t} = 0.5/ak_2 \) and the part attributable to centrifugal potential \( J_{2c} = 1/3ak_2 \):

\[
J_2 = J_{2t} + J_{2c} = \frac{5}{6} \alpha k_2. \tag{5}
\]

Formula (3) is obtained if we substitute not \( J_2 \) (5) but only the part of \( J_2 \), more specifically, \( J_{2t} \) into (3). Thus, it refers to an equilibrium rotating planet or satellite, i.e., to a similar but not the same problem studied here. Therefore, it would be natural to use the
Love number $k_2$ or $h_2 = 1 + k_2$ as a constrain when modeling any Galilean satellite.

**Conclusions.** In the report there are considered two (Fe-FeS core + silicate mantle) and three (Fe-FeS core + silicate mantle + crust) layers models of Galilean satellite Io. Two parameters $\rho_0$ – average density and the Love number $k_2$ for equilibrium figure of satellite are known. With help of theory of figure formulas there were obtained the principle moments of inertia $A$, $B$, $C$ and the mean moment of inertia $I$ for two and three layers models of Io using only $\rho_0$ and $k_2$ as boundary conditions. We conclude that when modeling the internal structure of Io, it is better to use the observed value $k_2$ rather than the moment of inertia derived from $k_2$ with help of the Radau-Darwin formula.

We calculated the periods of the Chandler wobble for considered models. This period is equal approximately to 460 days for three layers model.

**Acknowledgements.** This work was supported in part by the Russian Foundation for Basic Research (project no. 03-02-16195), The program no.13 of the Presidium of RAS, and by The Russian Science Support Foundation (http://www.science-support.ru/).

DEVELOPMENT OF FLUVIAL ACTIVITY AT DAO VALLIS, NIGER VALLIS, AND HADRIACA PATERA, MARS. W. Zuschneid1, G. Neukum1, S.C. Werner1, R. Greeley2, D. Williams2, and the HRSC Co-I Team, 1Freie Universitaet Berlin, Institute for Geosciences, Germany (ewill@zedat.fu-berlin.de), 2Arizona State University, Department of Geological Sciences

Introduction: The north-eastern Hellas Basin rim has been modified by volcanic activity at Hesperia Planum, and in particular from the volcanoes Tyrrhena and Hadriaca Patera[1-2]. The southern and eastern boundaries of Hadriaca Patera are cut by the Dao and Niger Valles outflow channel system. The formation of the outflow channels has been attributed to the melting of ground ice by volcanic intrusions and the catastrophic discharge of meltwater[3]. This discharge occurred on the surface or sub-surficially, producing subsided plains. On the basis of the newly available data from the High Resolution Stereo Camera (HRSC) on the ESA Mars Express, ages of the flanks of Hadriaca Patera and the outflow channels will be determined to provide constraints on the evolution of volcanism and fluvial activity in the eastern Hellas region.

Geologic Setting: The geologic evolution of the eastern Hellas Basin is dominated by large-scale volcanic activity. This activity has led to the formation of the volcanic plains of Hesperia Planum, but it is also associated with central-vent volcanism at the Tyrrhena and Hadriaca Paterae. Hadriaca Patera, located at the northeastern rim of Hellas, is a large, low relief volcanic construct with a radial drainage pattern. Dao and Niger Vallis form a system of outflow channels located directly at the southern margin of Hadriaca Patera. The outflow system was created by a combination of surface and subsurface discharge[3-4]. The latter has led to the formation of subsided plains due to the loss of material under a rigid surface layer.

Dao Vallis source depression: Dao Vallis originates from an extensive horseshoe-shaped source depression, 120km in length, up to 40km wide and up to 2700m deep. It shows partly smooth, partly chaotic floor material. Downstream, a second, smaller depression with similar morphology follows [4].

Niger Vallis: Niger Vallis has a far more complex morphology than Dao. It consists of a variety of depressions and connecting channels. Several channels were formed by surface flow, while others were created by removal of material underneath a brittle surface which subsequently fractured and filled the void [3]. These subsurface flow features connect Niger with Dao Vallis and the outflow system to Ausonia and Peraea Cavus in the east of Hadriaca. The larger depressions of Niger Vallis have different depths, their floors are smooth and show few chaotic areas.

Current Work: The evolution of the outflow channels and the volcanic features cannot be viewed independently. In order to examine the relationships of fluvial and volcanic activity, the different geologic units have been mapped. Images of the HRSC provide excellent coverage of the entire area at resolutions of 12.5 to 25m/pixel and stereo data. On the basis these images, crater retention ages [6-8]] will be measured. These age measurements provide a basis for establishing a chronology of events for the Dao Vallis and Niger Vallis region.


Figure 1: Overview of Dao and Niger Valles and Hadriaca Patera. Images of this mosaic were acquired by HRSC during orbit 528 and 550
THERMAL RADIATION IN THE LOWER VENUS ATMOSPHERE: THE EFFECT OF THE COLLISIONAL LINE BROADENING. T.S.Afanasenko¹, A. V. Rodin¹, Institute for Space Research, RAS, Moscow, 117997, Russia

rodin@irn.iki.rssi.ru.

