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Prabhut: m42_58, m42_59
S. G. Puğachev: m42_12, m42_60
J. Raitala: m42_35, m42_36, m42_37, m42_65
V. N. Raskhozhiev: m42_30
D. Reiss: m42_50
T. Roatsch: m42_68
A. V. Rodin: m42_01, m42_16, m42_33
E. L. Ruskol: m42_48
RSC Team: m42_21
A. K. Rybakova: m42_33
A. N. Sanovich: m42_46
D. E. Schea: m42_22, m42_58
V. V. Shevchenko: m42_60
S. N. Shilobreeva: m42_28
Yu. G. Shkuratov: m42_29, m42_34, m42_56, m42_62, m42_71
N. A. Shubina: m42_44
S. I. Skuratovsky: m42_61
V. V. Shuvakov: m42_51
B. A. Smith: m42_52
SPICAM Team: m42_33
D. Stankovich: m42_62
L. V. Starukhina: m42_63
R. Stesky: m42_30
H. Svedhem: m42_64
E. N. Sytnynik: m42_36
V. G. Tejfel: m42_52
D. V. Titov: m42_64
T. Törnänen: m42_36, m42_65
V. V. Tsyplakov: m42_30
G. K. Usinov: m42_66
A. A. Valter: m42_40
S. van Gassel: m42_06
S. G. Valeev: m42_67
Yu. I. Veliokosky: m42_02, m42_34
G. Videen: m42_29, m42_71
A. V. Vityazev: m42_45, m42_54
R. Wagner: m42_68
S. Werner: m42_06
L. Wilson: m42_21, m42_22, m42_69
M. Wolff: m42_29
O. I. Yakovlev: m42_70
T. Zegars: m42_50
Yu. I. Zetzer: m42_51
E. S. Zabko: m42_56, m42_71
M. Yu. Zolotov: m42_42
L. V. Zasova: m42_33
LINE MIXING AND COLLISIONAL BROADENING IN THE THERMAL RADIATION OF THE LOWER VENUS ATMOSPHERE: T.S.Afanasenko¹, A. V. Rodin¹², Institute for Space Research, RAS, Moscow, 117997, Russia. ²Moscow Institute of Physics and Technology, Dolgoprudny, 141700, Russia. rodin@im.iki.rssi.ru

Introduction: In the previous paper [1] we investigated the effect of the collisional line broadening on the spectrum and fluxes of the thermal radiation in the lower Venus atmosphere. It was shown that the Lorentzian form-factor at high pressure of the Venus atmosphere is no longer valid so that the absorption in the far wings of spectral lines is several orders of magnitude lower. We employed a theory of far-wing spectral line profile [2] which resulted in estimates of the thermal radiation fluxes in the lower atmosphere. However due to strong broadening exceeding typical rotational energy shift, spectral lines cannot be considered as isolated and the interference of states has to be accounted for [3,4]. In the current work we combine the two approaches and compare results with ground-based observations.

Line shape in the strong collision approximation: A simple and elegant way to take into account the effect of state interference on the rotation-vibration band is proposed in [3] by assuming a relaxation time parameter \( \tau_0 \) common for all states forming the band. In this approximation the spectral absorption \( \Phi(\nu) \) is expressed as

\[
\Phi(\nu) = \frac{1}{\pi} \text{Re} \left\{ \frac{U(\nu)}{1 - \tau_0 U(\nu)} \right\}
\]

where

\[
U(\nu) = \sum \frac{A_m}{\tau_0^{-1} + 2\pi(\nu - \nu_m)}
\]

and \( A_m \) are strengths of individual lines in a band.

The estimate of the relaxation time \( \tau_0 \) is based on the empirical data of line broadening under normal pressures, averaged over the band. It is important that the calculations be done over P, Q, R-branches of each vibration band separately. In spite of technical simplicity of the above model, its implementation for Venus atmospheric absorption is complicated by a priori unknown validity field. In particular, the asymptotic behavior of band shape consisting of multiple dense rotational lines is proportional to \(-\nu^{-4}\), whereas the far-wing approximation generally predicts the exponential asymptotic [3].

Line shape in the far wing: We have described in [1,2] the algorithm implemented for far-wing approximation neglecting the line mixing (state interference) effect. To account for both effects in practically useful calculations, the following approach may be employed. General shape of a spectral line is composed of two parts, central Lorentzian core and far wing suppressed by an exponential factor of power-scaled argument [2]. However in the central part of each line of the band, a Lorentzian is replaced by a line mixing profile presented above. Since most of strong vibration bands of interest for the Venus atmosphere have relatively dense distribution of rotational lines, within such bands the profile is identical to the model [3]. Far apart from the band, where the absorption is determined by a superposition of far wings of lines, it reveals exponential behavior. However this profile would not be identical to pure far-wing approximation since the parameters of the theory need to be updated.

Comparison with observations and implications for Venus Express: NIR transparency windows in the Venus atmosphere give us a chance to test theoretical radiative transfer models versus observations. A comparison of our calculations with data [5] kindly provided by B. Bezard, is presented in Fig. 1. Radiative transfer calculations were done according to [1] with the correction of error in partition sum evaluation. In some particular bands the discrepancy from observations is significant, while in other regions the model fits data with high accuracy. Tuning the model to available observations will allow us to derive physical parameters of the atmosphere from IR spectrometers of Venus Express project, as well as to remove the contribution of the atmosphere in observations targeting Venusian surface. This model presents a first attempt to construct fully ab initio radiative transfer model applicable for wide range of pressures and temperatures. Interpretations of Venus Express remote sensing data related to different atmospheric levels is only possible based on such or similar model.

ABSOLUTE CALIBRATION OF LUNAR SPECTROPHOTOMETRY DATA. L. A. Akimov, and Yu. I. Velikodsky. Astronomical Institute of Kharkov National University. Sumskaya ul., 35, Kharkov, 61022, Ukraine. E-mail: velikodsky@astron.kharkov.ua.

**Introduction:** A problem of absolute calibration of lunar albedo is very important and actual. Accurate knowledge of albedos of lunar areas (as well as its photometric functions) lets us use them as photometric standard for observations of planets and the Earth’s surface. Moreover, albedo is an important photometric parameter, which may be used for studying of physical properties of lunar regolith.

Absolute measurements of the Moon is a difficult task because magnitudes of possible photometric standards (the Sun or stars) greatly differ from the lunar one. Using the Sun is more preferable, because the Sun is a light source for the Moon and such a measurement is direct. But in this case there is a problem of taking into account possible changing of atmosphere transparence during observation. As a result of presence of these difficulties an accuracy of existing measurements is not enough.

We suppose that most precise absolute calibration has been obtained by Akimov [1] in red light (λ=660 nm). His albedo is based on Sytinskaya-Sharonov’s absolute system obtained with visual photometry. Phase dependence at large phase angles is based on Sytinskaya-Sharonov’s [2] and Peacock’s [3] data. And phase dependence near opposition is based on Wildey-Pohn’s data [4]. Moreover, Akimov experimentally studied the law of brightness distribution over lunar disk [5] that let him correctly calculate the albedo.

Akimov’s photometric system has a good agreement with Saary-Shorthill’s system [6] within accuracy about 10% [7]. So, 10% is the accuracy of existing data. But it is desirable to improve accuracy of absolute data to use them as photometric standard.

At the same time Clementine spacecraft lunar data have been calibrated using laboratory measurements of lunar samples and their albedo is 2.5 times greater than Akimov’s one [8]. Therefore it is necessary to provide new independent observations to check different absolute photometric systems.

**Observational data:** In 1986 we have performed a series of photometric observations of the Moon and the Sun during 4 days and 3 nights in 3 narrow spectral bands (440, 550, 660 nm). Phase angle was changed in interval 0.8-25° (observations includes near-eclipse phase). Observations have been performed at 70-cm reflector in Grakovo station of Kharkov observatory (near Kharkov, Ukraine) with photoelectric photometer. 75 lunar areas (from catalogue [1]) were measured with 50-cm objective diaphragm and center of the Sun was measured with 16-cm diaphragm. Full luminous intensity of the Sun was calculated with taking into account of darkening to the solar limb at corresponded wavelength.

**Data processing:** Using observations of solar center we have studied changes of atmosphere transparence during observation and have calculated “exoatmospheric” brightness of all lunar areas and solar center. Then absolute visible albedo A has been calculated with formula:

$$A = \frac{B_{\text{LA}}}{B_{\text{SC}}} \cdot \frac{R_{\text{S-L}}^2}{R_{\text{S}}^2 k_{\lambda} K},$$

where $B_{\text{LA}}$ – brightness of lunar area, $B_{\text{SC}}$ – brightness of solar center, $R_{\text{S-L}}$ – Sun-Moon distance, $R_{\text{S}}$ – radius of the Sun, $k_{\lambda}$ - coefficient that takes into account darkening to solar limb, $K$ – “instrumental” coefficient that contains diaphragm squares ratio and input resistances ratio. $K$ is equal to 470016, and $k_{660}$ is equal to 0.747 (for 440 nm), 0.803 (550 nm), and 0.825 (660 nm).

For calculating of precise phase angles and others photometric conditions with taking into account location of area on the Moon and location of observer on the Earth we have used formulas of coordinate transformation [9].

**Equigonal albedo obtaining:** We obtain from observation a visible albedo and can not calculate normal albedo (also named simply albedo) because photometric function of the Moon is not studied with enough precision. In [5,10] it was shown that photometric function can be separated up on two parts:

$$A_v = \rho(\alpha)\Psi(\alpha, i, i),$$

where $A_v$ – visible albedo, $\rho(\alpha) – equigonal albedo$ (albedo on “standard” conditions with mirror geometry, when incidence angle $i$ is equal to emergence angle $\varepsilon$ and is equal to half of phase angle $\alpha$ ($i=\varepsilon=\alpha/2$) [10]), $\Psi(\alpha, i, i)$ – disk brightness distribution function for fixed phase angle. For small phase angles ($\alpha<25^\circ$) function $\Psi$ is known precisely enough [5,11,12]. So, we can calculate equigonal albedo $\rho$ by (1) and study phase dependence $\rho(\alpha)$. The normal albedo can be obtained as $A=\rho(0)$.

**Phase dependence:** For approximation of phase dependence of equigonal albedo we used semi-empirical formula of Akimov [10,13]:

$$\rho(\alpha) = g \cdot e^{-\alpha} + m \cdot e^{-0.7\alpha},$$

where $g$ and $m$ are constant.
where $g$ and $\gamma$ – parameters of opposition peak, and the term with coefficient $m$ describes phase dependence on large phase angles (we used average value of exponent coefficient $0.7$ [10,13]). Obviously, albedo $A=g+m$.

**Preliminary results:** Now we have results for one lunar area (crater Le Monnier), which has the most number of measurements. Obtained equigonal albedo (in red band) at $\alpha\sim25^\circ$ (the night with best atmospheric stability) is exactly equal to old Akimov’s equigonal albedo [1] of Le Monnier 0.049 with standard deviation 3% (fig.1). For smaller phase angles observed albedo is slightly greater than old one, and near opposition ($\alpha=1-2^\circ$) albedo is greater by 15% (old normal albedo is 0.099, new – 0.115).

So, our new independent observations, as a preliminary, confirm Akimov’s absolute photometric system at least for $\alpha>15^\circ$. Equigonal albedo for smaller phase angles (and, as result, normal albedo) is greater than Akimov’s ones up to 15%. The last fact may be connected with the circumstance that Sytinskaya and Sharonov did not obtain phase dependence for small phase angles, and Akimov used data of Wildey and Pohn [4] for this phase interval.

Also we compared albedos of Le Monnier at 3 wavelengths with spectrum of lunar mares built by V.Kaydash on the base of C.Pieters catalogue. Our data repeat spectral slope of this spectrum with deviations about 4%.

We plan to obtain absolute albedo for all 75 lunar areas and to study phase dependence at small phase angles more detailed.

Also we plan to process data of our new absolute observations of 2005 year at large phase angles.

This work was supported by CRDF (grant UKP2-2614-KH-04).

**References:**
A laboratory modeling of the impact on the target presenting the surface layer of a comet is described. The impactor was a solid body, the target being a mixture of ice and organic matter. After the rapid impact, the ejection of the target fragments was not homogeneous. Against the background of an expanding cloud of drops (of water with an inclusion of organic particles), we observed a high-pressure jet. It is possible to choose an experimental regime similar to that realized in the Deep Impact NASA experiment.
“MUD VOLCANISM” ACTIVITY IS LIKELY TO EXIST ON TITAN. V.A. Alekseev. State Research Center of Russia (TRINITI) ‘Troitsk Institute for Innovation & Fusion Troitsk, Moscow region, Russia A.Getling@ru.net

Juxtaposing some images made by the Huygens probe near the surface of Titan [1] and the photos of a mud volcano region on the Earth (Kerch and Taman peninsulas, the Crimea and the Caucasus, accordingly; this volcano region has been studied in [2] and others) reveals similarity of geomorphologic features. Therefore, we put forward a hypothesis about existence of the mud-volcano activity on Titan. The liquid methane of Titan can manifest itself in producing it, instead of methane gas producing this activity on the Earth. For Titan, gas hydrates (hydrates of hydrocarbon gases) and water ice are the analog with the earthly clay breccia. Note, in accordance with [3], that gas hydrates are stable at the P-T condition known for Titan. The mud-volcano activity can explain: 1) general type of the landscape about the Huygens probe landing site on Titan; 2) the chain of bright islands seen during descent to the landing site, which can be marker of a fault line; 3) the conic form of a hill at the first plan of the image taken from an altitude of 8 km; 4) the rounded pebble-like form of the small blocks on the surface of Titan (compare similar form of blocks seen on the surface of Venus [4]); 5) the presence of solitary long white strips, each of which diverges at one of the ends (the methan wind can bring this picture into existence by extending matter of volcano-eruption pollution).

The solar radiation is the fundamental source of energy that drives the Earth’s climate. Climate models show that total solar irradiance variations can account for a considerable part of variations of the temperature of the Earth’s atmosphere in the pre-industrial era. Earlier we have considered the correlative connection between some climatic changes on the Earth and the activity of the Sun [1]. In this paper the analysis of data of monthly average and annual average regional variations in the surface air temperature for the last 125 years [2] in comparison with the solar activity variations in the 11-year [3] and secular cycles, is performed.

The variations of the solar activity and surface air temperature are compared in Fig. 1. The solar activity variations are shown for the 11-year (Fig. 1a, curve 1) and for the secular (curve 2) cycles in the interval 1700-2004 yrs. The analysis of the entire set of data in this interval has shown that over the background of periodic changes, the solar activity increase is observed on the average at ~0.2%/year. The tendency of solar activity increase in the interval 1880-2004 yrs is much higher: ~1%/year.

The temperature variations of the surface air for the last 125 years are shown in Fig. 1b. The well-defined tendency of the temperature increase in time is seen. The parameters of the regression line correspond to the increase of temperature over the analyzed period by 0.61±0.05 °C. The falls of the temperature (by ~0.1-0.2 °C) in the intervals ~1900-1920 and 1960-1980 yrs are probably conditioned by the large number of catastrophic volcano eruptions in these periods [4].

Distributions of average annual anomalies of the temperature of land and oceans during different periods of the solar activity in a secular cycle (number of a sun-spots Ri<120 and Ri≥120) within the interval 1880-2004 yrs are shown in Fig. 2. One can see, that the range of values of temperature anomalies for the land is almost twice is larger than for the ocean. Such a difference is connected, undoubtedly, with the large thermal inertia of oceanic masses. Median values of distributions for high solar activity years are significantly different from values for years of low solar activity (by ~0.3 °C).

Distributions of the annual average anomalies of temperature of land and oceans in the different periods of the 11-year solar cycle within the interval 1880-2004 yrs are shown in Fig. 3. Here, for each 11-year cycle of the solar activity, the years related to a high level of activity of the Sun in this cycle and the years with a low level of activity of the Sun were selected. The obtained data were used for construction of the time dependence of the annual average temperature anomalies for years with high and, separately, low solar activity in the 11-year cycle (analogue of Fig. 1b for all years). For each of such dependences, the regression line was calculated and the Δt value was determined (Δt is the increase of temperature for the whole time period). Such a procedure was made for the data related to different regions of the Earth. The obtained results demonstrate the clear-cut distinction between Δt values related to years with different levels of the solar activity. This distinction is most clearly seen for the land, especially for high latitudes of the northern hemisphere (Fig. 3a). For example, the Δt values for the range of latitudes 20°N-90°N for years of high and low solar activity are equal to 1.25±0.11 and 0.79±0.11 °C, respectively. At transition to the southern hemisphere Δt values decrease, but here higher values of Δt are also peculiar for the periods of high solar activity.

For oceans (Fig. 3b) the latitude dependence is practically not seen, though a tendency remains: higher Δt values correspond to the years with high solar activity.

The obtained data testify that the temperature on the Earth “has the time” to react to the changes of the solar radiation in a 11-year cycle. This reacting manifests itself most legibly for the land. The non-uniform distribution of the land and water surface, as
SOLAR ACTIVITY AND CLIMATE CHANGES OF THE EARTH: Alexeev V.A.

as well as the small thermal inertia of the land and the large inertia of the oceanic water, are apparently the main causes of the found peculiarities in the distribution of Δt values for different regions of the Earth.

References


Captions

Fig. 1. Comparison of the solar activity variations (a) in the 11-year (curve 1) and secular (curve 2) cycles with the temperature variations of the surface air of the land (b).

Fig. 2. Distributions of the average annual anomalies of air temperature in 1880-2004 yrs over the surface of land (a, a') and ocean (b, b') for different levels of the solar activity in a secular cycle. a, b - number of sun-spots in a secular cycle $R_i < 120$; $a', b' - R_i \geq 120$. The arrows mark median values (°C): -0.19±0.03 (a); -0.15±0.03 (b); 0.09±0.02 (a'); 0.11±0.02 (b').

Fig. 3. The increase of the surface air temperature for 125 years (1880 – 2004) according to the regression line (such as that in Fig. 1b) for different regions of the Earth: 90°N-20°N (mean latitude $L_{av}=55°N$), northern hemisphere ($L_{av}=45°N$), 20°N-20°S ($L_{av}=0$), southern hemisphere ($L_{av}=45°S$) and 20°S-90°S ($L_{av}=55°S$). The calculation of regression lines is carried out for years of high-level solar activities in a 11-year cycle (1), low-level solar activity (3) and for the entire set of data; a and b are the data for the land and ocean, respectively.
**Introduction:** This work is based on the photogeologic analysis of the High Resolution Stereo Camera (HRSC) images of the eastern flank of Olympus Mons volcano and adjacent lowland plains taken at orbit 1089 (Fig. 1). The HRSC-derived DTM’s and topomaps as well as MOC and MOLA data were also used. This study continues our analysis of the western flank of volcano, which has been published in [1-2]. Our initial results from the eastern flank have been presented in [3-5]. As in previous analysis we divide the study area into three domains: 1) the volcano summit plateau, 2) the volcano slopes, and 3) the lowland plains.

**Observations:** The summit plateau. The surface of the summit plateau’s eastern flank shows ~200 Myr old [3, 5] lava flows almost everywhere. In comparison to several small mesas on the summit plateau in the west, only one relatively large mesa-like feature (covered by lavas, Fig. 2) and a few small steep-sloped hills (all marked M in Fig. 1) are observed here. Layers in their slopes are rarely seen most likely because of a dust mantle, the presence of which is suggested by down-slope trending dark streaks [e.g., 7].

The S3 slopes are typically the most gentle and covered by lava flows continuing from the summit plateau down to the lowland plains. In the east there are no the S2 slopes, but the S1 and S3 slopes are abundant. A new type is also present: the S4 slopes, which appear when lavas flow over the rims of the S1 slopes (Fig. 3).

The slopes. In the west, they were classified into three morphological types: S1, S2 and S3 [2]. The S1 slopes are the steepest and often have ravinets. The S2 slopes are less steep. In their uppermost part there are several chaos-like depressions, from which the channel-like grooves trend downhill.

The lowland plains. Within the study area they typically have a smooth surface. In their southern part, a few networks of channels typically starting from the steep-sloped troughs are visible (blue lines in Fig.1). The channels intersect, anastomose and form networks, the largest network is about 10 km wide and 60 km long. The mean surface age of the...
network bearing subarea is ~80 Myr [3, 5]. The morphology of these networks, including the presence of streamlined islands and terraces (Fig. 5), resembles that of the Martian outflow channels, which were previously interpreted as formed by the catastrophic release of subsurface water [e.g., 8].

Recently, this interpretation has been challenged by the suggestion that highly fluid lavas could have cut the channels [e.g., 9]. Our observations do not support a particular hypothesis.

In the HRSC and MOC images of the lowland plains, ridges that are significantly more rectilinear than normal wrinkle ridges are observed. These sometimes merge into dome-like hills, while such hills occasionally form linear chains (Figure 7). Locally, these features are superposed on channels. We interpret these ridges and hills as extrusions of subsurface material, whose nature (lava, mud, ice?) we hope to determine in future studies.

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Discussion and conclusion. The summit plateau morphology of Olympus Mons’ eastern flank shows remnants of layered sediments, supposedly airborne, which suggests a combination of volcanic and surficial deposition processes in the formation of the volcanic construct. In the west, similar sediments contained and probably still contain [13] water ice. This interpretation is supported by the S2 slope morphology as well as numerous glacial-type landforms [1, 2]. In the east, we did not find evidence for water ice in the past or present within summit plateau layered deposits. At the foot of some slopes in the east, we found a few landforms suggesting glacial-type activity, but on a significantly smaller scale than in the west. In the east, the plains have trough and channel networks, which could be cut either by water or by lava. Here, tectonic compressional landforms are seen, which are not present in the west. Therefore, geologic activity in the eastern flank of Olympus shows similarities with that of the western flank in lava emplacement style and differences in surficial water/ice processes and tectonism.


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GEOMETRY AND FLOW PARAMETERS OF LAVAS OF OLYMPUS MONS VOLCANO: MARS. E. A. Bazilevskaya\(^1\) and G. Neukum\(^2\), \(^1\)Geological Department, Moscow State University, Vorobiiev Gory 119992, Moscow, Russia, Ek_Bazilevs@mail.ru, \(^2\)Freie Universitat Berlin, Berlin, Germany

Introduction: This work is a continuation of our previous study when using the MGS MOC images and MOLA profiles we have measured the thickness of lava flows in several localities of Olympus Mons volcano on Mars [1]. This time we report on the study of other geometric characteristics of lava flows observed on the surface of Olympus Mons volcano, such as their length, width and also slopes of the surface on which the lava flows flowed. Combining them with the earlier measured thickness values we estimate some parameters of the lava eruption: yield strength, effusion rate and viscosity of the lava.

Lava flows, which are subject of this study, are streaming downslope in directions generally radial to the volcano center. Some of them are cut by the basal scarp of the volcano, others extend over it onto the surrounding plains. Many of the individual flows have axial channels similar to those of the channel-fed terrestrial basaltic lava flows. We studied these flows, analyzing HRSC, MOC and MOLA data.

Morphology and Dimensions: We studied the mentioned parameters of lava flows on the southern slope of Olympus Mons volcano. For that we have made a schematic geologic map based on the analysis of the HRSC image 0037 with resolution of 20 m. This mapping allowed us to divide the southern slope into 4 zones (Figure 1).

Zone I is characterized by individual flows that originate from the caldera scarps and flowed towards west and east. Width of the flows (~500 m) generally exceeds values for other zones. Zone II (extends 70 km down slope) – zone of flows, which outlines are significantly camouflaged by wind erosion and airborne dust deposition so only several flows could be well seen. Among them, two unusually wide, short and thick palmette flows are observed. Specific distinction of Zone III (40 km down slope) – zone of flows, which outlines are areally extensive complex flows (~1 km wide), composed of tens of overlapping individual flows with vague outlines. Zone IV (110 km down slope, terminates at the foot of Olympus Mons) presents a net of clear-cut individual flows, mostly channel-fed. Within Zones II – IV collapsed lava tubes could be locally recognized as chains of pits ~150m wide and up to 1.7 km long.

On the geologic scheme, within Zones II-IV, we have outlined four 8.9 x 27.9 km areas where the measurements of lava width and length as well as the slope steepness have been made (Figure 2).

Measurements of width of the flows have been made for each of the four selected areas. The whole flow width (W) and the width of the leveed flow channel (w) have been determined if seen. Measurements of length of 30 individual lava flows of southern slope of Olympus Mons have been mostly made within the selected zones although in some cases we have traced the flow outside of the area. The average slope steepnesses have been measured using the MOLA database as mean values for each of four selected areas. The results of measurements are summarized in Table 1.

<table>
<thead>
<tr>
<th>Area</th>
<th>Mean flow width (W)</th>
<th>Mean channel width (w)</th>
<th>Lengt h (L)</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>~4.8 km; ~2 - 1.4 km</td>
<td>-</td>
<td>22.2 km</td>
<td>3.4°</td>
</tr>
<tr>
<td>2</td>
<td>377 m</td>
<td>-</td>
<td>4.9°</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>395 m</td>
<td>144 m</td>
<td>4.4°</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>484 m</td>
<td>220 m</td>
<td>6.9°</td>
<td></td>
</tr>
</tbody>
</table>

Measurements of lava flow thickness have been done in our previous work [1]. We have been searching for MOC images on which steep scarps of the volcano are seen: in any places of volcano, not only on the southern slope. In these places the
apparent thickness of the outcropped flows was measured and using MOLA measurement of the slope steepness, the apparent thickness was recalculated into the true one. The resulting mean values for true thickness are ~6.7 m for the volcano flanks and ~1 m for the caldera slopes.

The results of the width, length and slopes measured for the southern flank of Olympus Mons have been combined with the measurements of thickness made for different flanks and caldera and then jointly used for estimates of the parameters of eruption.

**Estimates of the parameters of eruptions:**

**Yield strength.** Lava flows are often modeled as a Bingham plastic controlled by two parameters, the yield strength and the plastic viscosity. Several geometric parameters, measurable from images, give an opportunity to determine the yield strength of lava flow:

\[
Y_1 = \rho g H \sin \alpha
\]
\[
Y_2 = \rho g H^4/W
\]
\[
Y_3 = \rho g (W-w) \sin^2 \alpha
\]

Gravity g is known as 3.7278 ms\(^{-2}\) and density \(\rho\) was chosen to be 2,500 kgm\(^{-3}\) following Hiesinger et al. [2], Zimbelman [9], H = thickness of lava flow. Depending on the equation used, we find a minimum yield strength of ~0.9 \times 10^5 Pa and a maximum yield strength of ~3.6 \times 10^5 Pa, with the average 7.9 \times 10^4 Pa.

**Effusion rate.** The effusion rate (R) is given as:

\[
R = 300 k w L/H,
\]

where \(k\) = is thermal diffusivity of the fluid (taken 7 \times 10^{-7} \text{ m}^2\text{s}^{-1} for mafic lavas following Zimbelman [9]), L is the flow length (m), and w and H is defined as above. For calculating effusion rate we took mean length of lava flows (22.2 km) for regions 2-4. As a result we find that effusion rates range from ~24 to ~137 m\(^3\)s\(^{-1}\), averaging about 53 m\(^3\)s\(^{-1}\).

**Viscosity.** Viscosity (\(\eta\)) of a flow can be calculated with the estimated yield strengths and effusion rates obtained above [8].

\[
r = w/(W-w)
\]
\[
\eta = w^{1/3}Y^{5/4} \sin^{1/4} \alpha /24 R g/l/4 \rho l/4 \quad \text{for} \ r < 1
\]
\[
\eta = w^{1/3}Y^{5/4} \sin^{1/4} \alpha /24 R g/l/4 \rho l/4 \quad \text{for} \ r \geq 1
\]

where \(r\) = is ratio of the channel width to the total levee width. As a result we calculated average viscosities of ~6 \times 10^3 Pa-s. Minimum viscosities are on the order of 1.4 \times 10^2 Pa-s, maximum viscosities are about 2.8 \times 10^5 Pa-s.

**Discussion:**

**Geometric characteristics of lava flows.** Calculated lengths of Olympus Mons lava flows (average ~22.2 km) are in good agreement with results of Hiesinger et al. [2] for flows of Ascræus Mons (average length ~19 km). According to [4] typical values for length of basaltic lava flows of Hawaiian volcanoes do not exceed 50 km.

Results of our measurements of the Olympus Mons lava flow thickness (mean values = ~11 m for the caldera scarps and ~6 m for the volcano flanks) are in a good agreement with Schaber et al. measurements [7] for different localities of Tharsis region of Mars and with typical thicknesses reported for the terrestrial basaltic flows (3-20 m [4]).

**Estimation of rheologic properties.** Our results of estimation of yield strength (average ~7.9 \times 10^5) are in good agreement with estimates for terrestrial basaltic and andesitic lava flows (~10^4-10^5 Pa and ~10^5-10^6 Pa correspondingly), and are comparable with estimates of Hiesinger et al. (~2.7 \times 10^5 Pa) [2] and Zimbelman (~2.1 \times 10^5 Pa) [9] derived for lava flows on Ascræus Mons.

On the basis of our calculations we derived the values of effusion rate for studied lava flows of Olympus Mons – average ~53 m\(^3\)s\(^{-1}\). These results are within the range of 5-1000 m\(^3\)s\(^{-1}\) effusion rates typical for Hawaiian basaltic eruptions [6] and are in excellent agreement with results of Hiesenger et al. [2] for lava flows of Ascræus Mons (average ~68 m\(^3\)c\(^{-1}\)).

We estimated average viscosities of studied lava flows on Olympus Mons to be ~6 \times 10^5 Pa-s. According to Macdonald [5] field measurements of basaltic lavas indicate viscosity of 10^5-10^6 Pa-s. Hulme [3] reported viscosities of 1.7 \times 10^5 Pa-s for andesites of the Paracutín volcano in Mexico. Previously published values for viscosity of Martian lava flows range from 5.2 \times 10^5 to 2.1 \times 10^6 Pa-s [2, 3, 9].

**Conclusions:** Calculated yield strengths, effusion rates and viscosities for flows on Olympus Mons are similar to values obtained for flows on other Martian shield volcanoes. Values for yield strength and effusion rates are in good agreement with data for terrestrial shield volcanoes, and calculated viscosities of the Olympus Mons lava flows partly coincide with the results of field measurements of viscosities of terrestrial basalts. On the basis of our investigation we conclude that the lava flows investigated are likely to be basaltic to andesitic in composition.

**References:**


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DETAILED STUDY OF ROUGHNESS ANISOTROPY OF THE VENUSIAN PLAINS.

N. V. Bondarenko1,3, M. A. Kreslavsky2,3 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 Institute, Kharkov, Ukraine; 3Dept. Geol. Sci., Brown University, Providence RI, USA.

Introduction. We analyze properties of Venusian surface using processed Magellan radar altimeter (RA) data [1] stored in the SCVDR PDS data set (Surface Characteristics Vector Data Record). These data contain estimates of Doppler centroid shift $f_D$ that characterizes along-track (north-south) anisotropy of the backscattering function of the surface. For plains, this anisotropy is caused by the north-south (N-S) asymmetry of the subresolution surface topography [1,2].

Our previous analysis of the $f_D$ distribution has revealed a global hemispherical trend of roughness anisotropy: in general, equator-facing small-scale topographic slopes are steeper than pole-facing [2]. This trend is in general agreement with the global pattern of wind directions inferred from wind streaks [3], if we assume that the observed anisotropy is due to the presence of microdunes on the surface [2].

Here we report on detailed comparison of Doppler centroid shifts and geological objects using Magellan SAR mosaics.

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]). For a globally horizontal surface with an isotropic backscattering function, the radar echo Doppler spectrum is symmetric with respect to the Doppler frequency corresponding to the nadir, and the strongest echo comes from the nadir bin. Non-zero Doppler centroid shift $f_D$ means that the strongest echo in the along-track direction comes from either ahead ($f_D>0$) or behind ($f_D<0$) the nadir.

With the $f_D$ estimates from the SCVDR, we generated a gridded map of $f_D$. Our map provides better visual sharpness and more suitable for comparison with the radar images than an analogous map from the GVDR data set.

One Doppler frequency bin (935 Hz) corresponds to the surface tilt of 0.4° close to the orbit periastris (low latitudes) and up to 0.8° in the polar regions. The single-burst RA footprint size is ~25 km near the Magellan orbit periastris (~10°N) and up to 220 km in the polar regions. Thus both sensitivity of $f_D$ to roughness anisotropy and the resolution change strongly with the latitude.

Observations: We searched Venusian plains for correlations between the $f_D$ map and surface geology seen in the Magellan radar (SAR) mosaics. We found that there is no global direct correspondence between $f_D$ and surface morphology. In many places, however, we do see a coincidence of sharp contrasts in the $f_D$ map and boundaries of geological units. This gives the strongest observational evidence that $f_D$ in the plains indeed reflects some intrinsic anisotropy of the surface.

Role of tectonic structures: The topography or roughness responsible for the observed anisotropy can be of very different scales ranging from centimeters to kilometers. The largest scales in this range, kilometers and hundreds of meters, are resolved by the Magellan SAR. Numerous tectonic features of these scales are seen amid the plains in the SAR images. There is no clear indication of pronounced asymmetry of these features.

The most abundant resolved tectonic features in plains are the wrinkle ridges. Their three-dimensional shapes are rarely resolved, however, the largest representatives as well as analogy with similar features on the other planets [4] indicate that the ridges often have asymmetric profile [5]. The sense of this asymmetry, however, often changes along individual ridges, and there is apparently no consistent asymmetry through large areas. Ridges typically occupy less than 10% of the footprint area [5]. Thus, we do not expect wrinkle ridges to play a significant role in formation of scattering anisotropy. We also do not see correlation between areas of high scattering anisotropy and the density of wrinkle ridges (according to [6]).

Typical regional plains. We found eight examples of relatively large ($>1.5\cdot10^5$ km$^2$) uniformly anisotropic areas in the very typical regional plains. These areas are far from apparent parabolas, diffuse halos and other features attributed to the presence of surficial deposits (e.g., [7] and references therein). All areas contain well-developed networks of wrinkle ridges. The areas are composed of different flow units; their boundaries have low or moderate contrast in the SAR images and do not appear in $f_D$ maps. There is no clear coincidence seen between the flows boundaries and the edges of uniform areas in the $f_D$ map. These typical uniform areas have rather high (although not extreme) values of $f_D$. Seven of the eight follow the hemispherical trend.

Correlation with volcanic units: There is a number of examples, where the distinctive contrasts $f_D$ correspond to boundaries of distinctive volcanic units in SAR images. We found 24 examples of lava flows showing rather uniform $f_D$ sharply different form the surroundings. All flows found are rather small; the largest of them, the bright lava flow in Sedna Planitia (Fig. 1), is about $10^5$ km$^2$. This flow, located at high northern latitudes, has strong slope asymmetry ($f_D \approx -2$ kHz) with steeper north-facing slopes, opposite to the hemispherical trend. Its surroundings show variable asymmetry, in average small but also negative $f_D \approx -0.2$ kHz. A few wind streaks
listed in [2] in the vicinity of this flow have the inferred wind direction to the north-west, also opposite to the hemispheric trend.

Five sites from the set form a tight cluster of similar flows in Rusalka Planitia. These flows are in the southern hemisphere and are characterized by positive $f_D$ contrary to the global hemispherical trend. Other sites do not occur in clusters. The remaining 9 flows in the southern hemisphere follow the global trend. The northern hemisphere flows exhibit both positive and negative $f_D$. Thus, in general, the distinctive lava flows do not follow the $f_D$ global trend.

All flows found are rather bright and some of them are very bright in SAR images and hence have rather rough surface at the scales of meters and decimeters. All flows show the degree of slope asymmetry higher than surroundings. No apparent correlation is seen between sign of $f_D$ and flow direction; this direction, however, cannot be reliably identified for some sites.

There is a principal possibility that the lava movement during the flow emplacement caused the observed surface roughness anisotropy. It is much more probable, however, that the high roughness favors accumulation of wind-blown material, and deposits of this material in wind shadows produce pronounced slope asymmetry.

**Splotches appearance:** Comparison between $f_D$ map and SAR mosaics showed that several areas of peculiar $f_D$ are associated with some largest splotches. Splotches are circular diffuse bright or dark features in SAR images (e.g., [8]) interpreted as "failed craters", the results of projectile explosion in the lower atmosphere. There are a few hundred of them on Venus. We found 17 clear examples of correspondence between a splotch and a feature in $f_D$ map (one of the best example is shown in Fig. 2). In the majority of cases (14 of 17) splotches have higher anisotropy than surroundings; in all these cases the sign of anisotropy follows the hemispherical trend. The remaining 3 splotches appear in $f_D$ map due to their weak anisotropy on highly anisotropic background.

**Discussion:** We believe that the deposits of wind-blown materials, either microdunes or deposits in wind shadows of topographic obstacles or both are the best candidates for explanation of the observed ubiquitous roughness anisotropy in Venussians plains. In the frame of the eolian hypothesis it is natural that we observe some correlation of anisotropy with geology, but this correlation is not universal and not uniform. Formation of anisotropic deposits is a complex process that depends on surface roughness characteristics, availability of proper loose material and its properties, and wind pattern during emplacement epoch. Furthermore, formation of anisotropic surface is interleaved with episodes of removal of anisotropy by emplacement of fresh lava flows, debris falls from large impacts, and eolian erosion. It is possible that eolian transport is active only episodically, during geologically short periods, for example, as has been hypothesized in [9], only after large impact events.

Our observations further confirm that the surficial deposits of wind-blown material are ubiquitous on Venus and can be present even in regions where we do not see diffuse features or wind streaks in the SAR images.

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**References:**

COLORADO RING STRUCTURE: A POSSIBLE ANALOG OF CORONAE ON VENUS.
G. A. Burba, Vernadsky Institute of Geochemistry and Analytical Chemistry, Russia’s Academy of Sciences, 19 Kosygin St., Moscow 119991, Russia <gburba@gmail.com>.

Introduction: Definition is given for the large circular structure on the Earth, named hereafter as Colorado Ring Structure (CRS), which is located in North America. Its center is at 37°N, 110°W. The overall topographic appearance of this structure looks to be similar to the large circular features on Venus termed Corona. So, it could have some importance for planetological studies to compare the coronae on Venus with terrestrial ring structures.

Geographic setting: The CRS is located in the western part of North America, between the Great Plains on the east and the Great Basin on the west. The structure have a clear appearance on the satellite images due to the sharp colour differences of its landscapes – dark-looking forests on the mountain ranges and bright-looking desert ground at the lower areas (Fig. 1). Its latitude/longitude framing is 33 and 41°N, 105 and 115°W.

The main rim of the CRS could be traced along the mountain ranges (Fig. 2), which are (counterclockwise from Flagstaff): Mogol on Rim at S, Manzano Mountains and Sangre de Cristo Mountains (within the Rocky Mountains system) at E, Sawatch Range, Gore Range and Park Range (Rocky Mountains) at NE, Elkhead Mountains and Uinta Mountains at N, southern segment of Wasatch Range at NW, Kaibab Plateau and Coconino Plateau at W.

Colorado Plateau occupies most of the inner part of CRS.

Topographic description: The outer diameter of CRS (between the rim foothills) is 840–880 km, the inner diameter is about 520 km, and the rim width is 150–200 km.

The rim altitudes are in general up to 3000 m. The higher areas within W and S segments are at 3000–3500 m, with maximal point at San Francisco Mount near Flagstaff (Humphreys Peak, 3851 m). E and N segments of the CRS rim are higher, around 4000–4400 m. The highest point at N segment is Kings Peak, Uinta Mountains (4123 m), and in E segment – Mount Elbert, Sawatch Range (4399 m), which is the highest point of Rocky Mountains range. So, there is a general topographic tilt of CRS from E to W, to be more precise, from NE to SW (Fig. 3). The plain areas within the inner part of CRS are at altitudes 1500–2000 m. The plains surrounding CRS are at altitudes 1000–1500 m.

The outer boundary of CRS have as a rule very sharp topography, expressed as steep megascarsps, especially in W, SW and S segments of the rim, at the areas of Hurricane and Aubery Cliffs at W outer foothills, and at Sedona Cliffs at SW segment.
**Hydrographic patterning:** The circular pattern of the CRS is outlined with the net of river valleys. Inner part of CRS is framed with Colorado river at N, NW and W, as well as with Little Colorado river from S an SW (Fig. 2). Just west of the Little Colorado mouth there is Grand Canyon, which cuts through the western rim of CRS. From E and SE there is a long canyon/valley of Rio Grande, which outlines the inner foothills of CRS rim. Gila river outlines the outer foothill of CRS from S. There are valleys of Muddy and Sevier rivers along the western outward side of CRS rim. As a whole, the CRS inner area is a catchment basin of Colorado river.

**Current geologic activity:** Colorado Ring Structure looks to be geologically active one.

**Seismicity.** Earthquake epicenters are concentrated along the rim of CRS. About 120 earthquakes occurred here during 1990–2000. All epicenters were at shallow depths – less than 35 km. Most of the epicenters are located within N, NW, W and SE segments of the rim.

**Volcanism.** There are about 20 volcanoes and lava fields along the rim of CRS (see red dots on fig. 2). The most recent activity took place at the Sunset Crater volcano, which is located on the SW segment of the rim. Its foothill altitude is 2120 m, cinder cone height is 320 m (Fig. 4). The latest activity occurred here 825 year ago, in 1180, when basaltic lava flow from the foothill have been formed, which is named Bonito Lava Flow.

**Comparison with coronae on Venus:** The overall topographic shape of the Colorado Ring Structure – a high mountain ring/rim around the lower, but still topographically high plateau, and lowland plains outside the rim – resembles the typical topography of the large circular features on Venus termed Corona [1]. Such topographic similarity could be result of the similarity in geologic evolution of these structures both on Earth and Venus. The data on Colorado Ring Structure could be applied for interpretation of the processes which form corona features on Venus.

NIKOLAEVA PATERA IN ASTERIA REGIO ON VENUS: A SMALL CORONA AT JUNCTION OF TWO REGIONAL FAULT BELTS. G. A. Burba, Vernadsky Institute of Geochemistry and Analytical Chemistry, Russia’s Academy of Sciences, 19 Kosygin St., Moscow 119991, Russia <gburba@gmail.com>.

Introduction: A rimmed depression in Asteria Regio on Venus have been named Nikolaeva Patera. The naming was done in 2005 by the International Astronomical Union in commemoration of Dr. Olga Nikolaeva, Russian planetologist and geochemist. The feature is located within the area of ancient terrains, but possess clear signs of younger geologic activity. It is located at the intersection of the two long, regional-scale fault belts, one of which is connected with the formation of Beta Regio upland.

The name origin: Nikolaeva Patera is named after Olga Vladimirovna Nikolaeva (1941–2000), who was a planetologist-geochemist at the Laboratory of Comparative Planetology, Vernadsky Institute, Russian Academy of Sciences. During 26 years she was involved in the planetary research. Her main field was geology and geochemistry of Venus, including both atmosphere and surface of this planet. Her main results in planetology include:

► development of the engineering model of lunar soil for the spacecraft design and landing (1974);
► discovering (with Venera 8–14 landers data) the diurnal variation process of the cloud cover of Venus – the geochemical cycle and day-to-night change in thickness of the cloud layer (1976);
► the first (and still the only) in situ measurements of the chemical data within the near-surface layer of Venus atmosphere, done with “Contrast” geochemical indicator designed and built by Olga (two patents for invention obtained); this indicator provided the unique data on the redox (reduction-oxidation) conditions in the layer as close as 15 cm from the planet’s superheated surface (Venera-13 and Venera-14 landers, 1982);
► estimations of terrestrial counterparts for the types of Venus rocks after their natural radioactive elements K, U and Th content (1982, 1997);
► development of geochemical aspects of Venus geology in application for the layered rocks origin (lava emplacement vs. sedimentation) – after the Venera landers’ close-up panoramic images (1983);
► definition of arachnoids as the specific class of endogenic geologic structures after the Venera-15 and Venera-16 SAR images (1986);
► definition of geochemical uniqueness of the rock type at Venera-8 landing site as Na-alkaline syenites (1983), which proposed as evidence of a continental crust on Venus (1990), and pancake domes suggested as source of such rocks (1992).

The naming of feature on Venus after Olga Nikolaeva have been done on the proposal of Prof. V.I. Feldman (Geological Department, Lomonosov Moscow State University).
Geographic description: Nikolaeva Patera is located NW of Beta Regio within the northern part of Asteria Regio. The center of patera is at 33.8°N latitude and 267.5°E longitude. As a whole this patera have an elongated shape with long axis 100 km and short axis 60 km. The feature consists of three oval-shaped depressions, which are nested within the joint rim (Fig. 1). Elongation is in NE – SW direction, which is along the same way as the narrow trenches of Agrona Linea fault belt (labeled A on Fig. 2). Nikolaeva Patera marks the SW end of this belt. Agrona Linea fault belt could be clearly traced from Nikolaeva Patera for 2300 km to the NE, then to E and SE around the northern outhills of Beta Regio upland. Patches of Sudenitsa Tesserae surround Nikolaeva Patera from every quarter.

Geological setting: Most ancient terrains in the vicinity of Nikolaeva Patera are patches of tessera – bright rugged areas located higher than dark plain terrains (Fig. 3). The latter are next in age. Dark plains fill the areas between tessara patches, embaying tessera. These plains in some areas undergone a tectonic reworking with numerous narrow sinuous trenches, which form elongated fault belts. The belts are younger in age than the dark plain areas. There are two main fault belts in the area. One of them have NE–SW direction and its segment located NE from Nikolaeva Patera belongs to Agrona Linea, which is circum-Northern Beta Regio fracture zone. Another fault belt in the area have NW–SE direction with further changing its running to E (Fig. 4). There are clear evidences of volcanic activity within Nikolaeva Patera and in the area adjuent to its rim. Small domes, both conical and flat-topped (“pancake”), are on patera bottom and on its rim. To the SE of patera there is a plain area with lobate flows running radial outwards the patera (Fig. 2). They looks to be originated from the patera’s rim.

Interpretation: Nikolaeva Patera sits at the junction of the two regional fault belts. Such position reflects the origin of the patera, which occupies most disturbed and so a weak spot of a crust in the area, being at the intersection of the two zones of disturbance, marked on the surface with numerous trenches. These trenches should resulted from the extension of the planetary crust. The NE – SW belt position is concentric to Beta Regio. So, the uprising of Beta Regio upland could be a cause for the formation of this belt.

The weak spot of a crust provided a channel for the interior material upwelling, which resulted in volcanic activity and formation of the rimmed depression of Nikolaeva Patera. As a whole Nikolaeva Patera looks like a small corona-type feature.

Fig. 3. Asteria Regio area radar image. Area size is 1500 x 1700 km.
SAR image: Magellan/NASA.

Fig. 4. Interpretation of regional linear features in Asteria Regio. Nikolaeva Patera is located at the junction of the two regional fault belts. The borders of these belts are outlined with white lines.
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 lasted until June 2004 when all the instruments started their routine operations. The MARSIS radar antennas, however, were deployed in May-June 2005, following comprehensive simulations of boom deployment and mitigation of potential risks, to benefit from nighttime conditions required for subsurface sounding before the pericentre natural drift in latitude, when illumination conditions become favourable to the other instruments. 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$_2$ ice occurrence in the South pole, showing where the two ices mix and where they do not. The Planetary Fourier Spectrometer (PFS) has confirmed the presence of methane for the first time, which would indicate current volcanic activity and/or biological processes. The UV and IR Atmospheric Spectrometer (SPICAM) has provided the first complete vertical profile of CO$_2$ 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. 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. Finally, preliminary results of the subsurface sounding radar (MARSIS) indicate strong echoes coming from the surface but lack of echoes under the young smooth Northern plains, which may indicate the presence of thick and homogeneous plains deposits.

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 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/)
AN ORIGIN FOR THE SOUTH POLE-AITKEN BASIN THORIUM. V.I. Chikmachev, S.G. Pugacheva, Sternberg State Astronomical Institute. Moscow University. Moscow. chik@sai.msu.ru.

Introduction: The lunar South Pole-Aitken (SPA) impact basin is one of the largest and oldest impact basins and has thus been heavily degraded by post-basin-formation primary impacts and their ejecta [1]. The general area of the SPA basin is characterized by numerous pre-Nectarian basin structures, for example the Hertzprung, Mendel-Rydberg, Korolev, Gagarin, Keeler-Heaviside etc. [2]. However, a few basins, for example the Al-Khwarizmi/King and Lomonosov-Fleming, probably formed before the SPA and located by near outermost ring of the structure but are now completely obliterated. This entire area to the north – westward of SPA consists almost exclusively of what are commonly called the "old features of the lunar surface". In figure 1 the highest crater frequency is found, that within the limits of the possible Al-Khwarizmi/King basin [3].

The SPA basin thorium map: The using data Lunar Prospector [4] the thorium distribution map demonstrated a hemisphere of the Moon which contains the SPA basin structure and its environs was constructed. Perspective azimuthal orthographic projection was used as cartographic basis of this map (fig.2). The principal thorium source is located almost at center of the SPA basin, exactly in Mare Ingenii ($\lambda = -166^\circ$, $\beta = -41^\circ$), where maximum thorium is more 5 micrograms/gram. Its height level is $H \leq -5$ km [5]. At edge of the SPA basin (near to its outermost ring) concentration of thorium is decreased up to a minimum (below 1 micrograms/gram). It is necessary to note, that secondary emissions of thorium to north-west from SPA where buried basins Al-Khwarizmi/King and Lomonosov-Fleming settle down the concentration of thorium is kept above 1.5 micrograms/gram.

Conclusions: On the basis of our study of the generalized structure of the SPA basin we conclude that ejecta of thorium from SPA have taken place at the center of basin in the moment of formation of all structure. Secondary emissions of thorium in northwest from SPA have taken place in a direction of motion a comet body caused impact and formation SPA basin structure [5]. Thus, SPA basin almost certainly exposes lunar material with very deep origin, almost certainly the lower crust and possibly the lunar mantle [6].

Figure 1. Cumulative size-frequency distributions of craters at least 20 km in diameter superposed on pre-Nectarian basins [3].

Figure 2. The distribution of thorium is shown on the map of the SPA basin. The center of the map represents the landing position $\phi=40^\circ$ S, $\lambda=180^\circ$. 
**Introduction:** Gullies have been observed along the flanks of a variety of surfaces in the mid/high-latitudes of Mars including crater and valley walls [1], and isolated topographic surfaces like central peaks, mesas, knobs, and raised crater rims [2]. Among the most striking of these features, however, are ones found along the slopes of dunes. Due to their small volume and lack of impermeable rock layers, dunes are problematic locations for groundwater accumulation and release, one of the candidate formation hypotheses proposed for martian gullies [1,3,4]. This has led other workers to conclude that gullies on dunes are triggered by surface accumulation of snowpacks that undergo melting to carve the observed channels [5], an hypothesis that has been offered to account for all gullies on Mars [6,7].