Introduction: The radiation balance of the atmosphere of Venus, a planet known for its strong greenhouse effect, is complex and still not completely understood. In particular, there are open questions concerning the impact of the dynamical heat transfer onto the thermal structure of the undercloud atmosphere, on the altitude range where convective zones occur and how stable they are, how the energy cascade is organized between zones where solar light is being absorbed, the planetary surface, and the outgoing thermal radiation. The necessity of the detailed consideration of radiation processes in the lower Venus atmosphere is also imposed by the development of the GCM modeling, as well as by paleoclimate research.

Simulation of the thermal radiation transfer in the lower Venus atmosphere is further complicated by the fact that in the optically thick medium, a major portion of energy is passed within transparency windows in between strong CO₂ absorption bands. The molecular absorption in these intervals is determined mainly by the far wings of spectral lines. Their form-factor may substantially differ from their classical Lorenzian shape providing large offset from the line center. This led authors to rely on the empirical model of the line shape[1] which relates to a narrow pressure and temperature range, moreover, its implication in energy flux calculations has not been proved by experiment. Thus the need for a physically grounded model of the line shape related to high pressure and temperature arises.

The model: In order to evaluate a physically plausible form-factor, we have employ a theory based on the approximation of the far wing and described in detail in [2]. The profile is divided by three intervals in addition to the central Lorenzian part, each interval being dominated by some particular component of the intramolecular interaction potential. In terms of the line shape, this results in a particular rate of vanishing of the line wings, generally following power law. Amplitude of the wing is proportional to squared pressure and therefore is indistinguishable from the pressure-induced absorption. In the case when the central part appears to be lower in amplitude than the neighbouring “wing”, an interpolation procedure between cetral part and the nearest “wing” interval was applied in combination with the renormalization of the full profile, so that the intergral line strength is preserved. A schematic of the model line shape is presented in Figure 1. Evidently, as high as at 60 km above the surface, at sufficiently far offsets the Lorenzian profile may exceed the actual absorption by several orders of magnitude.

Results: We have calculated the spectral absorption for the interval 1-200 micron corresponding to the range of altitudes up to 100 km and calculated spectral fluxes solving the radiation transfer problem in two-stream approximation. A standard atmospheric composition was adopted, with variable water vapor and other minor constituents. Cloud layer was assumed to be an absorbing medium with spectral

Figure 1. The model line shape corresponding to 60 km altitude. Central peak is the Lorenzian component, whereas dashed, dotted and dash-dotted curves correspond to different components of the collisional profile in the far-wing approximation.

properties compiled from available observations. Integration over wavelengths yields the net radiation flux that determines the radiation balance of the planet. The results are presented in Figure 2 for various water vapor contents. In the lower part the flux appears to be in better agreement with observations than previous calculations not using a physical line shape[3,4]. The flux at the level of Venus surface equals approximately 2.7 W/m².

Figure 2. Thermal radiation flux calculated for various contents of water vapor: (dashed) 10 ppm, (solid) 30 ppm, and (dotted) 100 ppm. An inflection at 45-50 km is caused by absorption of the outgoing radiation at the lower boundary of the cloud deck.

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INTERIOR STRUCTURE OF MARS. T.V. Gudkova and V. N. Zharkov. Institute of Physics of the Earth, Russian Academy of Sciences, B.Gruzinskaya, 10, Moscow 123995, Russia, gudkova@ifz.ru

Introduction. Based on available chemical models of the planet [1,2,3,4,5], a new set of global models of the Martian interior has been constructed. The model comprises four submodels - a model of the outer porous layer, a model of the crust, a model of the mantle and a model of the core. The first 10-11 km layer is considered as an averaged transition from regolith to consolidated rock. The mineral composition of the crustal basaltic rock varies with depth because of the gabbro-eclogite phase transition. As a starting point for mantle modeling there have been used experimental data obtained by Bertka and Fei [6,7] along the areotherm, iron content of the mantle being varied. New high P-T measurements of the density of Fe (γ-Fe), FeS and FeH enable us to refine the core model. Taking into account available chemical models and the fact that noticeable amount of hydrogen could enter the Martian core during its formation [8], such parameters as ferric number of the mantle (Fe#), sulfur and hydrogen content in the core are varied.

The construction of Martian interior structure models.

Crust. In the crust models of Babeiko and Zharkov [9], there is a transition in the outermost 10- km layer from highly porous Martian regolith (\( \approx 1.6 \) g/cm\(^3\)) to consolidated rocks (3.2 g/cm\(^3\)). Because of the gabbro-eclogite phase transition, the increase of density with depth in the consolidated crust depends strongly on the temperature gradient.