The apparent youth of martian gullies has led to debate as to whether or not they can be formed at ideal locations under current climatic conditions. Since obtaining data of these features is difficult due to their small size, focus has been given to models that simulate the current martian climate [6,8]. Hecht [6] concluded that the lack of heat provided by the thin martian atmosphere is balanced by increased direct solar insolation, and that liquid water can exist transiently at nearly any location on the martian surface. Therefore, if gullies on Mars are formed by accumulation and melting of surface snow deposits, they would be expected to form where ice is most stable: poleward slopes in the mid/high-latitudes of each hemisphere, where ice would not receive enough direct sunlight to increase the temperature sufficiently to sublimate.

While orientation measurements show that gullies indeed form on poleward slopes in the mid-latitudes of the southern hemisphere [4,9], little work has been done to decipher a geologic association between surface snowpacks and gullies based on available data. Christensen [7] presented evidence for a remnant snowpack within a suite of gullies on the poleward wall of Dao Vallis in Mars Orbiter Camera (MOC) image M09/02875, but images of this region show similar deposits at all seasons (e.g. M03/07529), implying that gully formation at this site is no longer active. In this contribution we document a set of channels incised into dunes in the presence of a seasonal frost deposit in the southern mid-latitudes and discuss implications with regard to gully formation and the stability of liquid water on present-day Mars.

**Observations:** MOC image M04/03432 shows a dune field found on the central peak of an 80 km diameter crater at 52°S, 33°E (Figure 1a). The southernmost of these dunes show distinct frost deposits on their pole-facing slopes, with no deposits observed on the equator-facing slopes. This image, with a resolution of 2.77 m/px, was acquired at $L_{\text{sun}} = 195.04^\circ$, which corresponds to the beginning of spring in the southern hemisphere. A detailed examination of one of the dunes (Figure 1b) reveals a set of small linear features trending downslope from the zone of frost accumulation towards the base of the dune (Figure 1d). Similar features of the same length and width are observed on neighboring dunes in the same orientation and geometry, and lighting from the northwest reveals these features to be channels incised into the dune face. These channels are no more than 10 m wide and they extend for no more than 175 m from the lower margin of the frost deposit towards the base of the dune, and they frequently terminate before reaching the floor. The easternmost of the channels in Figure 1b shows a high-albedo deposit at its terminus, presumably frost.

MOC image E17/00566 (Figure 1c) shows the same dune face, but in a different season. This image, with a resolution of 4.36 m/px, was acquired at $L_{\text{sun}} = 24.74^\circ$, which corresponds to the beginning of summer in the southern hemisphere. The frost deposits observed in the spring are no longer present, revealing an undulating topography to the underlying terrain. This texture is only present on poleward slopes, where frost is found in the spring. The channels observed in the spring are present, though they are at the limit of this image’s resolution and are barely visible.

The only other high-resolution data available of this dune is THEMIS image V09998003, which shows the dunes at $L_{\text{sun}} = 5.392^\circ$, also at the beginning of southern autumn. The resolution of this image (35 m/px) is insufficient to observe the channels, but it is important to note that no frost deposits are found on the poleward or equatorward slopes of the dunes.

**Discussion:** Despite the difficulties in analyzing the precise morphologies of the observed channels at limited resolutions, these features appear similar to well-documented larger gullies found on other dunes in the mid-latitudes of Mars [5]. Unlike gullies carved into other surfaces, gullies on dunes do not emanate from broad alcoves, they maintain very consistent narrow and straight courses, and frequently terminate before reaching the base of the dune [5]. Additionally, these features frequently form well-defined ridges at their terminus instead of depositing debris fans [5]. Higher resolution data will be necessary to study the channels in question in detail, but the visible morphology suggest that they are similar in nature to larger channels incised into dunes.

We feel that the association between frost deposits on pole-facing slopes and the occurrence of channels exclusively on the same slopes is suggestive that these channels have formed from the melting of these frost deposits. Efforts to model the present-day martian climate [6] have predicted that poleward slopes in the mid/high-latitudes of Mars are where gullies would form if they are active, and this location meets those criteria. Frost on poleward slopes at this latitude remains stable on the surface, while frost on equatorward slopes is exposed to greater direct solar insolation, increasing the temperature enough to induce sublimation. By early spring, when Figure 1b was obtained, the frost is still stable and is potentially beginning to melt and carve the channel. By the beginning of autumn, when Figure 1c was obtained, the frost has all been removed to reveal the underlying terrain.

Due to resolution constraints, the most important unanswered questions with regard to these features cannot currently be addressed. The composition of the frost deposits is unknown and the resolutions of TES and OMEGA are insufficient to obtain accurate spectra for the target. CRISM, part of the Mars Reconnaissance Orbiter (MRO) instrument payload, will obtain spectra at a spatial resolution as high as 18 m/px, which will be sufficient to determine the composition of the frost. Additionally, any seasonal
modification of the channels cannot be observed at MOC resolution. HiRISE, also part of the MRO instrument payload, will obtain visible wavelength images at a resolution as high as 30 cm/px. We feel that seasonal monitoring of this site could provide critical information with regard to frost deposition in the mid/high-latitudes, channel formation on dune faces, and the stability of liquid water on the present-day martian surface.


Figure 1. a) MOLA context map showing the host crater. Arrow refers to location of Figures 1b-d. b) Subframe of MOC image M04/03432 showing a frost-covered dune with channels running downslope in early spring. c) Subframe of MOC image E17/00566 showing the same dune in early winter. d) Sketch map for Figure 1b.
ENHANCED BACKSCATTERING OF POLARIZED LIGHT: EFFECT OF REGOLITH PARTICLE SHAPE ON OPPOSITION BRIGHTNESS PEAK EXHIBITED BY SOME SOLAR SYSTEM BODIES.

J. M. Dlugach and M. I. Mishchenko, Main Astronomical Observatory of the National Academy of Sciences of Ukraine, Zabolotny str. 27, 03680 Kyiv, Ukraine; dl@mao.kiev.ua, NASA Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025, USA; crmim@giss.nasa.gov.

1. Introduction

Photometric phase curves obtained for some atmosphereless bodies of the Solar system show a steep increase in brightness near the phase angle \( \alpha = 0^\circ \). Evidently, several physical mechanisms may be involved in the formation of such a brightening. In particular, it was suggested that for a wide class of astronomical objects, coherent backscattering of light can play a role in producing this phenomenon (see, e.g., [1-4]). Because coherent backscattering from discrete random media is known to give rise to a triangular shape near the opposition and half-width of few tenths of a degree, it is worthwhile to apply this mechanism to interpret the astronomical observations. In our papers [5-7], we have shown that coherent backscattering of sunlight from regolith layers composed of submicrometer-sized spherical grains can explain the opposition effect exhibited by Saturn’s A- and B-rings, Galilean satellite Europa, bright asteroids 44 Nysa and 64 Angelina, and some regions on the Martian surface. But, it is interesting and important to examine how the characteristics of coherent backscattering can be affected by particle nonsphericity, which is the purpose of this work. We present and analyze the results of our computations performed for semi-infinite homogeneous regolith slabs composed of polydisperse, randomly oriented oblate spheroids with varying degree of asphericity.

2. Basic formula

Let the scattering medium be a plane-parallel slab composed of randomly distributed, independently scattering particles. This slab is illuminated by a parallel beam of light incident in the direction \((\theta_0, \varphi_0=0)\), and \(S\) is the Stokes reflection matrix for exactly the backscattering direction \((\theta, \pi)\). We will also specify the direction of incidence by the couplet \(\{\mu \geq 0, \varphi = 0\}\), where \(\mu = -\cos \theta\). For a macroscopically isotropic and mirror-symmetric scattering medium, the matrix \(S\) has the block-diagonal form [8],

\[
S = \begin{bmatrix}
S_{11} & S_{12} & 0 & 0 \\
S_{12} & S_{22} & 0 & 0 \\
0 & 0 & S_{33} & S_{34} \\
0 & 0 & -S_{34} & S_{44}
\end{bmatrix}
\] (1)

In accordance with the microscopic theory of coherent backscattering [9], the matrix \(S\) can be decomposed as

\[
S = S^l + S^M + S^C
\] (2)

Where \(S^l\) is the contribution of the first-order scattering, \(S^M\) is the diffuse multiple-scattering contribution composed of all the ladder diagrams of orders \(n \geq 2\), and \(S^C\) is the cumulative contribution of all the cyclical diagrams. The matrices \(S^l\) and \(S^M\) can be found by solving the vector form of the Ambarzumian’s nonlinear integral equation [10]. Then the matrix \(S^C\) can be obtained from the exact relations [11]. The backscattering enhancement factor (the amplitude of the opposition spike) \(\zeta\) for exactly backscattering direction defined as the ratio of the total backscattered intensity to the incoherent (diffuse) intensity is given by:

\[
\zeta = \frac{I_{\text{diffuse}} + I_{\text{coherent}}}{I_{\text{diffuse}}}.
\] (3)

For the case of unpolarized incident light, the amplitude of the opposition effect is described as follows [11]:

\[
\zeta = \frac{S_{11} + 0.5(S_{11} + S_{22} - S_{33} + S_{44})}{S_{11} + S_{11}^M}
\] (4)

To determine the elements of the matrix \(S^M\), one must first calculate the elements of the normalized Stokes scattering matrix for the particles forming the medium. In this study, we have used the method that was developed in [12] and is based on Waterman’s \(T\)-matrix approach [13]. Then the elements \(S_{11}^M\), \(S_{22}^M\), \(S_{33}^M\) and \(S_{44}^M\) were computed by means of a numerical solution of Ambarzumian’s nonlinear integral equation as described in [10].

3. Numerical results and discussion

To model the potential effect of particle nonsphericity on the amplitude of the opposition spike \(\zeta\), we have chosen randomly oriented oblate spheroids distributed over surface-equivalent-sphere radii \(r\) according to the power law. The shape of a spheroid is fully described by just one parameter, the aspect ratio \(E\) (i.e., the ratio of the larger to the smaller spheroid axes), along with a designation of either prolate or oblate.

We have performed computations of the amplitude of the opposition spike \(\zeta\) for a semi-infinite homogeneous slab composed of spheroids with the real
part of the refractive index \( m_R = 1.2, 1.4, \text{ and } 1.6 \), the imaginary part of the refractive index \( m_i = 0 \) and 0.01, a range of values of the effective size parameter \( x_{\text{eff}} = 2\pi r / \lambda_i \) (\( \lambda_i \) is the wavelength of the incident radiation in the surrounding medium), and aspect ratios \( 1 \leq E \leq 2 \). The effective variance of the size distribution \( \nu_{\text{eff}} \) was fixed at 0.1. In our opinion such wide region of adopted parameters gives possibility to analyse possible influence of nonsphericity for a number of various submicrometer-sized grains covering the surface layers of different Solar system bodies at a wide spectral interval (from UV up to IR).

The main results of our computations are shown in the form of color diagrams of the amplitude of the opposition spike \( \varsigma \) as a function of the effective size parameter \( x_{\text{eff}} \) and aspect ratio \( E \) for \( \mu = 1, 0.642, \) and 0.156. Let us first analyze the case of conservative scattering, \( m_i = 0 \) (Fig. 1). One can see that the calculated values of \( \varsigma \) lie in the range of 1.3 – 1.7. In the case of \( m_R = 1.2 \), the amplitude of the opposition effect does not depend on the shape of particles. The dependence on the value of \( E \) increases with increasing real part of the refractive index and/or with increasing effective size parameter (i.e decreasing wavelength). But it is not a monotonous function of \( m_R \), the aspect ratio \( E \), and effective size parameter \( x_{\text{eff}} \). The maximum value of this dependence is seen for \( E \sim 1.4, m_R = 1.6 \), and it does not exceed 20%.

Fig. 1. Amplitude of the opposition effect versus effective equal-surface-area-sphere size parameter and aspect ratio for \( m_R = 1.2 \) (left-hand column), 1.4 (middle column), and 1.6 (right-hand column), and \( \mu = 1 \) (top row), 0.642 (middle row), and 0.156 (bottom). The imaginary part of the refractive index is fixed at \( m_i = 0 \).

Fig. 2 shows the result of computations of the amplitude of the opposition spike \( \varsigma \) in a case of absorbing particles (\( m_i = 0.01 \)). The most obvious effect of increasing absorption is a significant decrease in the values of \( \varsigma \) and an appearance of weak dependence on particle asphericity also for \( m_R = 1.2 \). But also as in a case of nonabsorbing particles, the maximum difference in values of \( \varsigma \) caused by different particle asphericity, does not exceed 20%.

Fig. 2. As in Fig. 1, but for \( m_i = 0.01 \)

4. Conclusions

Using the model of a semi-infinite homogeneous slab composed of randomly oriented, polydisperse oblate spheroids with varying aspect ratios, we have demonstrated that for unpolarised incident light, the amplitude of the opposition effect caused by coherent backscattering depends very weakly on particle shape.

References
DEEP RADAR SOUNDING OF MARTIAN POLAR DEPOSITS: RADIATIVE TRANSFER MODELS.
Ya. A. Ilyushin, Moscow State University, Russia 119992 GSP-2 Moscow, Lengory, Physics Faculty, Atmospheric Physics Department e-mail: ilyushin@phys.msu.ru

Introduction: Investigation of the internal structure of martian polar caps, being a challenging task during past decades [1], is now going to be solved with the ground penetrating radar (GPR) instruments, both orbital and landed. In the present report the propagation of ultra wide band (UWB) chirp pulse in martian polar ice caps is investigated. The specific nature of the problem is that rather inhomogeneous layered structure of the caps lead to formation of diffuse structure of the radar signal field. Application of coherent radar signal processing techniques fails to resolve individual internal scatterers within the bulk of the polar caps, so that some incoherent interpretation technique should be utilized. In the present work the results of application of radiative transfer theory to the problem of radar sounding of the martian polar deposits.

Electrical model of the caps: we essentially exploit the electrical model of the caps, previously developed in earlier papers. It has been repeatedly argued that the primary constituent of the northern polar cap is water ice. The following observations provide basis for this statement: high water vapor concentrations over the cap during summer [2], low density [3], a thermal inertia [4] and albedo [5] corresponding to dirty water ice. The surface of the southern polar cap does not look like water ice, but there are evidences that water ice is its major volatile constituent, too [6,7].

There are also numerous observational evidences of layered structure of Martian polar caps [1]. Thickness of individual layers has been estimated to be 14 – 45 m [8] and 80-120 m for northern and southern polar caps, respectively. Recent studies of layered structure of northern polar cap [10] have revealed more complicated structure, largely consistent with earlier rough estimates. Precipitational and sedimentational models predict that layers of dirty ice are separated from each other by a dusty cover about 1 m thick [3]. Such cover is formed from the layer material due to ice ablation, which occurs periodically according to variations of Martian orbit parameters.

A model of Martian polar caps as a stack of layers of two types, the so-called “icy” and “dusty” layers, has been previously suggested [11]. It has been shown there that only electromagnetic waves at frequencies below 1 MHz can propagate through such a layered medium without significant distortion. Above 1 MHz there are frequency bands where the medium is opaque. On the other hand, at frequencies below 1 MHz the northern cap is relatively transparent. However, most currently developed and previously proposed radar instruments operate well above 1 MHz.

Radiative transfer model: The basic idea of this part of the study is to apply the non-coherent theory of radiative transfer to the problem of subsurface radar sounding of martian polar layered deposits, which are the random layered media. Within the approach applied in the present paper, the compressed radar pulse is regarded as a short plane wave packet, reflecting from, and penetrating through, the interfaces between layers. It is correct from the mathematical point of view because, for the simulation purposes, the compression of the pulse and wave propagation through medium can be interchanged. Assuming the layered structure of polar caps to have many internal parallel interfaces reflecting the sounding wave, one can consider the limiting case of continuously scattering medium. The theory of radiative transfer is a well developed tool for treating such problems. In recent papers [12,13] this theory has been applied to the considered problem for orbital and landed radar GPR instruments, respectively.

![Simulated signals](image-url) Simulated signals [12] for the northern polar cap are shown in the figure. Solid and dashed lines represent the exact solutions of electromagnetic equations and the asymptotic solutions of the radiative transfer theory. Mean loss tangents of the dust are labeled nearby the curves.

Conclusions: a new approach to the problem of radar sounding of martian polar deposits, based on the radiative transfer theory, has been developed. Numerical simulations, showing validity of the developed approach, have been performed.

MAPPING OF ADSORBED AND BOUND WATER IN MARS REGOLITH BASED ON MARS EXPRESS/OMEGA DATA. N.A. Evdokimova,\textsuperscript{1,2} R.O.Kuzmin,\textsuperscript{1} A.V.Rodin,\textsuperscript{1,2} A.A. Fedorova,\textsuperscript{1} J.-P.Bibring\textsuperscript{3} and OMEGA team,\textsuperscript{3} Institute for Space Research, RAS, Moscow, 117997, Russia, nadca@mail.ru \textsuperscript{3}Moscow Institute of Physics and Technology, Dolgoprudny, 141700, Russia, \textsuperscript{3}Vernadsky Institute of Geophysics and Analytical Chemistry, 119991 Moscow, Russia, \textsuperscript{3}Institut d’Astrophysique Spatiale, Batiment 121, 91405 Orsay Campus, France.

Introduction: OMEGA is a mapping spectrometer operating in visible and NIR spectral range (0.38-5.1\(\mu\)m). The instrument, inherited from the unsuccessful Mars-96 mission, is dedicated to the identification of molecular and mineral composition of Martian surface and atmosphere by means of spectral analysis of the outgoing reflected solar radiation\cite{1}. The instrument acquires spectra in 352 spectral channels, provided by three detectors: visible (0.38-1.05\(\mu\)m) and two infrared ones (0.93-2.73 \(\mu\)m and 2.55-5.11\(\mu\)m). Imaging capability is reached by combination of spacecraft orbital motion and transversal sweeping of the Field Of View (FOV) by means of special scanning device. OMEGA provides surface coverage at mean spatial resolution of 0.3-5 km from the orbit of 300-4000 km elevation, which allows mapping of minerals and volatiles on the Martian surface. The latter includes ice, as well as bound and adsorbed water, as the working spectral range involves important bands of CO\textsubscript{2}, H\textsubscript{2}O (including adsorbed and bound phases), oxides, hydrates, and other minerals\cite{1,2}.

Correction on the atmosphere: Since spectral features of interest (1.41\(\mu\)m, 1.46 \(\mu\)m, 1.5 \(\mu\)m, 1.91 \(\mu\)m etc.) overlap molecular CO\textsubscript{2} absorption bands at 1.4 \(\mu\)m and 1.9 \(\mu\)m, their analysis requires accurate subtraction of the atmospheric contribution to the observed spectral radiances. To correct OMEGA spectra on atmospheric absorption, we have chosen \textit{ab initio} approach and calculated several atmospheric transmittance spectra for mean temperature profile and pressures derived from the European Mars Climate Database \cite{3} for the seasons, local time and geometry corresponding to OMEGA observations.

An approximate estimate of the atmospheric contribution to the observed radiance is a convolution of the monochromatic transmittance with instrumental function. According OMEGA calibrations, a symmetric trapezoid with top width and both wings each equal to one pixel has been adopted.

Results: The corrected spectra were analyzed in order to derive features corresponding to condensed phases of water \cite{4}. As a first step, qualitative spectral index have been adopted as a measure of band depths and accordingly – the amount of absorbed in the Martian soil. For narrow bands at 1.41 \(\mu\)m and 1.46 \(\mu\)m, the spectral index are equal to band depths normalized by corresponding continua, so that the maximal band depths corresponds to the maximal index; for broader 1.5-\(\mu\)m water ice band, the index equals to the integral reflectance within the band normalized by integral reflectance in the band region with the spectral continuum approximated by linear function. In the latter case, maximal value of absorption corresponds to minimal index value. The examples of mapping these indices on the orbit 941 are presented in Figure 2. The orbit is suitable for testing mapping techniques because the frame contain such

![Figure 1](image-url)

**Figure 1.** Instrumental function (a) and simulated transmittance (b) used to correct OMEGA spectra on atmospheric absorption. \(n\) is pixel number, \(\Delta\lambda\) - distance between pixels in spectral scale.
Figure 2. Maps of condensed phases of water derived from OMEGA data. Orbit 941, latitude range 67°-84°N, Ls=100.2°. Shown are spectral index based on relative depths of related spectral features. (a) 1.46 μm adsorbed water feature; (b) 1.41 μm bound water feature; (c) water ice absorption at 1.5 μm (inverted); (d) reference plot: albedo at 1 μm. (a) and (b) contain spatial structure not related to local topography; they may be indicative to the mineral composition of the Martian surface.

Different morphological features as ice deposits of the Northern polar cap, near-cap aeolian dusty deposits, and seasonal inventories of ground water outside the cap. Upper maps (a,b) shows adsorbed and bound water spectral index. The area of maximal index of the adsorbed water (a) coincides with the belt of dust deposits around the polar cap and may reflect the extended adsorbing capacity of dusty surface, as well as a result of intensification of the nighttime water deposition onto dusty areas due to lower thermal inertia. The map of index reflecting distribution of hydrated minerals (b) shows little correlation with that of adsorbed water;

DELTAS FORMED IN AN ANCIENT CRATER LAKE IN THE NILI FOSSAE REGION OF MARS.
C. I. Fassett and J. W. Head, Dept. of Geological Sciences, Brown University, Providence, RI. 02912 (Caleb_Fassett@brown.edu).

Introduction: Despite the vast amount of new information which has been learned about Mars over the past ten years, there remains substantial uncertainty about the nature of the early Martian surface environment and climate. In particular, the record of valley networks which incise the ancient highlands has led to the idea that liquid water was stable on the surface and that early Mars was warm and wet [e.g., 1], though this is disputed [e.g., 2]. In part, it has been hard to distinguish between various models for the conditions that existed when valley networks were formed, because there have been few unambiguous examples of sedimentary deposits directly associated with valleys that have been observed.

Recently, Malin and Edgett [3] and Moore et al. [4] discovered a spectacular example of fluvial deposits associated with small valley network systems around NE Holden crater, which has provisionally been renamed Eberswalde. The deposits observed in Eberswalde provide a direct record of sedimentary emplacement, though it is not obvious whether these were formed in a lacustrine environment or as alluvial fan deposits and there is continuing disagreement on this point [5, 6]. Distinguishing between subaerial and subaqueous modes of emplacement given a sedimentary fan record can be difficult in the absence of other geological constraints.

We have recently described two sedimentary fan deposits (~56 km² in areal extent) in a 40-km diameter unnamed crater (centered at 77°40'E, 18°25'N) in the Nili Fossae region of Mars (Figure 1) [7]. As we discuss in detail below, compelling geological evidence exists that these fan deposits formed as lacustrine deltas. To examine these deposits, we gathered coregistered THEMIS, HRSC, MOC, and MOLA data of the Nili Fossae region in the ArcMap GIS environment.

Observations: Input Valleys. The fan deposits we observe in the Nili Fossae region are fed by ~200-km and ~80-km long sinuous valley networks. These drain a region of 15000 km², which is substantially larger than the region drained by the valleys that feed the Eberswalde crater. The drainage density of these input valleys is between .026 and .044 km⁻¹, which is consistent with what is observed using new data in other Noachian regions of Mars [see, e.g., 8], which is greater than was thought to be typical based on mapping using Viking [9], though typically smaller than what is found on Earth. The large, distributed drainage area of these input valleys that deposited the fan deposits suggests that they likely formed due to precipitation and surface runoff.

Fan Deposits. The fan deposits at MOC scale are shown in Figure 2. We interpret inverted channel deposits, based on the presence of what we identify as cross-cutting ridges that we believe are inverted channel segments (Fig. 2a). There is also extensive layering evident (Fig. 2b) which suggests the fan was episodically active, at least in a given location.

Geological Context and evidence for a crater lake. As seen in Figure 1, the eastern rim of the crater is substantially incised by a large valley. The minimum elevation of this valley (~2395 m) is above that of the fan deposits and most of the crater floor. Based on these elevations, this outlet valley appears to have formed after a substantial crater lake was formed and the eastern rim was breached. The outlet valley then incised the eastern rim by ~100 m to its present elevation.

Based on the present topography, we can estimate the volume of water that would have been required to fill the crater. If we flood the current topography to the -2320 meter level, we the volume of water would be at least ~350 km³. This provides a direct estimate of the minimum amount of water that must have gone through these valley systems, though the actual volume involved could have been much greater.

Implications: The biggest differences between the Nili Fossae fan deposits and those found in Eberswalde crater is that the Nili Fossae deposits appear to clearly been deposited in a lacustrine environment. We also have the ability to directly estimate a strong constraint for the minimum water volume involved, given that the crater had to be filled to breach the eastern rim. With this minimum volume of ponded water and estimates of the flux of water into the crater, we can obtain estimates for the minimum length of time over which the valley networks must have been active.

Water flux estimates. We have used Manning’s equation scaled for Martian gravity as well as empirical relationships [10] to estimate the likely discharges of the input valleys to the crater. Although such estimates are subject to substantial uncertainty [10], we believe that the best estimate of the channel-forming flux for these valleys is ~700 m³/s.

Minimum Formation Time and Intermittency. If flow was maintained at channel-forming conditions constantly (and water was not lost to infiltration or the
atmosphere), the minimum water volume could be reached in ~16 (Earth) years. However, there are several reasons to believe that such an estimate is too small (perhaps far too small). First, maintaining channel-forming discharges for such a substantial length of time seems extremely difficult, at least if the water results from precipitation, given that such flow conditions are equivalent to ~0.5 cm/day over the entire watershed. On Earth, the percentage of time where channel-forming discharges are maintained is generally small [11].

**Qualitative indicators of the formation time for the fan deposits.** It is difficult to derive quantitative estimates of the intermittency on Mars in absence of the knowledge of its past climate. Nonetheless, there are qualitative indicators that the valley systems discussed here were active for longer than the minimum possible length of time. First, the input valleys show significant signs of avulsion and meander evolution. Secondly, the present maximum evolution of the fan deposits is quite consistent with the minimum breach elevation and none of the fan deposits appears to have been stranded above this minimum stand. This suggests the fan surface had enough time to adjust to the drop in lake level as the outlet valley cut down the eastern rim. Third, OMEGA data reveals the presence of clays as well as hydrated minerals in the watershed of the input valleys [12]. Though it is difficult to specify the minimum formation time for these clays without knowledge of their formation environment and the kinetics of their formation, the presence of such clays suggests water was available for a geochemically-significant amount of time.

**Conclusions:** Given the minimum formation time that we calculate and the qualitative evidence for persistent surface water in the Nili Fossae region, we believe the valley networks in this location were active for at least hundreds of years (consistent with peak flow < 20% of the time) and possibly much longer (intermittency < 1% are common in arid environments on Earth)[11]. Further work needs to be done to place more precise quantitative constraints on the valley network intermittency, as well as on the length of time over which liquid water played an important role on the early Martian surface.


**Figure 1.** THEMIS daytime IR Mosaic of the fan deposits deposited in the 40-km crater by two input valleys. The breach in the eastern rim strongly suggests that the crater was ponded and formed a crater lake.

**Figure 2.** (left) Mosaic of MOC images R23-00833 and S04-00725, showing the pervasive layering (inset A) and cross-cutting inverted channel deposits (inset B) which we interpret to have been deposited in a lacustrine environment.
EROSION AND TERRAIN INVERSION IN NORTHEAST ARABIA TERRA. C. I. Fassett and J. W. Head, Dept. of Geological Sciences, Brown University, Providence, RI. 02912 (Caleb_Fassett@brown.edu).

Introduction: The northeast part of Arabia Terra directly west of Syrtis Major is marked by two widespread morphological units: dissected terrain (Npld) and etched terrain (Nple) [1]. In both of these units, but especially in the etched terrain, there is evidence for inverted topography, such as old craters and valleys that now lie above the surrounding plains. The formation of this inverted terrain requires infilling of formerly low-lying regions (gradation) and preferential removal of the former surrounding topography (erosion). Grant and Schultz [2] made a detailed examination of this region based on Viking data, with a focus on the timing of gradational and erosional episodes. Here, on the basis of new data, we reexamine the geological history of the etched and dissected units in northeast Arabia Terra, specifically focusing on a region from 45-55°E and 5-25°N, which is centered on the transition from the dissected unit (SW) to the etched unit (NE) (Fig. 1).

Regional/Unit Characteristics: Virtually the whole northeast Arabia Terra region has low thermal inertia (I<200 J/(K m²s²°K)) [3, 4], which is consistent with widespread surface dust or lightly indurated fines, and minimal surface rock abundance. The regional slope is from the south to north, averaging 0.07°. Figure 1 shows a THEMIS IR daytime mosaic of the region, and the boundary between Npld and Nple.

Basement material in both the Nple and Npld is quite ancient. Counting all craters on Viking data (including the most degraded ones), Grant and Schultz [2] derived a common N5 age of 3.5 (cumulative log₁₀ number of craters > 5 km in diameter per 10⁶ km²) for both units, with craters smaller than 20 km diverging from the expected production function. Based on their work, the crust in this region thus dates to the mid-early Noachian [5]. However, much of the surficial material has been substantially reworked subsequently.

Etched terrain (Nple). The etched unit is marked by the widespread (but not universal) presence of an eroded/eroding layered material (100-200 meters in thickness) which fills the interior of craters and valleys. In some locations, this fill material sits above its surroundings and terrain inversion has occurred (Fig 2.). Locally, the fill material is cracked and/or pitted. In general, its surface age is much less than the etched terrain as a whole, based on the relatively infrequent superposition of pristine small craters on the etched/fill material; in Grant and Schultz’s crater counts, this unit could be as young as the early Amazonian (N5 age 1.6-1.7), though continual (but perhaps declining) modification may have occurred since its emplacement. As the etched terrain has been eroded, ancient underlying material is likely exposed. There are only a few valley networks in the etched region.

Dissected terrain (Npld). The dissected terrain is marked by widespread erosion by dendritic valley networks. At regional scale, valley networks appear similar to those in other portions of highlands [e.g., 6], and probably date to the late Noachian like most highland valley networks (N5=2.7 [2]). In low elevation portions of the dissected units (predominately in degraded craters), crater fill material is present similar to what is observed in the etched terrain.

Transition: Given that the fill material, which is the most prominent characteristic of the etched terrain, extends into the dissected unit, it is tempting to think of the transition between these morphological units as gradational. However, new data illustrates that the transition between valleys and inverted valley forms is (at least locally) abrupt (see Fig 3.) Even more intriguingly, although valleys exhibit a graded profile (south to north) consistent with their erosion by surface water, the surface elevation of the inverted deposits that these valleys transition increases northward, counter to our expectations that these might mark the surface of old channel bed deposits (for example, like those in NE Holden (Eberswalde) crater, [8]).

Synthesis: Two primary questions that we seek to answer about NE Arabia Terra are: (I) What is the nature of the fill material? (II) What mechanism accomplished such widespread erosion (and inversion?)

We know that the fill material had to be (1) extensive but not universal (discontinuous); (2) preferentially deposited in preexisting low topography; and (3) more resistant to erosion than surrounding material. The hypotheses that we are currently testing include ashfall/tuff, distal ejecta, surface volcanics, and sediment emplacement. We are also focusing on the possible role of volatiles.

Our working hypothesis, following Grant and Schultz [2] and earlier workers [e.g. 1], is that the mechanism for etching/terrain inversion/exhumation was likely aeolian. A major question that remains regarding this hypothesis is the scale and breadth of the erosion needed—up to hundreds of meters of material must be removed over a region of several hundred thousand square kilometers. Though it is conceivable that this erosion may have been taking place over most of Mars’ geological history, this would still require average erosion rates several orders of magnitude higher (~25 nm/year) than those typical over Mars history [9, 10].

Figure 1. THEMIS IR mosaic of the NE Arabia Terra region (45-55°E, 5-25°N). The dotted line represents the boundary between Npld and Nple as mapped by Greeley and Guest [1]; the solid line marks where we would place the boundary on the basis of new data. The white box marks the location of Figure 2.

Figure 2. MOLA false color superposed on THEMIS IR daytime mosaic. Both a ~30-km-diameter crater and the valleys connected to it are now inverted; the fill material must have been more resistant to erosion than what once surrounded it.

Figure 3 (right). Example of the transition from typical highland valley networks to unusual inverted forms (MOC images R10-00363 & R20-001681 of valley at 50°9'E, 15°30 N; see also region around 46°45'E 10°12').

Introduction and Background: It has long been known that Mars contains polar caps and that they are comprised largely of water ice and dust [e.g., 1]. Less well-understood is the presence and nature of glacial flow in polar regions [e.g., 1], the possibility of glacial processes operating outside the polar regions [e.g., 2], and the mode of formation of circumpolar craters that contain significant high albedo mounds and deposits [e.g., 3]. Recent studies have shown that debris covered glacier-like deposits occur in mid-latitude impact craters [e.g., 4, 5] and that circumpolar craters (65°-80° latitude) currently contain remnant ice deposits some of which may be glacial in origin [e.g., 3]. Analysis of new Mars data and a better understanding of glacial processes in terrestrial hyperarid, cold polar deserts analogous to the Mars environment, have documented the presence of tropical mountain glaciers and their distinctive deposits [e.g., 6-7]. These latter studies have shown that many glacial deposits on Mars represent the process of cold-based glaciation [6], which produces a distinctive set of features known as drop moraines. These features, well displayed in the Antarctic Dry Valleys [6,8,11], form when sublimation from the glacial front and ice forward velocity are in equilibrium so that there is no net movement of the front. In this case ice and contained debris move forward to the margins of the glacier, the ice sublimes, and the debris falls out of the ice and drops to the front, forming accumulations known as ‘drop moraines.’ Periods of increasing sublimation or decreasing velocity can result in ice front retreat; a new drop moraine will form when equilibrium is again reached. In this manner, a series of drop moraines can form parallel concentric ridges marking the successive stands of the cold-based glacier margin. Dozens of such parallel ridges form drop moraines on the Arsia Mons tropical mountain glacier [6-7].

Most circumpolar craters that show evidence of icy fill have distinctive concentrations of ice around the central peak (e.g., Korelev), lobate deposits attached to polar layered terrain, or isolated mounds of material usually along the base of the pole-facing crater wall [e.g., 3, 9]. Here we report on a crater in the same 65°-80° latitude range, but with a distinctly different ice-related crater interior deposit. We describe the crater occurrence and characteristics, its distinctive deposits that we interpret to be remnant moraines, and the conditions and sequence of events implied in its origin and evolution. We conclude that this feature represents a cold-based glacier that formed as a result of snow and ice accumulation on the eastern rim of the crater due to localized environmental conditions, and that the cold-based glacier flowed down the crater wall, climbed the central mound and was passively diverted by them, and underwent several phases of advance and retreat. These deposits appear to be very young in age.

Description: The 26.8 km diameter crater at 70.32°N, 266.45°E (Fig. 1) is about 1.6 km deep and was formed in the Vastitas Borealis Formation. It is characterized by a somewhat degraded lobate ejecta deposit, a distinctive central peak rising about 500 m from the crater floor, and a flat floor (Fig. 1a). Topographic and slope profiles (Fig. 1b-c) show that the SE portion of the floor is shallower and lies about 250 m above the NW floor; crater walls are steeper on the SE than on the NW. THEMIS data reveal the presence of a complex set of ridges, tens to several hundred meters across, extending from near the SE crater rim crest, down the crater wall and out onto the crater floor (Fig. 2, 3). Along the crater wall the ridges are oriented generally radially to the crater, begin to become arcuate on the crater floor, and then display multiple, tight lobate patterns in the vicinity of the central peak (Fig. 3). Some ridges, particularly the marginal ones, are very continuous and extend as much as ~15 km from the crater wall down onto the crater floor. The pattern of ridges forms a pincher-like structure on and around the central peak (Fig. 3). Perspective views (Fig. 2) show that the set of ridges forms a contiguous deposit extending down the wall and out onto the floor, rising up onto the central peak summit and then bifurcating and forming two marginal...

Fig. 1. A) MOLA gridded altimetry data superposed on THEMIS image V05259017; B. Topographic profile along A-A’ (Fig. 1a); C. Slope along same profile.
lobes. The southern lobe extends ~2 km further than the northern one, actually rising up onto the base of the western crater wall. All concentric ridge structures are convex outward, away from the SE wall. Associated features include a few dark patches, interpreted to be colian deposits superposed on the lobate deposit; no evidence of additional structures such as fractures or channels, were observed. Superposition relationships show that the deposit largely overlies radial wall textures and deposits at the base of the walls (Fig. 2, 3). The internal stratigraphy of the deposit shows that the ridges form several sets of broad, continuous lobes of different sizes that are superposed on one another, often with no disruption of the underlying ridges, but in many cases apparently obscuring them due to deposition (Fig. 3). The superposition relationships imply several phases (at least 4) of successive lobe formation. In some cases, narrow concentric ridges are seen forming tight fold-like features (SW lobe of the pincher).

**Interpretation and Conclusions:** The ridges and the deposits are unlike structural wrinkle ridges, inverted streams, eskers, exhumed dikes, landslide scars, crater ejecta, or typical mass movements. They are most similar to ridges associated with cold-based glaciers on Earth [8,11] and Mars [4-7]. On the basis of the size and morphology of the ridges, their convex outward shape, their clear control by the topography of the wall and central peak, their discrete patterns, and their multiple superposition relationships, we interpret the ridges to be moraines associated with phases of advance of glacier lobes from the southeastern margins of the crater wall and rim. In this interpretation, snow and ice accumulated on the SE rim and wall, incorporated debris from the crater wall, and flowed down the wall and out onto the crater floor, riding up on the central peaks, and then bifurcating and extending out onto the NW crater floor. Changing climate conditions caused the retreat of the ice, leaving moraines and deposits of glacial till. Distinctive superposition relationships suggest that advance and retreat occurred several times, and sharp preservation of underlying moraines below later moraines suggests cold-based conditions [6-8, 10-11]. We are currently establishing the internal deposit sequence of events in detail and exploring the range of conditions that might favor the accumulation of snow and ice in this particular configuration, in order to explore links between these deposits and climate cycles.


![Fig. 2. Perspective views of crater interior (THEMIS V05259017 on MOLA altimetry). A. Looking NW. B. Looking SE.](image)

![Fig. 3. Crater interior wall and floor structure. A) Portion of THEMIS V05259017. B) Sketch map of the ridges and their relationships.](image)

Introduction: One of the most topographically distinctive regions of the global dichotomy boundary extends to the highest northern latitudes (30-50°N), Deuteronilus-Protonilus Mensae (DPM), from ~290-360°W. This region, unlike most others along the boundary, is characterized by fretted terrain, in which the northern lowlands give way to progressively larger upland mesas and ultimately to sinuous valleys that extend into the adjacent highlands [1] (Fig. 1). The mesas are surrounded by lobate debris aprons (LDA), largely attributed to mass wasting aided by periodic flow of ice in pore spaces derived from diffusion of atmospheric water vapor [e.g., 2]. The fretted valleys are broad, steep-walled, flat-floored and extend deep into the uplands; they contain lineated valley fill (LVF). The similarity of mass wasting from the walls of the fretted valleys and the lobate debris aprons surrounding the mesas has led many to the conclusion that LVF may have formed largely by mass wasting [e.g., 3]. In Viking images, the lineated valley fill that occupies these valleys was shown to be characterized by axial lines (ridges and grooves) on the valley floors, and it was noted that these could be caused by convergence, in the middle of the valley, of linear debris aprons derived from opposite valley walls [e.g., 2]. On the other hand, Lucchitta [4] showed examples in which wall ridges and grooves bend downvalley to form axial lines, suggesting down-valley flow and a more important role for ice-related processes. In an update using MGS MOC and MOLA data, Carr [5] pointed out that the fretted valleys are poorly graded and concluded that they initially formed by fluvial processes similar to other valleys on Mars, but were subsequently widened and filled by the same mass wasting processes that formed the debris aprons; furthermore, Carr [5] observed down-valley topographic slope reversals leading him to conclude that down-valley flow of lineated valley fill was minor. Here we use new data from the Mars Observer, Odyssey, and Mars Express spacecraft to study the nature of the fretted valleys and the contained lineated valley fill to analyze their mode of formation and emplacement and their modification histories.

We completed a reconnaissance study of several dozen fretted valleys in this region and chose a typical example for detailed analysis at ~34°E, 41°N in the central region between Deuteronilus and Protonilus Mensae (Fig. 1). Using MOLA altimetry, Viking, THEMIS, HRSC, and MOC images, we systematically analyzed the LVF from the narrow parts of the deeper upland plateaus progressively toward the northern lowlands. In the most inland regions of the fretted valleys analyzed (areas A and B in Fig. 1b) the valleys are narrow (~10-20 km in width), often have distinctively cuspatate (convex outward) shoulder-to-shoulder alcoves along the walls (region A) or larger tributary valleys often forming hanging valleys along the wall margins (region B). In contrast to the sharp parallel valley walls and parallel along-valley lineated fill typical of many fretted valleys described elsewhere [e.g., 2,3,5], these alcoves are sources of lineated valley fill (Fig. 1c) that can be seen to emerge from individual alcoves as lobes, descending onto the valley floor, bending and turning in a downslope direction, and merging with adjacent lobes, deforming in the process ultimately to form the distinctive along-strike ridges of LVF. A particularly good example of the transport of material from an alcove and its deformation and inclusion into the main along-valley lineations (Fig. 1c) is seen as area C merges with the flow emerging from area B. Here, convex-outward, concentric-lobed flows are merging with adjacent lobes and converging with the along-valley flow, becoming progressively compressed and folded until they form the parallel along-valley lineations that are the hallmark of LVF. These types of alcoves, the arete-like borders between them, the arcuate-ridged lobes extending from them, the progressive compression and folding of the lobes as they converge onto the valley floor and lose their identity, are all hallmarks of glaciated valley landsystems on Earth [6,7]. In the context of such valley glacial environments, we interpret the alcoves on Mars to represent microenvironments where ice and snow accumulated, alcove walls shed debris onto the ice, and debris-covered glaciers emerged and moved down slope into the main valleys, widening the alcoves and creating aretes. We mapped the trends represented by the emergence of lobate material from these alcoves (Fig. 1b, narrow arrows) and the general trends with which they merge into the larger-scale LVF (wide arrows). The patterns in areas A, B, and C clearly indicate the convergence typical of glaciated valley landsystems [6,7] where abundant small debris-covered glaciers contribute to larger-scale valley glaciers. As the generally north-trending LVF emerges from areas A-C (Fig. 1b), the individual LVFs merge into two main trends, areas D and E. LVF C merges with B and continues ~95 km to the NNW where it bifurcates around a mesa to form LVF F and G, each continuing 50-75 km to the northern lowlands. Western portions of LVF B turn west into D, merging with LVF A to produce complex fold patterns. Portions of A turn NW and extend almost 100 km to the northern lowlands.

Several basic trends are observed over this region extending from A-B-C to the northern lowlands. First, valley widths in the north tend to be wider (up to a factor of 2-3) than in the southern regions. Valley wall alcoves are seen virtually throughout the system and continue to contribute material to the LVF through individual and adjacent lobate flows (Fig. 1c), even along the most distal large mesa. The distribution of the alcoves tends to be preferentially on north pole-facing slopes (Fig. 1b), but they occur with all orientations. The elevation of the LVF floor decreases by ~1500-2000 m from the proximal areas (areas A-C) to the end of the LVF in the northern lowlands. LVF slopes tilt toward the north, are less than 1° throughout most of the LVF floor, but increase rapidly to ~1-2° in the northern 25-50 km of the LVF, adjacent to the lowlands. These latter areas are characterized by piedmont-like lobes extending into the northern lowlands between mesas. These latter features are very reminiscent of the fronts of terrestrial glaciers in terms of their lobate nature and relatively steep front, and in many cases are characterized by flow lines, pits and moraine-like ridges.

These general trends are accompanied by what appears to be a progressive fragmentation or "mesa-ization" of the extensive upland plateau. The large mesa between LVF areas A...
and B (Fig. 1b) is barely connected to the plateau itself at its southern end, but to the north, widening of the valleys and enhancement of the alcoves has fragmented plateau segments into smaller and smaller mesas. Although the summits of the three mesas directly north of the plateau are the same, the size of the mesas has decreased substantially. This trend is consistent with the effects of the interpreted valley glaciation, in that the debris-covered glaciers in the alcoves along the margins erode progressively into walls, causing their retreat and widening over time. The lateral migration can form aretes and ultimately cut headward through narrow parts of the plateaus to form isolated mesas.

In summary, examination of new spacecraft data shows compelling evidence for an integrated picture of LVF formation, with a significant role being played by regional valley glaciation [e.g., 6,7] in the modification of the valley systems. We find evidence for: 1) localized alcoves, which are the sources of dozens of narrow, lobate concentric-ridged flows interpreted to be remnants of debris-covered glaciers; 2) depressions in many current alcoves, suggesting loss of material from relict ice-rich accumulation zones; 3) narrow pointed plateau ridge remnants between alcoves, similar to glacially eroded aretes; 4) horseshoe-shaped ridges up-valley and upslope of topographic obstacles; 5) convergence and merging of LVF fabric in the down-valley direction, and deformation, distortion and folding of LVF in the vicinity of convergence, all consistent with glacial-like, down-valley flow; 6) pits and elongated troughs in LVF, particularly in areas where LVF is distorted, suggesting localized preferential sublimation of an ice-rich substrate lying below a rocky sublimation till; 7) distinctive lobe-shaped termini where the LVF emerges into the northern lowlands, with associated pitting; we interpret these to be the location of the relict glacial system margin. This assemblage of features can be traced and mapped in an integrated fashion (Fig. 1b) for over 200 km in length, and covers an area (~30,000 km²) ~10 times that of the Antarctic Dry Valleys. The extensive development of lineated valley occurrences with very similar characteristics in fretted valleys in the DPM region suggests that glaciation may have been instrumental in the modification of the dichotomy boundary over an area as large as several million km². The age of the LVF in different parts of the DPM region has been estimated as Amazonian [e.g. 8,9]. The process of glaciation has clearly contributed to the erosion of the dichotomy boundary, broadening and extending valleys, and creating isolated mesas from glacial erosional activity, suggesting retreat over distances of many hundreds of kilometers.


Figure 1. Fretted channel and lineated valley fill system in the central DPM region. a) Viking Orbiter image mosaic and location map b) Map showing the location and trend of small alcoves and lobate flow-like features (narrow arrows) and broad flow trends of LVF on the valley floor (wide arrows). c) Outflow of two ridged lobes from alcoves (bottom left and right); lobes join major LVF of area C (upper right) near convergence with B (Fig. 1b). Left lobe is swept westward, forming broad arcuate folds; right lobe is increasingly compressed until it resembles a tight isoclinal fold. Both lobes ultimately merge into the general LVF parallel to the valley walls. MOC R1702578.
**Introduction:** On Mars, the presence of linear volcanic vents, narrow graben (the surface manifestation of near-surface dike intrusion) [1], and narrow linear ridges (the erosional remnants of near-surface dike systems) [2] represent the best evidence for the nature and distribution of dike systems. We report here on the discovery and documentation of a laterally and areally extensive set of narrow ridges that are interpreted to be the near-surface manifestation of a major dike system emplaced in the Late Noachian-Early Hesperian period of Mars history and associated with one of the most widespread magmatic and volcanic events in the history of Mars. Hesperian ridged plains (Hr) are "broad planar surfaces, rare lobate deposits, and long, parallel linear to sinuous mare-type (wrinkle) ridges..." interpreted as "Extensive lava flows erupted with low effective viscosity from many sources at high rates.". The paucity of distinctive flow fronts that might confirm a volcanic origin, and the weakness of secondary evidence to support a volcanic origin (such as the presence of wrinkle ridges) has led to some uncertainty in the origin of Hr (5). Furthermore, in regions where volcanic edifices are observed (e.g., Hesperia Planum) there is evidence for pyroclastic activity, rather than effusive volcanic activity, playing a major role (e.g., 6, 7). Because of its global distribution and significance in the history of Mars, we have undertaken a regional and global assessment of Hr as a means to assess evidence for the thermal, volcanological and climatic evolution of early Mars (e.g., 8, 9). In the course of examination of exposures of Hr in the area north and east of Huygens Crater (Fig. 1), we discovered a series of linear ridges associated with Hr. We used MOLA, MOC, THEMIS and HRSC data to trace the surface exposure of these ridges and characterize their nature and relationship to other units.

**The Nature and Distribution of the Linear Ridges:** At least fourteen exposures of narrow linear ridges comprising two major ridge systems have been detected (Fig. 1). Each ridge system is broadly arcuate to slightly sinuous in its surface pattern and remarkably consistent in character over its lateral extent, which although discontinuous, extends for distances of 575 and 700 km, perhaps extending further in the near subsurface. The nature of the ridges is remarkably consistent over their length, as revealed in THEMIS and HRSC data. The width of the ridges range from ~400-800 meters but is typically ~600 meters. The height of the ridges, as revealed by numerous MOLA PEDR profiles, averages less than ~30 meters. The broad ridge seen in the THEMIS and HRSC images is caused by flanking erosional debris surrounding a thin central ridge. The ridge system does not change detailed character as a function of local topography. As ridge segments climb broad wrinkle ridges or descend into craters their width and height remain virtually unchanged. The general shape of the ridge strike is linear, but in regional view it is broadly arcuate. Locally the ridge can meander slightly, but there is no evidence to date for multiple ridges, braiding, bifurcation, sinuosity or tectonically disrupted patterns along strike. The ridge is in contact with several southern upland geological units [4]. The ridges are seen to be both superposed/cross-cutting these plains units and embayed by them. Impact craters and their ejecta are superposed on the ridges as well as cut by them. In some cases there is evidence that lateral emplacement of ejecta took place subsequent to the ridge formation, as ejecta is preferentially piled up in the vicinity of the ridge.