Mantle. For the modeling of the density profiles in the mantle, we use the experimental results of Bertka and Fei [6,7]. They have performed high-pressure multi-anvil experiments with an analog of the Dreibus and Wänke composition [10] to determine the model mineralogy up to core-mantle boundary pressures along a model areotherm. The Martian mantle are assumed to consist of 12 mineral assemblages. The weight fraction of each mineral assemblage is calculated from the mass balance of the experimental products. We added \( \delta_p \) to the B-F mantle density profile (Fe#25), when calculating the models with lower and higher iron content. When increasing Fe# by 1 \( \delta_p \) is increased by about 0.01 g/cm\(^3\) for olivine zone, 0.0083 g/cm\(^3\) for \( \beta \)-zone, 0.011 g/cm\(^3\) for \( \gamma \)-zone and 0.0125 g/cm\(^3\) for perovskite zone.

Core. The Martian core composition is considered to be sulfur-rich, consisting of Fe with 14.2 wt % S, 7.6 wt % Ni [10]. New high P-T measurements of the density of Fe (γ-Fe) and FeS [11] enable us to refine the core model by Zharkov [8]. In this study the Martian core is assumed to be a mixture of iron-nickel alloy, sulfur and some amount of hydrogen. The addition of 10 mol % of hydrogen to the iron reduces its density by 0.16 g/cm\(^3\) [8]. Experimental data of high-P-T phases of γ-Fe and FeS [11] were obtained for a solid state and the temperatures of about 1300-1600K. When the temperature is increasing from 1600K to 2100K, the density decreases by about 0.125 g/cm\(^3\). When melting core material, the density decreases by about 0.2-0.3 g/cm\(^3\).

Modeling. In our modeling we varied the following parameters: ferric number of the mantle (Fe#), sulfur content in the core (S\(_{\text{core}}\)) and hydrogen content in the core (H\(_{\text{core}}\)). Core mass (M\(_{\text{core}}\)), core radius (R\(_{\text{core}}\)), pressure at the core-mantle boundary, crust thickness, dimensionless value for the moment of inertia, calculated bulk Fe content, weight Fe/Si ratio and the thickness of a perovskite layer for the calculated models are listed in Table.

Figure 1 shows the core radius as a function of the Martian mantle Fe# for different amount of hydrogen in the core (0-70 mol %), assuming a core composition of 14 wt % S (according to the DW model) and a 50-km-thick crust. If there is no hydrogen in the core, the Fe/Si ratio ranges from 1.34 to 1.37, and Fe# ranges from 0.26 to 0.21, respectively. The following tendency is seen: the presence of hydrogen leads to the increase of the Fe/Si ratio and the decrease of Fe# in the mantle due to the increase of the core radius. The incorporation of 50 mol % of hydrogen into the core leads to the increase of Fe/Si ratio up to about the chondrite ratio.

We have calculated a series of Martian interior models with core density profiles calculated for core compositions ranging from pure Fe (0 wt % S) to FeS (36 wt % S). Figure 2 indicates the relation between the core radius, the mantle Fe#, the sulfur content in the core and the moment inertia factor. The Fe/Si ratio is lower than the chondrite ratio for any of these models.

The higher sulfur and hydrogen content in the core and the smaller mantle Fe#, the less likely a perovskite layer exists.

Conclusion. Based on available chemical models of the planet [1,2,3,4,5], a new set of global models of the Martian interior has been constructed. Quantitative studying the effect of hydrogen in the core on planetary structure is one of the main goals of the paper. If there is no hydrogen in the core, a model produces a Fe/Si ratio that is smaller than the chondritic value of 1.71. The presence of hydrogen in the core significantly increases the Fe/Si ratio up to about 1.7, and reduces the melting temperature of the core material. To satisfy the bulk chondritic ratio, more than 50 mol % of hydrogen must be incorporated into the core. Then, a problem of consistency of the cosmochemical DW model with the internal structure model of the planet is solved. It will confirm the idea that terrestrial planets were formed from chondritic material. This is a fundamental problem on the formation of Mars and its evolution.

The determination of the core radius continues to be of great importance, in case of a reliable determination of the core radius uncertainties concerning the composition of Mars will be resolved. From cosmochemical point of view, it is difficult to assume that the core contains more than 20 wt % of sulfur. The radius of such core is about 1600 km. Therefore, if the core of Mars turns out to be larger, hydrogen could be such an admixture element. According to numerical modeling hydrogen increases the core radius and decreases Fe# of the mantle.

Acknowledgements. This research was made possible by Grant No. 03-02-16195 from the Russian Foundation for Fundamental Research and State Programme No.13.

Table. Parameters of the models

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Fig. 1. Core radius as a function of Martian mantle Fe#, assuming a core composition of 14 wt % S and a 50-km-thick crust (a), and a 100-km thick crust (b) for different amount of hydrogen in the core (0-70 mol %). Dashed lines show the lower (left) and upper (right) limits of the moment inertia factor. Fe/Si ratio is given for boundary models.

Fig. 2. Core radius as a function of Martian mantle Fe# for a core composition ranging from 0 wt % S (Fe-core) to 36 wt % S (FeS core) assuming a 50-km thick crust (a) and a 100-km thick crust (b). Dashed lines show the lower (left) and upper (right) limits of the moment inertia factor. Fe/Si ratio is given for boundary models.