**Origin of the Linear Ridges:** Of the several processes that might be responsible for the ridges, we find that the most likely origin is related to dike emplacement events, caused by the propagation of magma-filled cracks in the crust, to form long linear fractures filled with magma that solidifies and can leave a surface ridge following exhumation (e.g., 10). Dikes are generally vertical or near vertical in nature, cross terrain irrespective of topography, are in the range of meters to hundreds of meters in width, are extremely linear and often broadly curving, and can extend in the near subsurface for thousands of kilometers (e.g., 1, 10-12). The geometry and extent of the ridges provides important constraints on the nature of the dikes and the eruption conditions that might have accompanied their emplacement. The broad ridge in lower resolution images is typically ~600 m across and the details revealed in high-resolution images suggest that the surface exposure of the ridge crest is 50-100 m. This implies that the actual width of the dike at depth might range up to ~200 meters. These values indicate that the depth from the original surface to the top of the dike would range from zero (the part that erupts) to as much as 1000-1500 m [1]. If, for example, an eruption was active along a 5 km long segment of the dike (less than 1/100 of the total dike length), the rise speed of mafic magma in a dike ~100 m wide could be up to 20 m/s, and the volume flux (velocity x width x length) could be up to 10 x 10^9 m^3/sec [12]. Flows erupting from such vents would travel over flat to gently sloping terrain for hundreds to thousands of km [12], or in rougher topography, pond in local to regional lows such as craters, burying the vent and the associated eruptive products in the process. In summary, on the basis of models of the ascent and eruption of magma on Mars and the structure of its crust (e.g., 1, 12), dike widths of this magnitude are likely to involve very high effusion rate and high volume eruptions, in the range of those typical of flood basalts on Earth. These models are completely consistent with the patchy nature of Hr plains in the cratered uplands and the dearth of individual flow fronts.

**Summary and Conclusions:** Two major low, narrow broadly arcuate linear ridges in the region of western Terra Tyrrenhia northeast of Huygens crater cross Noachian terrain for distances of hundreds of kilometers irrespective of topographic and geologic unit changes. The longest of these is over 750 km in length, is remarkable in its continuity and consistent nature, and its exposure relationships. We find that an origin of these ridges as the surface manifestation of near-surface magmatic intrusions (dikes) is the most plausible. The consistent width, vertical orientation, linear continuity, and broad linear-arcuate nature are all consistent with dike emplacement. Detailed, along-strike stratigraphic relationships show that the ridges cut some Hesperian-aged ridged plains but are embayed by others. We thus interpret these features to be associated with the emplacement of the nearly globally distributed Early Hesperian ridged plains. Hesperian plains have previously
been interpreted to have been emplaced volcanically, but the lack of clearly identifiable flow fronts and the presence of mantling sediments has often led to alternative hypotheses. The width and geometry of these dikes is consistent with very high effusion rate, high volume flood basalts eruptions, emplacement events that would result in the volcanic flooding of rough Noachian cratered terrain topography, likely leaving little evidence of flow fronts. The current nature and consistent exposure of the ridges strongly argues that they have been exhumed, with overlying material up to several hundred meters thick being removed. The present topographic relationships imply that exhumation was a very efficient process, and thus that the removed layer must have been very volatile-rich and fine-grained, susceptible to sublimation, possible aqueous erosion and eolian redistribution. The close linkage in time of the valley networks characterizing the dissected terrain and the emplacement of the volcanic plains suggests that volcanic degassing into the atmosphere may have changed the climate sufficiently to cause deposition and erosion of a volatile-rich cover.


Fig. 1. A) Distribution of ridges (thick lines) northeast of Huygens crater. B) Ridge crosses rim of an ancient 18 km rectangular crater. C) Typical character of the ridge system showing narrow and linear behavior. D) Ridge cutting across wrinkle ridges with a superposed impact crater with a pedestal-like elevated ejecta deposit. E) Ridge, knobs and associated flow-like lobes. F) Lack of deflection as the ridge passes across crater rim crest and down onto crater floor. G) Central portion of ridge covered by lobate ejecta from crater in upper right. H) MOC image of linear ridge. Note sharpness of ridge crest and material along sides of ridge, responsible for making it appear broader in the lower resolution THEMIS images (B-G).

Introduction: The melting of planetary interiors, segregation of melt products, ascent, intrusion and eruption of magma, and outgassing to the atmosphere are among the most fundamental processes in the thermal and geological evolution of terrestrial planets. Commonly, much of the record of these processes remains in the subsurface, and an understanding of the processes relies heavily on interpretation of surface deposits and exposures, which are often easily erodable, covered by their own effusive products, or are difficult to distinguish in a sea of similar deposits. Dike systems (solidified magma-filled cracks propagated from the source regions toward the surface) represent one of the key elements of the shallow subsurface manifestation of these processes, but are rarely exposed on planets with little differential uplift and erosion, such as the Moon, Mars, Mercury and Venus. On Mars, the presence of linear volcanic vents, narrow graben (the surface manifestation of near-surface dike intrusion) [1], and narrow linear ridges (the erosional remnants of near-surface dike systems) [2,3] represent some of the best evidence yet seen for the nature and distribution of planetary dike systems. Examples of dike emplacement events ideal for study would be those that 1) represent single emplacement phases, 2) form on non-volcanic deposits so that eruptive products associated with the dike emplacement event can be distinguished from previous material, and 3) are not covered by continued post-emplacement volcanism. Amazonian-aged deposits on the western flanks of the large Tharsis Montes volcanoes (Arsia, Pavonis, and Ascraeus Mons) have been recognized to be of glacial origin [4-7] and represent such a late-history non-volcanic baseline.

We report here on the documentation of a linear and laterally extensive cone and lava flow system (Fig. 1) interpreted to be the near-surface manifestation of a major dike emplacement event in the Amazonian, postdating the major glacial deposits and unmodified by subsequent volcanism. Analysis of this feature permits the assessment of eruption conditions associated with the emplacement of a single large dike in the Tharsis volcanic province.

Geological setting: The volcanic cones and flows shown in Fig. 1 occur just northwest of Arsia Mons, one of the major shield volcanoes of the Tharsis region. They are largely superposed on the Arsia Mons fan-shaped deposit, interpreted to be of glacial origin [e.g., 5], which provides a key baseline to help distinguish the dike emplacement event from other volcanic deposits of the Arsia and Tharsis region. The deposit itself consists of a row of cones and adjacent flows superposed on the ridged facies of the glacial deposit, thought to represent drop moraines from the previously emplaced cold-based piedmont glacier [5]. The strike of the row of cones is ~N5E, not radial from the center of Arsia. The row of cones and related flows has been traced for a distance of ~27 km in the fan-shaped deposit and several cones aligned along strike suggest that it extends for several tens of km to the NW of the fan-shaped deposit.

Within the fan-shaped deposit, two major developments of linear cones occur. The southern one, covered in a MOC image, consists of at least 44 aligned discrete cones over a distance of about 7.3 km. The second one, covered in a THEMIS image, is about 11.3 km in length and consists of about 50 cones. Individual cones are equidimensional to slightly elongate along the direction of strike and average less than 100 m in width and are typically 20-25 m in height. Several of the cones in each area of development are broader, up to ~600 m in width, and noticeably taller. Some cones, typically toward the ends of the rows, show no surrounding flow deposits, while others show evidence for flanking lava flows extending for distances of 2-3 km from the cones. The larger cones, some of which have summit craters, appear to be the focal point for more extensive flanking flows, extending as much as 5-15 km from the vicinity of the cones. The paths of these latter flows are controlled by 1) the regional NW slope on the flanks of Arsia, 2) topography of the underlying lava flows and the glacial deposits, and 3) local slopes associated with the ridges of the glacial deposits.

Nature of dike emplacement events and interpretation of eruption conditions: Commonly, when a dike is emplaced, it rises to the near surface along most of its length and the top stalls at a depth of a few meters to a few hundred meters, depending on local crustal density distribution, stress levels, and orientations, magma overpressurization levels and gas content and exsolution patterns [e.g., 8-9]. Often, a small portion of the dike intersects the surface, and creates a short-duration "curtain of fire" eruption developing into a longer-duration centralized vent (e.g., 10), forming a row of small spatter cones, a centralized tephra cone and flanking lava flows. The geometry of the cones and the typical distances of flanking flow features can be used to infer eruption conditions [e.g., 11-12]. On the basis of the geometry of the cones and flows described above, dike widths of less
than a meter are implied with rise speeds in the range of a few cm/s, comparable to those typical of Kilauea eruptions on Hawaii. Cone widths imply pyroclastic velocities of about 50 m/s, which could be generated by magma containing 0.1% exsolved water as the magmatic gas. On the basis of the length of the longest flows, the eruption duration is likely to have been of the order of days.

**Conclusion and implications:** This preliminary analysis shows that a dike emplaced in the Tharsis rise volcanic complex has characteristics similar to those of a typical terrestrial eruption at Kilauea, Hawaii. The scale and geometry of the cones also indicates that Tharsis magmas were not depleted in volatiles late in the eruptive history of the region. This example underlines the importance of searching for additional cases which provide the unambiguous information necessary for this type of reconstruction. Unlike this example, many of the dikes that characterize Tharsis volcanoes earlier in their history are much larger and are radial to the volcanoes themselves [1], and Tharsis flows are often many tens to hundreds of km in length. Further, the orientation of the dike feeding this eruptive phase (Fig. 1) is not clearly radial to any particular edifice, and its orientation and timing raise the question of magma and dike sources late in the history of the Tharsis region. Further analysis of this feature and its relation to Tharsis lava flows outside the fan-shaped deposit will assist in addressing these questions.


![Figure 1](image1.png) Rows of volcanic cones and adjacent flows superposed on the Arsia fan-shaped cold-based glacial deposit [5]. a) Southern segment (MOC image S07-01314, ~3.5 m/pix). B) Northern segment (THEMIS image V12087009, ~35 m/pix).
DICHOTOMY BOUNDARY GLACIAL MODIFICATION AT MID-LATITUDES: DEGRADATION OF THE WALLS AND CENTRAL PEAKS OF MOREUX CRATER (135 KM DIAMETER). James W. Head¹ and David R. Marchant², ¹Dept. Geol. Sci., Brown Univ., Providence RI 02912 USA (james_head@brown.edu, ²Dept. Earth Sci., Boston Univ., Boston MA 02215 USA (marchant@bu.edu).

Introduction: The dichotomy boundary represents one of the most important geological and geophysical features of Noachian Mars [1] but the scarp currently representing the boundary in many parts of Mars has been heavily modified. Identification, assessment and quantification of the degradational processes operating to modify the original boundary will provide important information on the amount of lateral migration that has taken place since boundary formation, and thus help to constrain models of the dichotomy origin. Processes thought to have been operating to modify the dichotomy boundary include tectonic, volcanic, fluvial, groundwater sapping, mass wasting, eolian, ice-assisted creep, and glaciation [e.g., 2-4]. We have been assessing the nature of modification processes along the dichotomy boundary at mid-latitudes in the Deuteronilus-Protonilus region and have found evidence for extensive regional glaciation [e.g., 5-6] in several areas of lineated valley fill in the fretted terrain. Here we assess the degradational processes operating on the 135 km diameter crater Moreux, superposed on the dichotomy boundary at ~44°E, 42°N (Fig. 1).

The Setting of Moreux Crater: Moreux is located at the edge of the plateau of the southern uplands and its southern portion disrupts the regional scarp representing the dichotomy boundary (Fig. 1). The majority of the crater is superposed on, and destroys or heavily modifies the population of mesas that characterize the dichotomy boundary in this area. On its exterior, ejecta from the crater can be identified to the south, on the plateau; particularly prominent is a crater chain extending SSW from near the rim crest. To the north, ejecta can be seen on the summits of the plateaus and in the intervening valleys in some places, but in others, lobate debris aprons and lineated valley fill clearly postdate the ejecta. Significant mesa formation does not appear to have occurred subsequent to the formation of Moreux and the dichotomy boundary scarp appears not to have migrated laterally subsequent to the Moreux impact event. Thus, Moreux is interpreted to have formed at a time subsequent to the formation of the current dichotomy scarp and the abundant related mesas, but prior to the end of the modification of the scarp and mesas by lobate debris aprons and lineated valley fill. Therefore, Moreux offers a unique template on which to study degradational processes on a landform at the scale of the dichotomy boundary (many tens of km), but different in morphology and structure. The interior of Moreux is 2-3 km below the rim crest (Fig. 2) and the central peak rises more than 2 km above the crater floor, approaching and sometimes slightly exceeding the rim crest elevation. Although relatively sharp, the rim crest has undergone significant local degradation where it has been eroded into a series of isolated massifs of various sizes. Here we focus on landform degradation of the crater rim crest and interior wall (analogous to the dichotomy boundary scarp) and the central peaks (analogous to the mesas).

Crater Rim Crest and Interior Wall: Topography on the crater rim crest is characterized by a series of rounded massifs and intervening smooth plains (e.g., Fig. 3). Between the massifs the plains change morphology, transitioning into lineated, flow-like lobes, extending 3-15 km down the crater interior walls (Fig. 3, center). Often characterizing the transition are scallop-like alcoves forming convex-outward scarps. Where developed, the alcoves mark the beginning of the lineated lobate deposits. The inter-massif lineated lobate deposits are joined by similar lobate deposits originating in alcoves farther down the crater wall, sometimes resulting in a broad digitate apron deposit extending from the crater wall down toward the crater floor (Fig. 3). Sometimes multiple lineated lobes originate on the crater rim in alcoves, converge and show evidence for the flow-like deformation of their lineations, and then descend downslope through a major break in the rim crest, with their often sinuous pathways controlled by wall topography. In some cases, the lobate deposits terminate in a distinctive moraine-like ridge (Fig. 3). Associated sinuous esker-like ridges and channels suggest that some flow took place.

Crater Central Peaks: At the broad scale, the flanks and margins of the central peaks appear lobate (Fig. 1). More detailed examination of the near summit areas at the top of the broad lobes (Fig. 4) reveals the presence of abundant arcuate alcove-like scarps that are the sources of the linear-ridged, lobate flow-like features. These features display parallel ridges oriented predominantly downslope; the features merge with one another, and merging ridges deform into tight and often crenulated folds. Some lobes remain distinctive and have arcuate concentric ridges at their termini (Fig. 4, lower middle), while others continue to converge (Fig. 4, left side), ultimately merging into the large basal lobes (Fig. 1).

Summary and Conclusions: On the basis of the characteristics and morphology of these lobate lineated flow-like features, their association with alcoves, evidence for their convergence and deformation, the presence of esker-like features and small sinuous channels, and proximity to lineated valley fill interpreted to be glacial in origin [5-6], we interpret these features to be due to the accumulation of snow and ice in alcoves, its compaction and flow, and its eventual sublimation, melting and loss. The presence of these glacial-like features in the interior of an impact crater with morphologies similar to the dichotomy scarp
and associated mesas strengthens the likelihood that glacialization was a significant process in the modification of the dichotomy boundary, a process that must have taken place under a climatic regime different than that of today [5-7].

Acknowledgments: We gratefully acknowledge financial assistance from the NASA MDAP Program (JWH) and the assistance of Amanda Nahm and James Dickson in data preparation and discussion.


Fig. 1. MOLA gradient and altimetry map.

Fig. 2. Altimetric profile of Moreux interior. MOLA gridded data.

Fig. 3. Crater rim crest and wall with massifs and lobate flows (THEMISV12805004).

Fig. 4. Central peaks; alcoves, and linear lobate flows converging toward base of peaks (THEMIS V01122003).
Introduction: Impact cratering is one of the most significant processes shaping planetary geomorphology, particularly in early planetary history [1]. Keys to the nature of the impact cratering process lie in exposures on Earth of impact craters planed off to different depths by erosion [2], on the Moon in the exquisite preservation of pristine surface textures and fresh crater morphology formed without an atmosphere [3], and on Venus in the very well-preserved morphologies representing interaction of the ejecta and the atmosphere [4]. On Mars, a wide range of distinctive lobate ejecta deposit morphologies are interpreted to be related to substrate volatiles [5] or interaction of the ejecta with the atmosphere [6]. Active eolian processes and episodic fluvial activity on Mars, as well as regional blanketing and exhumation, however, have obscured the detailed finescale morphology of the all but the most recent impact craters [7]. Thus, in using different planetary environments to study impact cratering, Mars tends to lie between the Earth and the Moon, with more gradation than the Moon, but much less planation than the Earth.

Recent Mars exploration has obtained high-resolution imaging, spectroscopy and altimetry data that permit the further analysis of impact cratering deposits in much more detail than previously possible. These new data, combined with an increased understanding of both the impact cratering process in general and deposition/exhumation processes on Mars, has led to renewed interest in the role that Mars can play in further decoding important aspects of the cratering process. In this paper we report on the discovery of a complex system of ridges on the floor of an impact crater that we interpret to be breccia dikes formed in concert with the impact cratering event and subsequently exhumed. We document the characteristics of these features, show the nature of the overlying deposits and their exhumation, assess their role in the cratering process through comparison with terrestrial breccia dikes, develop criteria for the recognition of these features and distinguishing them from magmatic dikes. In this abstract we outline the nature of breccia dikes on Earth and summarize criteria for their recognition on Mars.

Breccia Dike Characteristics in Terrestrial Craters: Breccia dikes are a common feature in eroded terrestrial complex impact craters [e.g., 8]. In a classic study, Lambert [8] proposed a classification system of breccia dikes in crystalline rocks distinguishing Type A (range up to a few cm in width, consist of small rounded commonly monomineralic fragments embedded in a cryptocrystalline matrix which often displays fluidal texture and which can be subdivided into liquid and solid particle flows) and Type B (angular to subrounded rock, mineral and glass fragments with a wide size distribution in a matrix of similar but finer grained material). Type B dikes can be subdivided into monomict and polymict subtypes and the polymict dikes show both no wall displacement (B1) and relative displacement of the fracture wall (B2). B2 dikes range from cm to m wide and form simple, straight continuous dikes. B1 dikes are very complex and variable in geometry and size. They bifurcate, anastomose, and show sharp changes in direction and thickness. On the basis of these characteristics and relationships, Lambert [1] concluded that 1) breccia dikes are not limited to the central uplift zone, but influence large areas of the crater target area; 2) A single impact can produce several successive generations of fractures and breccia diking; 3) Type A dikes form first as part of the initial shock compression; 4) Type B dikes form during and/or after pressure release in the modification phase, with B1 dikes emplaced by high-energy intrusion into the transient crater as it grows, and B2 dikes forming during the modification stage as blocks are displaced. Early stage fracture-producing processes serve to reduce target strength and angle of internal friction, enhancing movement in the modification stage.

Impacted sedimentary rocks exhumed from even deeper levels show abundant evidence of breccia dike development, particularly in the central peak region. The Upheaval Dome structure in Utah, USA, exposes a complex sandstone dike network emplaced and injected during crater formation and central peak formation (Figs. 1, 2) [9]. The dike system is characterized by extreme variations in thickness (0.1-10 m) even over short distances, decreasing mean dike width with increasing distance from the crater center, flow bulges with greatest thicknesses at nodular points, marking branch points where the dike bifurcates in two or more directions, all forming a honeycomb-like interconnected network of breccia dikes. Erosional outliers of dikes peripheral to the central peaks suggest that breccia dike networks were common throughout the crater subfloor at shallower levels.

Larger impact structures produce even wider and more extensive impact-related dikes. The 200-250 km diameter Sudbury impact structure displays a series of steeply-dipping dikes oriented radially and concentrically around the structure. Known as the "Offset Dikes" because they often terminate and then reappear laterally offset by a few km [10], these features occupy footwall faults and fractures related to the excavation and modification stages of the impact event. One such dike feature, the Hess Offset, is at least 23 km long, 10-60 m wide, and located within, and oriented sub-
concentrically to, the large Sudbury crater. The dike is granodioritic, undulates in thickness along strike, and splays locally to form claw shaped apophyses, originating during the modification stage of the impact event [11]. Similar Sudbury dikes (Whistle-Parkin Offset Dike, 12 km long, ~30 m wide [12]; Foy Offset Dike, 36 km long, 50 m wide and up to 400 m wide at the source [13]). Multistage emplacement (compression, excavation and modification) is implied for these dikes and the offsets are interpreted to be due to modification-stage crater adjustments.

Often encountered in larger impact structures are pseudotachylites and breccia belts, which are in turn cut by breccia dikes. For example, at Sudbury, the South Range Breccia Belt is a 45 km long and tens to hundreds of meters thick breccia with a very fine-grained matrix, interpreted to have been emplaced as a result of high strain-rate seismogenic slip during rim collapse of the Sudbury crater [14], and often referred to as a 'superfault'[15].

Summary: In summary, breccia dikes are seen in most of the eroded terrestrial complex craters and are developed in both crystalline and competent sedimentary target rocks [8]. They are characterized by wide variations in petrology and mineralogy (commonly related to the target material and the stage in the cratering processes), can range up to many tens of meters in thickness, and tens of kilometers in length. They occur in complex honeycomb like patterns and are often offset along late-stage crater-related faults. Individual dikes can undulate in width and branch, anastomose and bifurcate along strike. Although terrestrial weathering and vegetative cover inhibit complete mapping of breccia dikes, where detailed mapping has been done [e.g., 9] breccia dikes can be comprehensively developed (Fig. 1, 2) and drilling data on terrestrial craters further suggests that breccia dikes are very common [8]. In a separate abstract we use these basic characteristics of breccia dikes associated with impact craters on Earth as criteria to assess the breccia-dike-like features discovered on Mars [16].

BRECCIA DIKES IN IMPACT CRATERS ON MARS: EXPOSURE ON THE FLOOR OF A 85-KM DIAMETER CRATER AT THE DICHOTOMY BOUNDARY. J. W. Head¹ and J. F. Mustard¹, ¹Department of Geological Sciences, Brown University, Providence RI 02912 USA (james_head@brown.edu).

Introduction: Recent Mars exploration has obtained high-resolution imaging, spectroscopy and altimetry data that permit the further analysis of impact cratering deposits in much more detail than previously possible. These new data, combined with an increased understanding of both the impact cratering process in general and deposition/exhumation processes on Mars, has led to renewed interest in the role that Mars can play in further decoding important aspects of the cratering process. In this abstract, using criteria developed from a synthesis of breccia dike occurrences in terrestrial craters [1], we report on the discovery of a complex system of ridges on the floor of an impact crater that we interpret to be breccia dikes formed in concert with the impact cratering event and subsequently exhumed. We document the characteristics of these features, show the nature of the overlying deposits and their exhumation, assess their role in the cratering process through comparison with terrestrial breccia dikes, and develop criteria for the further recognition of these features on Mars.

Geological Setting: The dichotomy boundary, separating the northern lowlands from the southern uplands, is an area of enhanced erosion and degradation of a variety of landforms, including impact craters [2]. We have examined a row of four aligned but non-overlapping Noachian-aged impact craters, each 85-100 km in diameter, and extending along and adjacent to the dichotomy boundary over a distance of about 400 km (Fig. 1). Each impact crater shows a somewhat different state of degradation including heavily fractured floors, graben cutting the crater rims and floors, sapping channels and valleys eroding floors and rims and breaching craters, exhumed and stripped deposits on crater floors, completely breached crater rims, and extensive eolian deposition. We focus here on the analysis of complex arrays of ridge structures exhumed and exposed on the floor of the 85 km diameter crater designated C in Fig. 1.

Geological Relationships and Stratigraphy on the Floor of Crater C: The uppermost youngest materials on the floor of the crater are of eolian origin and consist of a large low-albedo patch of dunes in the northern part of the floor, and abundant relatively high-albedo linear and arcuate dunes arranged in clusters and patches elsewhere on the crater floor (Fig. 2-4). The very young age of these dune deposits is indicated by the apparent lack of superposed craters. Underlying these eolian deposits is a regionally relatively smooth flat-lying medial unit that occurs over a significant portion of the crater floor. Examination of this unit shows that it is composed of several flat-lying layers that appear to have been partially stripped, exhuming successively underlying layers in this unit. In some places, this medial unit has been completely removed, exposing the lowermost unit, which consists of a distinctive network of lattice-like ridges. The thickness of the flat-lying medial unit is of the order of hundreds of meters and individual subunits within it range from a few, to a few tens of meters thick. The uppermost part of the medial unit is slightly hummocky and moderately cratered. Exposed surfaces of subunits lower in the medial unit show a higher crater density, and craters on the lowermost subunit have been filled by the next overlying unit and subsequently exhumed. The lowermost medial subunit is transitional with the basal lattice-like unit with the ridges in the lattice appearing at the margins of the unit and becoming progressively exhumed with greater distance from the contact (Fig. 2, 3, lower corners). The medial unit weathers to large irregularly-outlined areas with stepped and terraced margins (representing the subunits), and rounded to oval mesas up to ~1 km in diameter (Fig. 2, 3). In summary, the nature of the medial unit suggests that it is flat lying crater floor fill, that it represents a series of filling and exhumation events, and that the materials making it up is easily eroded and removed, most likely being the source of the upper unit dune deposits. On the basis of geological relations elsewhere on Mars, deposits characterizing crater fill might include initial impact crater ejecta fallback and melt, subsequent distant ejecta, mass wasting, atmospheric dust and ice deposits, fluvial deposits, and volcanic tephra and lava flows.

Geological Characteristics of the Lowermost Lattice-Like Unit: The lowermost unit contains a distinctive set of linear ridges of broadly similar width forming a lattice-like pattern. Ridge exposures ranged from ~1-4 km in length and ~50-100 m in width; if the ridge represented a dike, this width certainly overestimate the actual dike width due to the presence of a flanking erosional debris apron. Individual ridges are generally straight to slightly curving, (Fig. 2-4) but some are slightly sinuous (Fig. 4, middle). In some cases they form broadly parallel ridges separated by hundreds of m to a km (Fig. 2, 4); in others, these ridges are cross-cut by near-orthogonal ridges forming a box or lattice-like pattern (Fig. 2, center; Fig. 4, right). Some ridges terminate abruptly against other orthogonally-oriented ridges (Fig. 3, upper right). Others bifurcate (Fig. 2, middle; 3, upper middle), and a few show possible en echelon structure (Fig. 3, right). Cross-cutting relationships are often clearly observed, but as yet no clear relationships have been established for sequential developments of trends. Near-orthogonal terminations suggest faulting and lateral offset, but exposure is insufficient to
locate and restore possible offsets. The location of these images (Fig. 2-4) is from the crater floor, extending from the base of the wall to the edge of the central peak region. The general orientation patterns are crudely radial and concentric to the crater, and more detailed analysis of orientations is underway.

**Comparison to Terrestrial Breccia Dikes and Interpretation:** On Earth, breccia dikes are common in eroded complex craters in a wide range of target rocks [3-5], range up to tens of m in width and tens of km in length, occur in complex honeycomb-like patterns and are often offset along late-stage crater-related faults; individual dikes can undulate in width and branch and bifurcate along strike. These characteristics are very similar to those of the ridges prominently occurring in the exhumed basal unit on the floor of the 85 km diameter crater C. On the basis of these strong similarities, we interpret these ridges to be the erosional manifestation of breccia dikes formed below the floor of the crater during the impact event that created crater C, and subsequently exhumed.

The further documentation of the presence and nature of breccia dikes on Mars will help in documenting the nature of impact cratering processes there and aid in assessment of the levels and depths of exhumation processes. The presence of networks of breccia dikes below complex crater floor floors also has important implications for the structure of the crust of Mars. For example, models of Mars crustal hydrology need to take into account the development of significant lattice-like networks of solid dikes in considering the permeability of the impact-formed megaregolith. On Earth, breccia dikes are often associated with key mineral resources (e.g., Sudbury) and are also clearly related to late-stage thermal and mechanical readjustments of impact craters, thus forming candidate longer-term distributed heat sources analogous to magmatic hydrothermal vents. Upcoming very high spatial and spectral resolution instruments on MRO have the capability to analyze the geology and mineralogy of these type of breccia dikes in detail.


![Fig. 1. (a) Four impact craters with different states of degradation at the dichotomy boundary; A-D, left to right. (b) Crater C.](image1)

![Fig. 2. The floor of crater C; THEMIS Image V12779010.](image2)

![Fig. 3. The floor of crater C; MOC Image R18/00213.](image3)

![Fig. 4. The floor of crater C (central part of Fig. 2); MOC Image E14/01582.](image4)
Introduction: The presence of several giant impact basins on Mars (e.g. Hellas, Utopia, Argyre, Isidis), open a scientific window into early years of Mars geology. For example, the count of these basins has been used for the one of the first chronology system for Mars [1]. Recent data from gravity and magnetic fields give the basis for important speculations about early thermal and magnetization state of Martian crust and core (e.g. reviews [2-4], and papers [5-7]) when all giant basins have been formed. The impact basin formation is still poorly known process mainly modeled from geophysical data (e.g. [8]). Despite the presence of numerical modeling of impacts at all scales [9], the specific modeling of giant crater formation at a spherical planet with the specific thermal profile is still the unresolved problem due to many lacunas in our knowledge of material strength and thermodynamic properties of crustal and mantle rocks as well as the material of the Martian core. The presented work is devoted to the reconnaissance study of giant basin formation on Mars.

The numerical model consists of the material motion equations solver (SALEB hydrocode is used here [10, 11]), a set of equations of state (ANEOS code is used here [12, 13]), and a set of assumptions about the thermal state of the target.

Mars model. Spherical Mars is modeled at the rectangular grid of cells filled with 3 materials: basalt models crust, dunite models mantle, and iron models core. All 3 materials are described with ANEOS equation of state. The problem of fitting of (mainly) shock-wave derived equation of state to the real Martian rocks should be refined in the future. The same is valid for the unknown Mars core material, modeled here preliminarily with the available EOS for pure iron. The thermal profile of Mars has been estimated for various geological periods by many authors as reviewed in [2-3]. Here we use crust/mantle and core/mantle temperatures close to estimates in [14]. The general view of the target is shown in Fig. 1. The model simulates the planet with radius 3400 km, mass 6.44 $10^{23}$ kg, and surface gravity of 3.69 m s$^{-2}$.

Model runs have been done for vertical impacts of (basaltic) asteroids (diameters from 400 to 800 km) with velocities of 8 to 12 km s$^{-1}$. The typical outcome is shown in Figs 2 and 3. The final shape of the planet returns close to the sphere, however the mantle uplift and remote stresses are visible in the small-scale analysis. The main (so far) result we see in the evidence that the “melt pool” at the basin center is the inevitable consequence of basin-forming impacts. It means that the crust/mantle boundary estimated from geophysics is the “new crust/new mantle” boundary as the solidification of the «melt pools» repeats the primary crust separation process with possible geochemical peculiarities (such as “depleted mantle”). This is a valuable input model for possible further geochemical speculation about mineralogy of “new” crust and mantle inside basins.

Discussion. The model results give an opportunity to estimate the size of basin-forming impactors by the direct comparison of mantle uplift profiles, modeled numerically and geophysically. However it needs the solution of another model problem of solidification of the “melt pool” with stress field in the crust and mantle. The most similar problem is analyzed in [15, 16] with simplified initial conditions (hemispheric heated volume) but with the whole planet mantle convection. It is shown that magmatic activity in the impact site may exists for 100 Ma and longer depending on mantle viscosity. This activity may be enough intensive to produce the Tharsis rise.

For smaller impacts the solidification of “new” crust and “new” mantle is inevitable. It means that the interpretation of MGS gravity anomaly data for Mars (as well as for lunar basins) should take into account possible geochemical “novelty” of crust and mantle in originally molten zone, which may occupy areas of n*1000 km across (like in Fig. 3).

Conclusion and outlook. Analizing the data obtained with MEX, MGS and Odyssey spacecrafts, the modeling of impact cratering on Mars allows us to understand better Martian geology and geophysics.
GIANT MARTIAN BASINS: B. A. Ivanov

Fig. 1. (a) Modeled Mars with core (red), mantle (brown) and crust (red). (b) density profile for modeled with ANEOS basalt/dunite/iron Mars (solid line) compared with one of geophysical models [2].

Fig. 2. Selected time frames for modeling of 500-km asteroid (basalt) impact with $v_{imp}=8$ km s$^{-1}$. Top to bottom: 0, 450, 2300, and 10000 seconds after impact. Note the giant “splash” due to central dome collapse in (c) delivered a lot of melt a top of the crust around crater.

Fig. 3 The thermal state for 10,000 s (~2.7 hours) after the impact shown in Fig.2. While the current equation of state does not reproduce exactly solidus and liquidus for mantle, we estimate the melted state as partial melt at a given solidus (raised with pressure as for KTB peridotite), and complete melt as material overheated 200K above solidus for a local pressure. The melted zone is in an eddy motion in the computations. The thick layer and remote patches of ejected melt are visible up to 2000 km from the axis of symmetry. The fate of this (invisible now) mantle melt is unclear: (1) it may be heavily mixed with local crust material during the ballistic deposition, or (2) it may sink down through the heated crust (which deserve the further model analysis). Isotherms 1600 to 2000 K (black lines with numbers) depict the general geometry of the impact “hot spot”. The “melt pool” has a diameter of ~600 km with a depth of 200 to 600 km.

THE NATURAL THERMOLUMINESCENCE AND ORBITS OF METEORITES.

A.I. Ivliev, V.A. Alexeev. Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow 119991 Russia; e-mail: cosmo@geokhi.ru.

Introduction: The only way to obtain an accurate orbit is to simultaneously photograph the trajectory of the fireball from two or more widely separated locations. Several photographic networks [1] have been established to do just this. After almost 40 network-years of operation, however, these efforts have led to the recovery of only four meteorites: Pribram [2], Lost City [3], Innisfree [1] and Peekskill [4]. Orbital parameters have been estimated for another 40 meteorites from eyewitness accounts of their fall [5,6] but these are of low accuracy. Limited orbital information can also be obtained from the local time of fall [7]. Meteorites with perihelion q ~1 astronomical unit (AU) should usually strike the trailing side of the Earth, or the local PM, while meteorites with perihelia q<1 AU should be more evenly spread over the leading and trailing sides. The large abundance of ordinary chondrites falling in the local PM relative to AM is interpreted as indicating that most of these meteorites had perihelia of q~1 AU.

Periodic changes of the perihelion of chondrites during their cosmic-ray exposure history (up to ~n10^7 years) may stipulate diffusion losses of gases at q ≤0.2 AU (temperatures T≥400 °C) and reaccumulation of natural thermoluminescence at q ~1 AU during the last of ~10^5 years before capture by the Earth.

Thermoluminescence (TL) is an extremely useful technique for studying the metamorphism and recent histories of meteorite [8]. The induced TL of meteorites is predominantly controlled by the type and abundance of feldspar present, while the level of natural TL (TL of the sample "as received") is determined by the thermal and radiation history of the sample. In this study we use the natural TL of individual meteorites to estimate their closest approach to the Sun. Natural TL is energy stored in crystals of certain minerals by ionizing radiation, such as high-energy galactic cosmic rays [9,10]. This energy can be released in the form of visible light by heating. The experimental procedures have been described in more detail earlier [11-13].

Perihelion of the orbits: During occurrence in orbit, natural TL is accumulated in meteorite owing mainly to cosmic ray radiation. The natural TL level of a meteorite is initially very low because of shielding from cosmic radiation by its parent body. After the impact, which actually produces the meteorite-sized body (meter-diameter or less), natural TL levels build up relatively rapidly, reaching a state of "equilibrium" within about 10^5 years [8]. The level of natural TL, at least in the lower temperature and less thermally stable portion of the glow curve, is then in a state of dynamic equilibrium, varying slightly during each orbit as the TL drains slightly at the higher temperatures of perihelion which it regains at aphelion. The level of TL in the higher temperature and more thermally stable part of the glow curve is, however, relatively unaffected by orbital temperature variations and may reach saturation level at exposition by the cosmic rays for a sufficiently long period of time. The average level of low-temperature equilibrium TL can vary over the time span of 10^6-10^7 years if the meteoroid experiences changes in its long-term orbital parameter. After fall on the Earth, the natural TL levels gradually decrease to lower values as a result of the higher terrestrial temperatures and lower dose accumulation rates [14-16]. TL levels are also lowered by heating during atmospheric entry, but this only affects a thin (< 5 mm in thickness) outer rim of the meteorite.

The TL levels are within 20-80 krad (at T~250°C on the glow curve) in the most ordinary chondrites with known fall dates [17-19]. Calculation of the value of the equivalent dose of natural TL in ordinary chondrites allows us to suggest that the intensity of TL is a sensitive indicator of their degree of heating by Sun at passing the perihelion. In fact, at the lower the perihelion, we will have the higher the heating and the lower the equilibrium TL. Chondrites having orbits with the perihelion q<0.85 AU must show very low levels of natural TL (<5 krad at T~250°C on the glow curve), whereas those with q>0.85 AU must show wide ranges of natural TL values (>5 krad) with a considerable scatter related to the variations in the rate of dose accumulation (at a varying degree of shielding and albedo) [17]. However, comparison of the thermal and radiation histories of meteorites solely on the basis of natural TL is hampered by considerable variations in the sensitivity of TL accumulation in different meteorites. Thus, it appears reasonable to normalize the intensity of natural TL in each sample to its sensitivity through the measurement of the TL value per unit dose induced by a radioactive source. The ratio known as equivalent dose (ED) is determined for each temperature value of the glow curve using the formula:

\[ ED = D \times \left( \frac{TL_{nat}}{TL_{ind}} \right) \]

where \( TL_{nat} \) and \( TL_{ind} \) are the natural and induced TL, respectively, and D is the dose of laboratory radiation (rad). Using such an approach, Melcher [20] estimated the perihelia of 45 meteorites. However, investigations suggest that it is more reasonable to calculate ED for two temperature intervals on the glow curves; ED_TLT at T=100-240°C and ED_THT at T=240-340°C. This allows us to reduce...
the error of ED estimate to \( \leq 15\% \) and estimate more accurately the perihelion value.

Comparative measurements of natural TL and TL\(_\gamma\), induced by \( \gamma \)-radiation, and calculations of ED\(_{LT}\) and ED\(_{HT}\) using a special program were carried out for 21 chondrite samples (See Table). Some of these chondrites were studied in [20], including the Pribram chondrite with a known orbit. The ED values of Pribram correspond to its perihelion (\( q=0.8 \) AU) and coincide with the results of ED measurements reported in [20]. For the majority of chondrites, including Bjurbole L/LL4, Chainpur LL3.4, Dalgety Downs L4, Dhajala H3.8, Gorlovka H3.7, Grady H3.7, Elenovka L5, Khohar L3.6, Kunashak L6, Kunya-Urgench H5, Kyushu L6, Mezo Madaras L3.7, Nikol'skoe L4/5, Ochansk H4, Pervomaisy L6, Pultusk H5, Rakity L3.6, and Saratov L4, the perihelia of orbits (\( q \)) are within \( 1.0-0.8 \) AU. Lower perihelia were determined only for the L5 chondrites Malakal (\( q\approx0.5-0.6 \) AU), which is consistent with [20], and Dimmit H3.7 (\( q\approx0.6-0.8 \) AU). A value of \( q\approx1 \) AU was obtained for the Kunya-Urgench orbit, which agrees with the perihelion estimate from the radiant of the chondrite fall [21].

### Table. Equivalent doses (rad) for meteorites

<table>
<thead>
<tr>
<th>Meteorite</th>
<th>Type</th>
<th>ED(_{LT})</th>
<th>ED(_{HT})</th>
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<tbody>
<tr>
<td>Biurbole</td>
<td>L/LL4</td>
<td>38 ± 5</td>
<td>156 ± 4</td>
</tr>
<tr>
<td>Chainpur</td>
<td>L.L3.4</td>
<td>5.7 ± 1.2</td>
<td>77 ± 21</td>
</tr>
<tr>
<td>Dalgety Downs</td>
<td>L4</td>
<td>0.23 ± 0.02</td>
<td>30.1 ± 5.6</td>
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<tr>
<td>Dhajala</td>
<td>H3.8</td>
<td>1.7 ± 0.1</td>
<td>21 ± 1</td>
</tr>
<tr>
<td>Dimmit</td>
<td>H3.7</td>
<td>0.6 ± 0.3</td>
<td>10.6 ± 6.1</td>
</tr>
<tr>
<td>Elenovka</td>
<td>L5</td>
<td>3.9 ± 0.3</td>
<td>23 ± 2</td>
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<tr>
<td>Gorlovka</td>
<td>H3.7</td>
<td>46 ± 15</td>
<td>288 ± 44</td>
</tr>
<tr>
<td>Grady</td>
<td>H3.7</td>
<td>1.4 ± 0.4</td>
<td>80 ± 11</td>
</tr>
<tr>
<td>Khozar</td>
<td>L3.6</td>
<td>1.21 ± 0.16</td>
<td>51.3 ± 1.4</td>
</tr>
<tr>
<td>Kunashak</td>
<td>L6</td>
<td>16.6 ± 2.0</td>
<td>43.2 ± 7.0</td>
</tr>
<tr>
<td>Kunya-Urgench</td>
<td>L5</td>
<td>11.2 ± 1.0</td>
<td>100 ± 8</td>
</tr>
<tr>
<td>Kyushu</td>
<td>L6</td>
<td>30.0 ± 5.6</td>
<td>181 ± 11</td>
</tr>
<tr>
<td>Malakal</td>
<td>L6</td>
<td>1.2 ± 0.1</td>
<td>2.4 ± 1.0</td>
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<td>Mezo Madaras</td>
<td>L3.7</td>
<td>35.5 ± 4.6</td>
<td>368 ± 59</td>
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<tr>
<td>Nikol'skoe</td>
<td>L4/5</td>
<td>16.1 ± 0.7</td>
<td>159 ± 15</td>
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<tr>
<td>Ochansk</td>
<td>H4</td>
<td>10.7 ± 1.1</td>
<td>105 ± 10</td>
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<tr>
<td>Pervomaisy</td>
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<td>20.1 ± 1.9</td>
<td>71 ± 12</td>
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<tr>
<td>Pribram</td>
<td>H5</td>
<td>11.4 ± 1.5</td>
<td>127 ± 12</td>
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<td>Pultusk</td>
<td>H5</td>
<td>4.6 ± 0.9</td>
<td>81 ± 15</td>
</tr>
<tr>
<td>Rakity</td>
<td>L3.6</td>
<td>12.9 ± 3.0</td>
<td>309 ± 25</td>
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<td>Saratov</td>
<td>L4</td>
<td>17.1 ± 1.1</td>
<td>56 ± 5</td>
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</table>

**References:**

**Chemical Modification in the Lunar Olivine Microcrystals Under the Solar Cosmic Ray.**

L.L. Kashkarov, S.N. Shilobreeva, G.V. Kalinina, V.I. Vernadsky Institute of Geochemistry and Analytical Chemistry, RAS, Moscow, Russia (ugeochem@geochem.home.chg.ru)

**Introduction.** Previous investigations [1] of the lunar regolith silicate crystals demonstrate the possibility of measuring the radiation effects from the solar cosmic ray (SCR) nuclei of the iron VH-group in the individual silicate microcrystals. Due to charge, mass and energy of the cosmic ray nuclei in the exposed silicates different radiation effects can be influenced [2]. One of these is the chemical and phase composition modification of silicates. Modelling of these processes with the help of the accelerated low-energy ions H, D, He and Ar [3,4] indicate that chemical composition of the interplanetary matter could be essentially changed. For these reasons experimental investigations of the volume distribution inside an individual lunar regolith silicate microcrystals subjected to different exposure SCR protons and α-particle dose are important. This paper reports: (1) The new results on investigation of the radiation parameters in the individual silicate micrograins of the lunar regolith matter, determined by using fossil track method [1] and (2) Preliminary results of the chemical modification observed in the some searched crystals.

**Samples and method.** Olivine crystals (about 80 grains) from the Luna-24 soil sample No 24184 of 0.127±0.200 mm size fraction were taken for the investigation. The crystals were mounted in epoxy-resin tablet, polished and, on the freshly revealed internal cut-off surfaces the chemical composition profiles were measured with the help of Electron Microprobe technique. Then, on the same crystal surfaces chemically etched VH-nuclei tracks were observed [1]. Estimation of the total proton and α-particle exposure dose was done on the base of experimentally established VH-nuclei track production rate (dp/dt)α vs depth (X) of scoop near lunar regolith surface. For the shielding surface layer from 2×10^3 cm up to 10^7 cm (dp/dt)α = 1.2×10^{-7} track/cm² yr. [5]. Thus, the olivine grains that do not demonstrate a well-defined track-density gradient and have uniform track distribution of α = 10^6 ÷ 10^7 track/cm² have been exposed near the submillimeter regolith depth for the time T = 1±10 m.yr. Corresponding SCR proton and α-particle integral fluxes estimated in that way were obtained. Calculation of the implant range profile of protons and α-particles [6] indicate that at E ≥ 10 MeV/ion induced degree of crystal lattice damage varied along the olivine grain depth within the limits of 5 ± 50 %.

**Results and Discussion.** Track densities and accounted protons and α-particles dose values detected in the eight individual olivine grains, are presented in Table. Practically all observed tracks are due to VH nuclei (iron group 23≤Z≤28) of the solar cosmic rays. Note that track density of the spontaneous and induced fission of Th and U is negligible small (≤0.1%) and is not taken into account.

<table>
<thead>
<tr>
<th>n₀</th>
<th>Sample</th>
<th>Track Density, ρ cm⁻²</th>
<th>Exposure Age, m.y.</th>
<th>Dose, cm⁻² (10^17)</th>
<th>Protons</th>
<th>α-particles</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>I – 1</td>
<td>(1.5 ± 0.1)×10⁴</td>
<td>2.5</td>
<td>7.9</td>
<td>0.55</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>I – 2</td>
<td>(3.2 ± 0.4)×10⁵</td>
<td>0.06</td>
<td>0.2</td>
<td>0.019</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>I – 5</td>
<td>(5.0 ± 1.0)×10⁶</td>
<td>0.008</td>
<td>0.025</td>
<td>0.00175</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>I – 7</td>
<td>(1.5 ± 0.1)×10⁷</td>
<td>2.5</td>
<td>7.9</td>
<td>0.55</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>I – 8</td>
<td>(2.4 ± 0.3)×10⁸</td>
<td>0.04</td>
<td>0.13</td>
<td>0.09</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>II – 2</td>
<td>(1.2 ± 0.1)×10⁶</td>
<td>0.5</td>
<td>1.6</td>
<td>0.11</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>II – 8</td>
<td>(1.3 ± 0.1)×10⁷</td>
<td>2.2</td>
<td>6.9</td>
<td>0.48</td>
<td></td>
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<tr>
<td>8</td>
<td>III - 8</td>
<td>(1.4 ± 0.1)×10⁸</td>
<td>0.23</td>
<td>0.7</td>
<td>0.049</td>
<td></td>
</tr>
</tbody>
</table>

(*) Track density (ρ) values for VH-nuclei of the SCR, E_VH=10 (100) MeV/nucleon.

(**) Estimation by relation p_{0.1 cm} = 6.7×10^{10} track/cm²-m.y. (see text).

(****) Accounted on the integral flux of the SCR protons: I_{10 MeV} = 10^4 protons/cm²-s. Integral flux of the SCR α-particles I_{α10 MeV} = 0.07×I_{α10 MeV} particles/cm²-s
Some examples of the visible chemical changes, determined along profiles in the analysed olivine crystals are shown in Figure. The degree to which radiation-induced element redistribution can be expected inside of the olivine crystals on the ~ 100 µm scale, however, remains to be established in the future modelling experiments.

**Conclusions.** (1) On the base of the first-step results obtained we can state qualitatively, that in some lunar regolith olivine grains we can observe changes in Mg, Fe, Ca and Si concentration weakly-sloping from edge to edge of single crystals. This can be due to comparatively high (up to ~10^{18} protons/cm^{2}) dose values of the SCR irradiation. (2) Track vs chemical composition investigation in the eight olivine samples from the Luna-24 column soil gives the preliminary results about the possibility of the SCR influence on the chemical modification in the single grains. (3) Estimated age of the near surface (up to 0.1 cm depth) solar cosmic ray irradiation and the accountable proton and α-particle energy spectrum are the main quantitative parameters for the accurate calculation of the total proton and α-particle exposure dose. (4) Small portion of the relatively high-irradiated olivine crystal grains (ρ ≥ 10^7 track/cm^2) demonstrate the correspondingly low degree of cosmic ray influence upon the Luna-24 searched sample.

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**Fig.** Concentration profiles of MgO and SiO₂ across single olivine grains from the Luna-24 samples

Summary: We perform synoptic Martian wind measurements using global Hubble Space Telescope (HST) images in the season of perihelion. By tracking cloud movement we confirm retrograde zonal winds, poleward deflection for high southern latitudes and local deviation of wind due to topographic effects. Wind speed and direction are in general agreement with Mars GCM.

HST global Mars imaging: For measurements of wind speed and direction we use data obtained during HST observation program #9738 [1, 4]. This program was scheduled for the time of closest Earth-Mars encounter of year 2003, allowing the highest spatial resolution ever achieved from Earth. Five series of images of Mars were taken with the High-Resolution Channel of the Advanced Camera for Surveys [2] just before and after perihelion (see Table). The spatial resolution for the images is about 7 km/pix at the disk center. The observation timing provided imaging of the same hemisphere of Mars at all dates (disk center at 19°S, 20-35°W). The season was summer in the southern hemisphere (see the areocentric longitude of the Sun $L_S$ in the table). In each spectral filter, a series of 3 images with different polarization filters was taken. The maximal time lag between consecutive images in each filter was ~3.8 min. Polarization effects are weak and were studied in detail in [4].

<table>
<thead>
<tr>
<th>Date of Year 2003</th>
<th>$L_S$, deg</th>
<th>Image scale, km/pix</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aug., 24</td>
<td>247.0</td>
<td>6.70</td>
</tr>
<tr>
<td>Sept., 05</td>
<td>254.5</td>
<td>6.79</td>
</tr>
<tr>
<td>Sept., 07</td>
<td>255.8</td>
<td>6.84</td>
</tr>
<tr>
<td>Sept., 12</td>
<td>259.0</td>
<td>7.01</td>
</tr>
<tr>
<td>Sept., 15</td>
<td>261.0</td>
<td>7.15</td>
</tr>
</tbody>
</table>

Data processing: The standard dark current, flat field, and geometric distortion calibrations are performed routinely by the HST data retrieval facility. Cosmic-ray-track removal (most abundant in the near-UV) is performed with an original heuristic procedure [4]. Visual inspection shows an extensive and complicated system of clouds and hazes over the surface albedo features (Valles Marineris, Terra Meridiana, Thaumasia Planum) and rather dark surface details visible through clear atmosphere. The cloud system is best seen in the wide-band UV filter (330 nm effective wavelength). Images in this filter are used in this study.

We accurately transform the images into a common projection knowing the exact orientation of the planet. Ratios of successive images reveal shifts of the cloud features due to their movement relative to the surface features during a few-minutes-long interval between exposures. We measure these shifts and infer speed and direction of cloud movement, which we consider as a proxy for wind speed and direction at their altitude.

We begin by slightly smoothing albedo images with a Gaussian filter to suppress high-spatial-frequency noise. We then perform quantitative measurements of cloud shifts by maximizing the local covariation of images in each pair. About 650 shifts in total are obtained for all five dates. The entire set of our measurements covers the area 60°S-20°N, 100°W-20°E. Fig. 1 shows an example of the results. Deflection of the wind direction pattern from easterly to northerly near the eastern edge of Tharsis rise is seen. Individual measurements for all dates for the zonal ($U$) wind component range from -80 to 30 m/s with an average of -30 m/s; the meridional ($V$) component varies from -60 to 30 m/s with the average of -15 m/s.

Fig. 1. Measurements of wind speed and direction for Sept. 7 projected and superposed over topographic map.

Among the sources of error is the resolution in the local covariation procedure (+0.1 pix, while the absolute values of inferred shifts are within 2 pixels), inaccuracy of scale knowledge (at 10^-3 level), inaccuracy in the estimation of Mars disk center, and individual field of view distortion of the polarization filters (+0.3 pix). Compiling of these error sources results in a ±10 m/s accuracy of speed and ±7° in direction.

To study the global distribution of martian wind over the studied region we average individual measurements within the 10° latitude zones (Fig. 2). The zonal ($U$) component confirms domination of westward winds (negative values) with maximal ampli-
tude at the equator, decreasing at northern and southern latitudes down to near-zero values at 50ºS. The $V$ component shows a prominent latitudinal trend from negative (southward) values to zero and then positive values for low northern latitudes.

**Comparison with global circulation models.**

We compare our wind measurements with the European Mars Climate Database (EMCD) v 4.0 available at http://www.lmd.jussieu.fr/mars.html. The EMCD data are calculated with a global circulation model (GCM) [3] with parameter choice providing the best fit to the observed surface and atmospheric temperatures. We compare our results with the "MGS dust scenario" at different altitudes for the time of local noon in the "MGS dust scenario" conditions. **Fig. 3** compares our individual measurements of $U$ and $V$ with latitudinal profiles from EMCD for four altitudes from 20 to 40 km. The closest agreement of $U$ observations with the database is observed for 40 km altitude for equator and low northern latitude and for 30 km altitude for low southern latitudes. For higher southern latitudes, the observed zonal speed is shifted to positive values in comparison to the model. $V$ estimates also show the best fit with 30 km model profile for equator and low northern latitudes and 40 km profile for low southern latitudes. For higher southern latitudes the observational point scattering is high.

Our inferred 30-40 km cloud altitude is in agreement with [5], where 35-40 km altitude water vapor condensation level was mentioned for the season of observations.

**Conclusions.** We found general agreement between movement of cloud features observed with HST for mid-day times during the perihelion season and GCM-inferred winds at 30-40 km altitude. The discrepancies, especially for the zonal wind component in the southern hemisphere, are a subject of future analysis.

Our results show that "fast" (at time intervals of minutes) synoptic cloud tracking is a powerful source of new data on wind speed useful for further understanding Martian climate system. These data are complimentary to orbital measurements. Such observations can be carried out with HST or any suitable out-of-the-atmosphere astronomical facility with proper capabilities (high resolution in the near UV).

**References:**

1. The Earth. The decoding cosmic photographs of geological structures and the catastrophe theory. Seismic wave dynamic and strain field picture in case of inhomogeneous crust are described by the theory of catastrophes. Isoseismic curves and seismic caustic curves are connected with elementary catastrophe of wave fronts. In seismically turbid environment occurrence of caustic curves (or caustic) is inevitable. Concentration of seismic wave energy is come about in forms caustic. Energy can surpass background many times over and to form on a day time caustic surface, as well as in optical systems, original structure. The time day surface of the formed caustic structure will have deformed geological medium. Loss of stability and destruction (transformation) of the geological structure is one of objects of research of the theory of catastrophes. If geological structure with areas of instability influences of an intensive seismic wave train forms in its volume one of figures of catastrophes. So the new defect (foil) reflecting the form seismic caustic appears. According to properties of the seismic environment and waves these caustics can be formed and are classified. Considering a terrestrial surface as the projection of the three-dimensional structures which are cut off by destructed processes, allows reveal caustic pictures for case at various the original of cartographical and photographic materials (cosmic photo). Tested work was carried out on geological and geomorphologic areas of Pamirs, Tan`-Shan`, Pamirs-Altai and other territories. Thus the following was found out. All images are easily decoded not only on space pictures of M 1: 2.5 000.000 and 1: 5 000.000, but also under initial tectonic both topographical circuits and maps. The sizes of figures of elementary catastrophes on district are from first tens up to hundreds kilometers. Decoding of various forms is variably and depends on the geological structure of region. So we know after examination of the cosmic region picture new information about inhomogeneous geological medium and deformation fields.

2. The planets of terrestrial group. If planet is tectonic active the previous case has a place. The Moon is typical tectonic non-active object and the cosmic region pictures have few peculiarities.

3. The Moon. On first look the Moon crust has in general only craters surface because tectonic activity is absent. Therefore only crater appears after new meteoroid impact. These imagine is very simple. Really part of Moon surface is covered by thin layer of regolith and a dust. Seismic waves in the moment of meteoroid impact and after it which emerge from lithosphere on surface form the caustic figures. Seismic wave field from moonquake is existing long time and forms figures likeness acoustic Khladney pictures (fig.1, arrow A) on day surface. Unlike acoustic case these pictures will be instable, because next impact will destroy it.
At Vernadsky-Brown symposium and other planetological meetings we have given many examples of round features on surfaces of celestial bodies (not only solid but gaseous as well) that are not of an impact origin but appears due to crossing standing waves warping planetary spheres [1 & others]. It was expected that such planetary waves caused by movements of celestial bodies in non-round keplerian orbits with periodically changing accelerations will be seen on numerous bodies (and rings) of the complex saturnian system [2]. These expectations and predictions were well paid. Virtually, all its satellites and rings are affected by crossing wave warping producing regular nets of round features (craters). Some craters have very effective square or hexagonal shapes (by the way, this shape was marked out on the Eros’ surface but not adequately explained). Already the first images of Phoebe a year ago shown that shapes and structures due to waves are there [1]. The 4 theorems of the wave planetary tectonics were confirmed: 1. Celestial bodies are dichotomic; 2. --- are sectoral; 3. --- are granular; 4. Angular momenta of different level blocks tend to be equal. Darkish lowlands (floors) presumably filled with denser material than icy surroundings were observed in many cases but the most outstanding one is the double-face Iapetus. Its dark half (more precisely 1/3) is compared with basalt filled basins of Pacific and Vastitas Borealis [3] - different materials but the same law.

The wave shaping naturally is sharper in smaller bodies. Hyperion, Pandora, Yanus, Telesto attest this all showing a bean shape (2πR-structure) and longing acquire together with this diamond shape (πR-structure) better seen in Phoebe, Pandora, Epimetheus, Helene. Waves, very spectacular in rings, are prominent on satellites surfaces (more or less clear depending on a viewing situation – the phase angle). Regular round (or square and hexagon!) craters are very typical for all of them (Fig.11-14). Craters (granula) sizes or groove-ridge spacings are an another characteristics showing the wave action and the dependence on orbits. Higher orbital frequency – smaller crater size and v.v.

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Radius, km</th>
<th>Orb. period</th>
<th>Main granula size, km</th>
<th>Side granula size, km</th>
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<tr>
<td>Pandora</td>
<td>55x43x33</td>
<td>0.6285</td>
<td>0.07-0.04</td>
<td>1295-777</td>
</tr>
<tr>
<td>Epimetheus</td>
<td>70x58x50</td>
<td>0.6943</td>
<td>0.10-0.07</td>
<td>1648-1178</td>
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<tr>
<td>Yanus</td>
<td>110x95x80</td>
<td>0.6947</td>
<td>0.16-0.12</td>
<td>2590-1884</td>
</tr>
<tr>
<td>Mimas</td>
<td>197</td>
<td>0.9420</td>
<td>0.4</td>
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<tr>
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<td>-550.45</td>
<td>136-124</td>
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</table>

1. 2. 3. 4. 5. 6. 7. 8. 9. 10.
Fig. 1-3, 7-8. Dichotomy (wave1, $2\pi R$-structure, bean-shape): 1-Hyperion, image NASA/JPL/Space Sci. Inst.PIA06608; 2-Hyperion, 06645; 3-Yanus, 06613; 7-Telesto, 07546; 8-Pandora, 07523. Fig. 4-6, 9-10. Polygonal sectoral octahedron shaping (wave2, $\pi R$-structure, diamond-shape): 4-Pandora, 07530; 5-Prometheus, 07549; 6-Hyperion, 06608; 9-Helene, 07547; 10-Epimetheus, 07531. 11-Hyperion, 06244 (intersecting wave traces of 3 directions corresponding to 3 symmetry planes of structural octahedron which include 12 octahedron edges; note structural and size control of craters having wave origin. Four more or less blackish and whitish sectors converge in the center of the image); 12-Mimas, 06256, intersecting wave traces producing square craters, chains of craters over the whole surface; 13-Enceladus, 06252, wave woven chess-board terrain, aligned “boulders”~100 m across being tectonic granulas predicted in the table above; 14-Enceladus, a portion of image 06248, intersecting wrinkles-waves of 3-4 directions producing crater chains; 15-Graphical model of intersecting quantum-mechanical ($\pm$ alternation of + and -) waves producing lines and grids of even-sized rounded (polygonal) craters; this picture imitates the real wave structurization seen on surfaces of all saturnian satellites and rings and characteristic for all celestial bodies [1-3 & earlier publications].
A specific martian feature – lobate ejecta blankets around craters extending up to 1.5 times of a crater diameter (much wider than continuous ejecta of lunar – 0.6-0.7 diameter, and mercurian – 0.4-0.5 diameter, craters) is usually attributed for the frozen water in the martian crust. This water could fluidize under impact excavated material and make it flow like liquid. The geomorphological evidences show that the ejected material really behaves like liquid (resembles the heavy clay liquid for drilling), flows and skirts obstacles but usually does not show any significant settling. One would expect this settling for water rich material. Could it have, along with some water, compositional peculiarities? It seems that after Gamma-spectrometry of “Odyssey” and two landers – “Spirit” and “Opportunity” analyses one can say that a “peculiar rock’ is marked out. “Odyssey” show rather low silica over highlands (20-21% [1]) and, that is especially interesting, over high standing Tharsis bulge (18-20%) [1]. Low Fe signifies that this lowering in silica is not due to the basic rocks which were postulated in “entirely basaltic Mars’ model (the martian meteorites as a proof). MGS gravity data [2] have clearly shown that the martian southern highlands are composed of “light” (not dense) lithologies, much less dense than the northern lowland Fe-basalts, otherwise relatively flat even gravity signals over two hemispheres were not possible [2]. The previous global albedo data also have shown that the southern highlands are much lighter (average albedo 0.25) than the dark northern lowlands (0.15)[3] hinting at different rocks. Very long lobate formations around huge Tharsis volcanoes also require very low viscosity lavas. Density of martian soils on surface and to a depth of about 10 m according to various geophysical methods (radar, polarimetry, IR, “Viking” data) is lower than that of the Moon and is lower for light areas than for dark ones [3].

Now this “whitish” low density material is partly characterized on Meridiani Planum by “Opportunity” [4] – it is salt: sulfates, chlorides, bromides covering and penetrating layered sedimentary (mainly eolian?) rocks. The salts are discovered in craters and, as it shown by an artificial very shallow impact crater (after fall of the “Opportunity”’s heat shield), under thin cover of eolian reddish Fe-rich drifts. The salts cover large areas on Meridiani Planum. Their most probable origin is due to widespread hydrothermal activity, vents being craters and deep cracks (faults) draining depths of the highland crust. So, salts are not just a thin veneer but a significant constituent part of the highland crust. That is why Tharsis is surprisingly low in Si. Silicates are partially replaced by salts (low density substance), this is required by necessity to diminish the mean density of highly standing tectonic blocks of Mars – a rotating planetary body that obeys the physical law of keeping equal angular momenta of hypsometrically different tectonic blocks. The aqueous salts with constituent water not only diminish a mean rock density and explain the presence of hydrogen (H2O) at the equatorial zones in ‘Odyssey’ data but also bring down the melting temperature of impacted rocks making ejecta easily flowing (like a dough for pan-cakes).

The “Spirit”s results for the Columbia Hills at Gusev crater go farther [5, 6, 7]. They bluntly shown that an outlier of highlands consists of rocks completely different from lowlands basalts. Keeping the same as in basalts level of Si, these light color rocks are higher in alkalis, Al, P, Cl, S, Ti and lower in Fe, Mn, Mg, Ca, Cr. Al/Ca increases from basalts to this rock. Na is up to 4.5%, Al- 8.0%, P – 2.3%, Ti – 1.5% - these are values suitable for syenites. The thinly layered rocks of Columbia Hills resemble very much the layered nepheline syenites of the ring complexes of Kola Peninsula (Khibiny, Lovozero). One layer of the Columbia Hills massif (Wishstone rock) is high in phosphorus (calculatedapatite is up to 13%), another – in sulfur (Peace rock). All this resembles layers of apatite and feldspathoids rich rocks of Lovozero massif. Sodalite (Cl and S) and noseane (S) rocks are known there. Near the top of Columbia Hills at the “Independent Rock” target researchers have found less iron than expected (Internet, July 11, 2005: “Spirit Scuffs” communication). Again, sharp Fe variations in thin layers of nepheline syenites (urrites, lujavrites, foyaites) are very typical. In some micro images of the Columbia Hills rocks (Internet) one can discern directional (lujavritic, trachytic) and massive (foyaitic) textures. Some larger light colored crystals contain darker isometric inclusions – a hint of the poicilitic structure. A very typical process of feldspathoid alteration in the contact with water (ground and atmospheric) is formation of zeolites. These very soft and low density silicates (often shining like water ice) are perfect sinks for water, giving it out and taking back into their crystalline structure depending on P-T conditions and availability of water [8]. These syenites often containing also as a matrix albite – sodic plagioclase (detected by “Spirit”s Mini-TES [7]) are melted at temperatures much lower than basalts. Maybe, this is a reason why the lobate craters are widespread on the southern highlands. In addition, zeolitized syenites are soft, this property of Columbia Hills rocks was found by “Spirit”s grinding instrument –these rocks are much softer than basalts [7]. Very thin layering of rich in alkalis and thus very fluid nepheline syenites is typical at Earth.

Gusev crater lies at the contact between lowlands and highlands. In an earlier work [9], before “Pathfinder” landing, we insisted on “Possibility of highly contrasting rock types at martian highland/lowland contact”, namely on finding albrites, syenites, granites in addition to basalts.
“Pathfinder” has found andesites, but more acid and alkaline lithologies were discovered by THEMIS (MGS) and “Spirit”. Two localities at Syrtis Major have dacites originated probably from a crustal body long not less than 95 km [10]. “Spirit” has nepheline- normative rocks probably rich in salts penetrating them in form of own minerals and in feldspathoids structures (sodalite, noseane, and others). So, salts helping to diminish rock density, simultaneously lower its melting temperature what helps to produce such characteristic martian structures as lobate craters.

One more remark. This water, alkali-rich easily fluidized crust is often masked by ubiquitous eolian dunes and drifts rich in Fe-minerals originating from the northern lowlands. Now orbiting Mars “Mars Express” with “Omega” instrument measuring reflected light from drifts discovers signatures of not only salts but also olivines and pyroxenes [11]. Researches often make hasty conclusions about wide presence of basic rocks. Sometimes they are right because the presence of plateau basalts, basic sills and layered basic intrusions at highlands is quite possible (compare with Earth), but often they are wrong taking surface reflectance from widespread Fe-minerals surface contamination for an indication of the deeper geology. In this sense, lobate craters luckily sample deeper horizons and better show the real geology.


MARS ATMOSPHERE AND CLIMATOLOGY WITH MARS-EXPRESS: MAIN RESULTS OF EXPERIMENTS WITH RUSSIAN PARTICIPATION. O.I.Korablev1, L.V.Zasova1,3, A.A.Fedorova1, A.V.Rodin1,2, N.I.Ignatiev1,3, A.K.Rybakova1,2, V.Formisano1, J.-L.Bertaux4, J.-P.Bibring5, and all members of PFS, SPICAM and OMEGA team, 1Institute for Space Research, RAS, Moscow, 117997, Russia, korab@iki.rssi.ru 2Moscow Institute of Physics and Technology, Dolgoprudny, 141700, Russia, 3Istituto di Astrofisica Spaziale, Rome, Italy, 4Service d’Aeronomie, 91371 Verrières le Buisson Cédex, France, 5Institut d’Astrophysique Spatiale, Batiment 121, 91405 Orsay Campus, France.

Introduction: More than one and half year passed since insertion of Mars-Express into Martian orbit on 25 December 2003. The spacecraft carries instruments HRSC, OMEGA, PFS, ASPERA and SPICAM [1,2,3] which are heritage from unsuccessful Mars-96 mission. This paper highlights main results on the Martian atmosphere and climate obtained by three instruments, PFS, SPICAM and OMEGA, and with Russian participation. Temperature field, concentration of minor constitutes like water vapor, ozone, methane and CO, H$_2$O and CO$_2$ ice clouds, opacity, aerosol content and vertical distribution, were observed during almost one Martian year. These data are critical in understanding the dynamics, photochemical processes and history of Martian atmosphere, challenging new theoretical studies with general circulation models.

Temperature field: Temperature profiles from the surface to 60 km altitude, and their variations versus season and local time, are retrieved at each orbit on the basis of thermal emission in the vicinity of 15-μm CO$_2$ band.

Figure 1: Temperature field (isolines) vs. altitude and latitude along orbit 68, which passes across Tharsis (j0401S), Ascraeus Mons (101N) and marginal area of Alba Patera (401N). [4]

Due to high spectral resolution, a fine structure of temperature inversion in the polar region, related to descending branch of Hadley cell ($L_s$=342°) was observed as low as at 10-20 km. The winter temperature inversion is caused by downdraft of the air mass advected from the summer hemisphere by the main Hadley cell branch, and subsequent adiabatic heating. As the heating occurs at relatively high altitude (30-60 km), lower atmosphere controlled by radiative transfer processes appears colder, which renders as sharp temperature inversion.

Water vapor: The water vapor was observed by PFS, SPICAM and OMEGA instruments. Seasonal and spatial distribution of H$_2$O was obtained by SPICAM (in 1.38 μm band) and PFS (2.56 and 20-40 μm). Figure 2 shows seasonal distribution of H$_2$O from SPICAM IR channel and special distribution from PFS 20-40 micron band.

![Water vapor](image1)

Fig. 2 Seasonal distribution of water vapour measured by SPICAM IR (orbits 8-1640), and areographical distribution of water in Northern spring by PFS ($L_s$=330°-60°). Apparent zonal structure revealing a strong equatorial maximum at 10°-45°E and a weaker maximum at 200°-240° suggests contribution from stationary planetary waves to the global water cycle on Mars.

O$_2$ airglow at 1.27 micron in nadir viewing: Mars-Express implements 4 methods of ozone detection: UV stellar occultations; UV solar occultations; UV nadir viewing (ozone total column density); O$_2$(1Δ) emission at 1.27 μm (high altitudes). The latter approach gives an access to ozone density above 20 km. Measurement of ozone in the NIR range are also possible for SPICAM and OMEGA.
A singlet oxygen $O_2 (^1\Delta_g)$ dayglow at 1.27 $\mu$m was predicted just after the discovery of ozone on Mars by Mariner 9 [5]. On Mars the situation is similar to Earth, where a strong airglow arises from $O_2 (^1\Delta_g)$ produced by ozone photolysis. For the first time this emission was observed from the ground at high resolution by Noxon et al.[6]. The mapping of this emission was reported by Krasnopolsky and Bjoracker [7]. Krasnopolsky [8] argues that the $O_2$ emission provides even better insight to photochemistry than ozone, since it is more sensitive to the variations of the water vapour saturation level (10-35 km) than total ozone, which remains nearly constant.

**Fig. 3.** Seasonal distribution of $O_2 (^1\Delta_g)$ emission in Mars atmosphere (SPICAM). Colour code is in MR.

Ozone quantity is controlled by the abundance in odd hydrogen species (HOx). HOx are produced by the photolysis of water vapor. Expected anti-correlation between ozone and water vapor are well seen by OMEGA and SPICAM (fig.4).

**Fig.4** Autocorrelation of O3 and H2O (OMEGA – orbit). Top: 1.38 micron H2O band, bottom: Intensity of the 1.27 $\mu$m O2 emission.

**Vertical distribution of aerosol:** All spectrometers are capable of studying the distribution and composition of Martian aerosols. Observations of water ice clouds, insight into their microphysics and the discovery of fine fraction of submicron particles extending high up in atmosphere are among the important results of Mars-Express. In particular, vertical profiles of aerosols were observed by SPICAM IR channel in solar occultation mode (fig.5).

**Fig.5.** Slant optical depth and extinction profiles are obtained by SPICAM IR channel in range 1274 micron. Lower north polar profile contrasts with high extended south middle latitude extinctions.

We believe that the specific features of the vertical distribution of high-altitude aerosols, including evident depression near the poles and subtle inversions in midlatitudes, have dynamical origin. This conclusion is based on the numerical experiments with the GFDL’s Mars GCM including interactive transport and coupled with ab initio microphysical description of water ice clouds. In the equinox season, Hadley cell circulation lifts dust and cloud particles in the equatorial latitudes up to 35-50 km and then advects them out of the equator in the two symmetric branches. In midlatitudes (approximately at 45-60°o) the advective poleward flow fades. Larger particles settle out, while smaller ones are accumulated in the convergent areas of the circulation pattern. It is this location where SPICAM solar occultation data suggest inversion of the vertical dust profile. Upper polar latitudes are characterized by very weak circulation with dominating downdraft vertical air motion. Therefore those particles trapped in the polar atmosphere are transported downwards until they either precipitate at the surface or advected back to low latitudes within the low-altitude closing flow of the Hadley cell. These causes substantially lower, relative to midlatitudes, vertical extension of the aerosol layer.

**References:**
Introduction: Attempts to use the Moon as a photometric standard for calibration of spectral observations of planets and the Earth’s surface from the space were done many times. In particular integral observations of the Moon were used for this aim. We suggest here a recalibration of previous integral data including telescope and Clementine measurements.

Telescopic integral observations of the Moon and their correction: The integral photometrical observations carried out by Lane and Irvine [1] (a range of phase angles $\alpha$ is $6^\circ - 120^\circ$) and Rougier [2] ($2^\circ \leq \alpha \leq 152^\circ$) can be used for the calibration purposes. The data [1] were obtained in the 9 narrow spectral bands from the range 350-1000 nm and in $UBV$ wide bands. Rougier’s data were acquired in $B$-filter. Unfortunately applicability of these data for calibration purposes is limited because of systematical errors caused by the influence of libration variations (up to 2% errors). The data can be also used to study lunar surface properties. However changing the contribution of maria and highlands to integral brightness with phase angle variations is high and this gives up to 10% of errors. We have developed a procedure to compensate these effects, see also [3].

Describing phase angle dependence of lunar reflectance: To study the phase dependence of brightness of lunar surface, it is convenient to use so-called equigonal albedo (EA) instead of visible albedo [4]. The equigonal albedo of a surface area is reflectance measured at a fixed illumination and observation geometry, when the incidence angle $i$ is equal to the reflectance angle $\varepsilon$ and is equal to half of the phase angle $\alpha$ ($i = \varepsilon = \alpha/2$). Bringing all lunar sites to similar photometrical condition allows us to avoid 3D image transformations, if one takes into account lunar libration variations. We have developed the procedure for the transformation of the integral data from relative intensities into averaged relative equigonal albedo and back [3].

For approximation of the experimental data obtained for all filters we used the following empirical formula:

$$EA(\alpha) = m_1 e^{-\mu \alpha} + m_2 e^{-0.3 \alpha},$$  \hspace{1cm} (1)

where $m_1$ and $m_2$ are empirical constants, $\mu$ characterizes the slope of phase curves (an effective roughness factor). This formula gives good agreement with observations over phase angles $6^\circ \leq \alpha \leq 120^\circ$.

Spectral effect: To study the influence of wavelength on phase dependence of brightness we have constructed a diagram “$\mu$ – wavelength” (Fig. 1). One can see, that at relatively small phase angles ($16^\circ$-$43^\circ$) decrease of $\mu$ with wavelength is observed, and for $41^\circ$-$120^\circ$ $\mu$ is practically constant being approximately 0.7. It means that at relatively small phase angles the phase dependence is formed substantially by micro-topography, and it is observed suppressing the shadow-hiding effect due to increase of transparency of regolith particles with wavelength increasing (illumination of shadows decreases value of $\mu$). At large phase angles the phase dependence is mainly formed by meso-topography and almost does not depend on wavelength.

Variations of integral lunar spectra with phase angle: Unfortunately because of internal calibration problems the Lane-Irvine multispectral observations do not allow one to study directly the real spectrum of the Moon. However we have studied the behavior of a relative slope of the lunar spectrum with phase angle. For each filter a fit of phase curve to experimental data with formula (2) was done. Using obtained fitting parameters, model curves for $6^\circ \leq \alpha \leq 120^\circ$ with a step $\Delta \alpha = 1^\circ$ were calculated. This allows us also to
obtain spectral curves for each phase angle. Each of these curves was normalized on the spectrum at 6°. For all relative spectra a linear regression was found. The slope of the relative spectrum is equal to slope coefficient of the regression line. Blue curve in Fig. 2 has been constructed using this procedure. One can see that there are the minimum near $\alpha = 10^\circ$ and the maximum near $\alpha = 50^\circ$. That confirms early uncertain conclusions of some authors on phase dependence of the lunar color-index and coincides very well with the result of model calculations [5]. For comparison we show also observation data (red plot in Fig. 2). The behavior of both of curves is generally similar (note that the blue curve is not a RMS approximation of red points).

![Fig. 2. The dependence of relative slope of the lunar spectra on phase angle.](image)

**Integral brightness variations from Clementine data:** Clementine UVVIS data are unique whole-Moon survey characterized by 100m/pix spatial resolution in several spectral ranges from near-UV to near-IR [6]. The Clementine data are available in form of digital photometric mosaics. To characterize the lunar integral brightness with these data, we averaged all pixels over the whole lunar disk with the center at the zero latitude, varying the longitude (changing with 10° step). In other words we calculated the dependence of the integral brightness on the lunar disk rotation. In Fig. 3 we present five such curves for wavelengths from 0.42 to 1.00 microns. The symmetry center (and minimum) for all curves is shifted to -10° (west longitude) value. To make a comparison we normalize each curve at its minimum. The changes of brightness by factor of 1.30 - 1.35 with rotation phase reveal the mare-highland dichotomy of the lunar nearside and farside. When the central longitude is gradually approach to the farside, all the curves clearly show increasing the color-index (IR / UV) of the lunar disk. We note the importance of these estimates for integral disk photometry (especially for the farside) performed onboard of lunar space missions.

![Fig. 3. Dependence of integral brightness of the Moon on its rotation phase from Clementine observations.](image)

**Conclusions:** Our posteriori processing of integral observations of the Moon [1, 2] provides more reliable phase curves of lunar brightness. Using these data together with the Clementine integral brightness curves enables using the Moon as a photometrical standard for photometry from spacecrafts.

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FLUVIAL CHANNEL RESULTED IN ALLUVIAL FAN FORMATION IN ICARIA PLANUM, MARS.

J. Korteniemi¹, J. Raitala¹, M. Aittola¹, V.-P. Kostama¹, E. Hauber², P. Kronberg³, G. Neukum⁴ and the HRSC Co-Investigator Team. ¹Astronomy, Univ. of Oulu, Finland, (jarmo.korteniemi@oulu.fi, jouko.raitala@oulu.fi), ²DLR, Berlin, Germany, ³TU, Clausthal-Zellerfeld, Germany, ⁴Freie Universität, Berlin, Germany.

Introduction: The Claritas-Thaumasia region is highly modified by tectonic forces. It has several indications of previous water activity phases (Fig. 1, 39°S, 258°E; Raitala et al., 2005). The Mars Express HRSC data were used in finding additional details of its fluvial history.

The Regional Background: Volatiles were transported from the peaks (1 in Fig. 1) into the basins like the one studied here, found in southern Claritas Fossae. Water filled the paleolake (2 in Fig. 1) up to the level of the lowest valley, breached through the saddle valley and formed a channel (3 in Fig. 1) from the paleolake to the west.

Along the channel, sapping from an old impact crater in the north (4 in Fig. 1) provided additional water. Close to Icaria Planum, the channel broke into an impact crater (5 in Fig. 1) and formed a temporary lake with a delta at the channel mouth. The flow breached further through the western crater rim (6 in Fig. 1). The crater floor is lower than this channel neck indicating a paleolake phase. Water spread onto the Icaria Planum lowlands (7 in Fig. 1) and formed an alluvial fan. The structures and alluvial fan units in Icaria Planum, visible in the MEX-HRSC data, were studied using the color HRSC images.

The Channel-Related Features: At some point when the wide basin in the southern Claritas area ("paleolake" in Fig. 1) was filled, the water broke through the western saddle valley forming a channel from the paleolake into Icaria Planum (Figs. 2,3).

The basin depression in the south was subsequently drained into the direction of Icaria Planum.

Fig. 2. The RGB combination of the three visible HRSC channels of the MEX orbit 357 over the area where the channel breached through the saddle valley into the west.

Fig. 3. The units identified along the course of the channel.

The flow channel begins from within the middle basin area (upper right part in the HRSC red channel image; Fig.3). It broke the paleolake rim at its lowest area (image center in Fig.3) and drained the lake through an impact crater (left in Fig.3) and further into northern Icaria Planum. Additional sapping channels were formed by the water flowing from the direction of the large crater (top in Fig.3).

Alluvial formations: The channel ends into the 30-km wide impact crater that was, at least partially,
filled by water (Fig. 4). A temporary paleolake phase is indicated by the morphological features in the crater: A) The delta within the crater at the mouth of the channel was formed in a standing body of water. B) The crater rim has terraces and its floor was smoothened by sedimentary deposits. C) The neck of the outflow from the crater further into the west is higher than the floor of the crater.

The four-channel HRSC data set allows to map materials, units & formations: During the period of lacustrine environment in the crater its western rim breached and part of the water was led out onto the Icaria Planum lowlands. The water-carried particles were spread as sedimentary flood deposits onto Icaria Planum in front of the short channel out from the crater. These alluvial deposits are made visible by the unsupervised four-channel MEX-HRSC image classification (right: red = shadows; purple shades = higher grounds; brown & yellow & dark green = identified deposits).

**Fig. 4.** The HRSC RGB image from the Mans Express orbit 068 indicates the flooded impact crater and the related formations.

**Fig. 5.** The four-channel MEX-HRSC classification shows shadows with red, higher grounds with purple tones and alluvial deposits with brown, yellow and dark green.

**Conclusion.** The region is ideal in studying the fluvial channel formation tied to climate changes, hydrothermal activity and local geology. Melting of snow and ice from the peaks surrounding the major basin was the most probably responsible for the formation of the paleolake and the adjoining channel. The resulted alluvial structures reflect the amount of water available, topography and regional slopes along the course of the channel. The hi-resolution multi-channel MEX HRSC data give advanced views into the alluvial structures, erosion and sedimentation in the channel formation processes as well as into other geologic features of the area. Advanced remote sensing approaches will facilitate further mapping of the characteristic phases in development of faulting, volcanism, morphology and other geology within the Claritas-Thaumasia area.

WATER AND ICE IN CENTRAL NOACHIS TERRA, MARS? J. Korteniemi¹, M. Aittola¹, T. Öhman², T. Törmänen¹ and J. Raitala ¹, ¹Division of Astronomy, Department of Physical Sciences, P.O. Box 3000, 90014 University of Oulu, Finland (jarmo.korteniemi@oulu.fi), ²Department of Geosciences, Univ. of Oulu, Finland.

Introduction: The studied area (Fig. 1) is located in the central part of the Noachis Terra on Mars (36-47°S, 20-30°E). The region is part of the southern highlands W of the Hellas basin, and has been generally described as ancient terrain with large, eroded craters modified by e.g. fluvial and aeolian processes [1]. Impact craters in the whole Noachis Terra exhibit many, mainly intracrater dune fields [1,2], as well as several examples of depressions and collapses on the crater floors [3,4].

Data and methods: We have studied the area using the freshest data sets available – THEMIS and MOC – in conjunction with Viking imagery, to find out what input they can give to the geological analysis of this highland region. The topography is determined from the MOLA 128 pixel/degree DTM.

Tectonics: The Noachis region has a large SW-NE graben system ~2000 km to the (N)W from Hellas basin [1]. All tectonic structures in our study area are roughly parallel to 1) the graben set, 2) Hellas basin and 3) Hellespontus Montes. This may indicate that the local tectonics is controlled by the Hellas impact event. This interpretation was concluded from the study of polygonal craters in the Hellas area [5].

A large 102-km crater (at 45°S, 28.5°E) exhibits depressions on its floor. The depression walls are layered, steep and up to 300 m high. Additionally, they are often straight and follow the regional trends [3], radial to Hellas basin and the main tectonic lines.

Aeolian features: The intracrater dune fields of the southern, mid and high latitudes – including Noachis Terra – are thought to be the most significant accumulations of sand on the planet [6], and dunes were first identified in the Noachis Quadrangle [7]. Therefore, the dune fields in the region have been targets of a number of studies [e.g. 2 and references therein]. An interesting observation is that there are almost no dune fields in the studied area, despite their regional abundance. The only dune field lies on the floor of the aforementioned 102-km crater.

Water activity: The region does not have any huge-scale fluvial features such as giant outflow channels like the E side of Hellas. However, it does exhibit evidences for a multitude of small-scale fluvial activity. The most prominent examples are the several channels found in the study area (dotted lines in Fig. 1). They terminate mostly in local low elevations, such as impact craters or other basins. These have usually smooth interiors, which appear dark in THEMIS day-IR images and bright in night-IR.

There are also few candidates for lake chains, which are not uncommon features in the Hellas region [8].

Figure 1: MOLA topography of the studied region (36-47°S, 20-30°E). The dotted lines are channels (mapped from Viking images). Deposits at the end of a long channel, in a depression, is shown in more detail in Fig. 2 (large box). Fig. 3 is a close-up of the small box area.

Deposits in depressions: The topographical depressions associated with the channels exhibit a distinct albedo, seen especially well in THEMIS IR images. Two examples are shown, one in the N part of the study area (Fig. 2a; 37.4°S, 28.1°E) and the other in the S (Fig. 2b; 43.7°S, 24.9°E). In Viking images both appear smoother than the surrounding higher terrain. The basins are rather polymorphous in THEMIS day images, as they are generally darker than the surroundings, but have also some albedo variations within the lowland. The darkest areas (in day-IR) appear very bright in night-IR. These regions are the ones directly connected to the incoming fluvial channels. THEMIS images indicate the thermophysical properties of the deposits, showing that the ‘night-bright’ basin material is not fine grained but rather rocky or consolidated. The channels also show up bright in night-IR, probably caused by a concentration of coarse – and thus warm – material on the slopes relative to flat surfaces [9].

The basin deposits are without doubt different from 1) the surrounding smooth area as well as 2) the highlands. The incoming channels give a reason to suggest that these deposits are probably the accumulations of liquid(s) which used the channels as routes to topographically lower areas. If that is the case, the deposits must be consolidated material, as it is bright in night-IR. Unfortunately there are no complete high-res image sets (THEMIS VIS, HRSC...
or MOC). Thus crater statistical age determination has not been done. However, based on crater mapping, there are just a few 10–40 km craters superposing the smooth basins; the highlands in turn exhibit many more craters of similar size, indicating that its surface is older.

The intercrater plains that appear bright (warm) in night-IR have also been explained by widespread sedimentation across Noachis [2]. These areas would thus represent remnants of Noachis-wide sedimentary unit(s) which have later been exhumed. However, this idea does not consider the incoming channels. In the same paper it is proposed that in Noachis Terra sand is not transported far from its origins, ruling out a distant source of sand. If this is the case, where did the material overlying the night-bright deposits move – there are (almost) no dunes in the study area?

**Evidence of ice:** Pingo s are described as cone-shaped mounds with cores of ice. They are formed by the upward expansion of freezing water surrounded by permafrost. Studies have proposed that there are evidences of pingo-like structures on the Martian landscape, e.g. within a fluvial channel [10]. It has also been proposed that the central mounds of craters could have pingo-like origins [11,12].

There are small (diam. 20–120 m; Fig. 3) mounds inside a crater (45.98°S, 24.41°E). The origin of the material unit they reside on seems to be a channel which breaches the crater rim in SW. The peaks are strictly connected to material apparently accumulated from the channel, and do not appear elsewhere on the crater floor. If the channel has offered a source of additional water, the mounds may very well have been created as hydrostatic pingos. Loss of local water, permafrost aggradation and the formation of a sub-surface ice core could have formed the peaks.

**Discussion and Conclusions:** Although the studied area does not show evidence of massive outbursts of local water, there is clear proof of small-scale water activity in the region. The most evident examples are the channels, which are usually associated with craters and basins. As a result, crater lakes and even lake chains probably formed. The channel-associated deposits in the lowlands, pingo-like features and the collapses on the crater floor may all indicate that there have been and probably still are reasonable amounts of water/ice/permafrost below the surface. Thus this area should be of great interest to future investigations, e.g. by the OMEGA, HRSC and MARSIS instruments on the Mars Express probe.

According to this preliminary study, Noachis Terra has been modified by several processes, which have characterized the unforeseeably varied geological history of the region. The after-effect of the Hellas impact event is evident, as can be seen from the orientation of the tectonic structures.

**References:**
THE EVOLUTION OF THE REULL VALLIS FLUVIAL SYSTEM, MARS. V.-P. Kostama1, M. Ivanov1,2, J. Korteniemi1 and J. Raitala1. 1Planetology group, PO box 3000, FIN-90014 University of Oulu, Finland; 2Vernadsky Institute, Moscow, Russia; <petri.kostama@oulu.fi>.

Introduction: The region between Hesperia Planum and Hellas basin is one of the main areas on Mars where large outflow channels occur [1-7]. One of these, Reull Vallis, begins in Hesperia Planum and runs from east to west across the northern portion of Promethei Terra. Here we present a new hypothesis for the evolution of Reull and its fluvial system. We suggest that the fluvial system of Reull Vallis consists of several parts that were formed during several distinct phases.

Data and methods: We defined and analyzed the different parts of the fluvial system and correlated temporally the processes that led to their formation using available images and topography data (MDIMs, MOC, THEMIS (IR and VIS), HRSC, and MOLA-gridded topography (128 px/deg)). Crater counting served to derive relative and model absolute ages of different portions of the Reull Vallis system; measurements from MDIM and MOLA were used to estimate volumes and their balances for different parts of the system.

Parts of the Reull Vallis fluvial system: The Reull Vallis fluvial system consists of five parts (Fig. 1). The first, northernmost part (depression with provisional name "Morpheos Trough"), begins within the southeastern portion of Hesperia Planum at ~31°S, 246.5°W. At ~32.5°S, 246°W, it is transformed into a morphologically distinct channel, "Morpheos Vallis" (segment 1 of Reull of [8]), which represent the second part of the system. This channel runs southward, and disappears at the northern edge of "Morpheos basin" at about 35°S, 246°W. The Morpheos Basin is the third part of the system; it represents a closed topographic depression elongated in W-E direction in the SE portion of Hesperia Planum, which received and stored water from Morpheos Vallis. Morphology of Morpheos Vallis suggests a catastrophic outflow through it. If we assume velocity of the flow to be 60-70 km/h, it would take ~5-8.5 days to fill the basin up to 650 m contour (~200 m deep); the volume of stored water is ~11-17 x 10^3 km^3. Morpheos basin appears to be the source area of Reull Vallis (segment 2 of Reull of [8]). The fourth part of the fluvial system is the upper Reull Vallis (Fig. 1). It begins as a full-sized topographic and morphologic feature at the western edge of the Morpheos Basin (~37.5°S, 247°W). For the largest portion, the channel of Reull Vallis is on a very shallow slope (Fig. 2). At ~42°S, 254°W, slope of the channel increases, whereas the regional slopes on both sides of it remain to be the same (Fig. 2). Volume of the upper Reull is ~1700 km^3. The fifth part of the system is the lower Reull Vallis (Fig. 1). It starts at ~42.5°S, 257°W as a broad and deep canyon and runs to the west along a shallow slope, which is similar to the regional slope outside of the channel and to the slope of the upper Reull (Fig. 2). Volume of the upper Reull is ~8300 km^3. The longitudinal profile (Fig. 2) shows that Teviot Vallis, an apparent tributary to Reull Vallis (its volume is ~3100 km^3), continues topographic trend of the lower Reull.

Apparent sequence of events: Morphologic characteristics and topographic configuration of different parts of the Reull Vallis fluvial system (especially a major break in slope separating the upper and lower Reull) suggest the following sequence of major events during formation of the system. The oldest parts of the system appear to be the Teviot-lower Reull Vallis and the sub-system of Morpheos Trough-Vallis-Basin. Probably, these two oldest parts formed independently; relative ages between them cannot be established. Within the Morpheos sub-system, a flow below the lava plateau removed supporting material and resulted in formation of the trough due to subsidence of the lava plains. When the flow broke through to the surface, it carved the Morpheos Vallis, filled a part of the Morpheos basin, and formed there a standing body of water. The last episode of the evolution of the Reull Vallis fluvial system was formation of the upper Reull due to discharge of the Morpheos basin. Topography along the upper Reull and secondary channels near its end suggest that water was pounded during formation of the upper Reull. Such a pounding may have resulted in erosion, deposition, and resurfacing of an area around both the upper and lower Reull. Note that the volume of water that potentially may be stored within the Morpheos basin is about an order of magnitude larger than the volume of material removed from the upper Reull.

Figure 2. The longitudinal topographic profile of Reull Vallis consists of two distinct parts, the upper and lower Reull separated by a major break in slope. A "tributary" to Reull (Teviot Vallis) continues the topographic trend of the lower Reull.
Crater counting: In order to estimate ages in different portions of the Reull Vallis fluvial system relative to Hesperia Planum and to each other, we have counted craters in three large regions: 1) Hesperia Planum ($1.50 \times 10^6 \text{ km}^2$, 3266 craters, 1.2-49.9 km in diameter), 2) Morpheos Basin ($0.24 \times 10^6 \text{ km}^2$, 357 craters, 1.2-31.4 km in diameter), and 3) Reull Vallis region (both upper and lower Reull, $0.27 \times 10^6 \text{ km}^2$, 261 craters, 1.2-39.7 km in diameter). The crater size-frequency distributions show that the Morpheos and Hesperia curves are practically coincided, while the Reull Vallis curve is distinctly lower (Fig. 3). Thus, the crater retention age of the Morpheos basin is indistinguishable from that of Hesperia Planum and the area around Reull Vallis is younger, which is consistent with apparent sequence of events during formation of the Reull Vallis fluvial system.

Conclusions: Our analysis suggests that the Reull Vallis fluvial system consists of several distinct parts that have different origin and age. The whole evolution of the system appears to be consisted of three major episodes: (1) formation of the lower Reull (apparent beginning of it is the Teviot Vallis in ~44°S, 258°W), (2) formation of the Morpheos fluvial sub-system (these two episodes may or may not be contemporaneous), and 3) formation of the upper Reull that connected the Morpheos sub-system with the lower Reull.

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).

Analogs of ice spiral structures are not known. The troughs have been attributed to the action of aeolian erosion [2-3], sublimation [4] or to “accumulation” hypothesis (glacial flow + sublimation + accumulation) [5-7]. It is supposed that an ice mass transfer occurs by sublimation from equatorward-facing slopes and subsequent accumulation on pole-facing slopes. No ideas on origin of spiral pattern have been moved forward with the exception of an attempt to explain trough revolving by combined effects of accublation and ice movement [8].

Hypothesis: Analysis made by Fishbaugh and Head [9] suggests that Chasma Boreale - the greatest trough of the north pole ice sheet - is a giant scour generated by subglacial water outflow. This mass of water can arise in consequence of a number of reasons, such as intensive ice sheet melting provided by geothermal flux increasing. It is natural that water lubrication decreases bed friction. In view of this the ice sheet will spread radially with high speed. Augmentation of radial stress over the breaking point generated ice that fractured entirely. The last takes place in the zones of maximal shear strength, which look like as helical glide lines, producing a helical system of crevasses.

Basing of hypothesis: theoretical investigation:
Start from the assumption that a stationary plane stress state of an axisymmetric elasto-plastic body (with no bed friction) satisfies the equation of equilibrium of the tension and gravitational forces and Mizes yielding condition.
In the present epoch, it seems, the parabola is the best approximation for the profile of the Martian north pole ice dome [10]. Thus, under a parabolic profile the gravitational force that causes the ice sheet to widen, increases proportionally with increasing radius. Under our statement of this problem, the central part of the ice sheet remains elastic whereas the outside is in the state of plastic yielding (Fig. 2).

The formula for the boundary radius between elastic and plastic region is the following

\[ r_s = \sqrt{\frac{2\tau_s}{gH\rho(1+\nu)}}R \]  

(1)

where \( \nu \) is Poisson ratio. (For the first time the solution for uniformly strained axisymmetric elasto-plastic body has been found in [11,12].) Experimental values of a limit shear stress lie within the limits 0.1-1 MPa [13]. Providing \( \tau_s = 0.1 \) MPa and \( g = 4 \) m/sec², \( H = 2800 \) m, \( \rho = 10^3 \) kg/m³, \( \nu = 0.3 \), and \( R = 6 \times 10^5 \) m, the radius of elastic kernel is equal to 73 km (~ 0.1 R). Providing \( \tau_s = 1 \) MPa it reaches 227 km (~ 0.3 R). One can see that the plastic zone occupies the larger part of the ice sheet for the whole
range of the limit shear stresses. It is worth noting that its percentage increases along with rising of ice temperature.

For many materials an erratic behavior of plastic yielding, driving to the localization of deformation bands, is observed nearly always on reaching a critical level of irreversible deformation. Laboratory rock testing [14] shows that the deformation is accompanied by evolution of initial microcracks and pores, leading to formation of new faults and modification of material properties. This process depends on both the level of actual tension, and the interaction of crack banks. This leads to dilatation, which is an irreversible expansion of the volume, caused by expansion of pores and exposure of cracks. The most powerful change of material structure proceeds in the vicinity of peak tension, before formation of the narrow fissured macroscopic abnormalities containing an abundance of microdefects. These abnormalities, referred to as “slip/glide lines/bands”, represent localized bands of plastic deformation. Their origin is concerned with an internal instability of material, i.e. with qualitative changes in the damage accumulation. The last reduces to a bifurcation of an initially uniform deformation process [15].

Under plastic flow, as opposed to brittle failure, cracking takes place along glide lines oriented at an angle to the tension axis [16]. Elementary theory of plane stress in a point (under the static balance of forces and moments) gives the value of angle $\phi$ with the maximum tangential stress. It is $\pm 45^\circ$. For the axisymmetric plastic body these glide lines/bands look like as two families of logarithmic lines (Fig. 2).

The resemblance of these lines/bands and the Martian north pole trough structure is undoubted. As Mars (as well the Earth) rotates from the west to the east, in the north Martian hemisphere the cracking took place just in the glide bands bending clockwise. To prove the influence of planet rotation on trajectories of crack propagation look at the following scheme (Fig. 3).

![Fig. 3. Influence of Coriolis force on crack trajectory.](image)

As known, the role of the Coriolis force reduces to a deviation of a moving particle trajectory (to the right in the north hemispheres of the Earth and Mars). In consequence of slight moving of the every particle on the crack bank to the right during cracking, its left part is rather more stretched that its right one. This drives the crack vertex trajectory to the right. Thus, under all other conditions being equal, in the north Martian hemisphere the cracking took place mainly in the glide bands, which are bending clockwise.

**Conclusions:** This investigation shows that after its bed thawed, the Martian north pole ice sheet began to transform to an ice body resembling an ice shelf (no basal traction). This transformation was accompanied by drastic amplification of radial tension that fractured ice entirely with formation of crevasses along helical glide lines. After bed temperature fell and ice sheet collapse ceased, the cracked bands began to undergo a smoothing owing to continuous slow ice spreading and mass transfer from the north crack slope warmed by the sun to the shady south slope. This process transformed the helical structure of crevasses into the helical structure of troughs.

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**References:**
CLIMATE-RELATED ALTERATION OF CRATERS IN THE NORTHERN PLAINS OF MARS.

M. A. Kreslavsky1,2 and J. W. Head1, 1Dept. Geological Sciences, Brown University 1846, Providence, RI, 02912-1846, USA; kreslavsky@brown.edu, 2Astronomical Institute, Kharkov National University, Kharkov, Ukraine.

Introduction: Impact craters are very helpful for studies of many aspects of surface processes on the Solar system bodies. Here we analyze population of impact craters in the Northern Lowlands of Mars, more specifically, within the typical Vastitas Borealis Formation (VBF) [1]. This is the largest geological unit on Mars. It is thought to have rather uniform age approximately at the Hesperian/Amazonian boundary and accumulation population of relatively large impact craters [2]. These craters were presumably formed in rather similar environment and hence initially had similar morphology. Their formation age is uniformly scattered through the Amazonian.

Fig. 1 A – C show examples of craters studied. Crater A has rougher ejecta and obviously fresher ejecta morphology, and hence is younger than crater B. We study systematically morphology and morphometry of the whole crater population to infer information about crater alteration and surface modification processes and their rates through the Amazonian.

Observations. We considered only craters on the VBF with diameters D from 10 to 25 km. This narrow range of D assured the maximal possible similarity of a pristine ejecta pattern and wall morphology. Almost all these craters have double-layer ejecta. [3] For smaller craters, the ejecta are too poorly sampled with MOLA; for larger craters, the multiple layer ejecta dominate [3]. We excluded craters where large dune fields, ejecta of larger craters, etc. complicate morphology. There is a strong latitudinal trend of subkilometer-scale roughness on Mars caused by climate-related surface alteration [4,5]. In this study we limited ourselves to high latitudes >52°N, that is well within the “smoothed” zone. In total there are 141 craters that match these selection criteria.

We use THEMIS VIS and MOC NA images to study morphology of the craters in detail and MOLA data to obtain crater depth, slope steepness and ejecta roughness. As a measure of ejecta roughness, we used the median total surface curvature, which we calculated in the following way. We used gridded topographic data at 128 pixels per degree resolution. In each pixel we calculated two principle curvatures $C_1$ and $C_2$ of the surface at 0.9 km baseline and the “total” curvature $C = (C_1^2 + C_2^2)^{1/2}$. For each of the 141 craters, we manually outlined the test ejecta area as a ring loosely inscribing the inner ejecta lobe and excluding the crater rim and floor (Fig. 1D). For clearly asymmetric craters we drew a non-circular outline. We preferred this subjective way to mark the test areas, because it allowed us to exclude small craters that overlap the ejecta, as well as to treat overlapping ejecta carefully. Steepness of the northern and southern crater walls at 0.3 km baseline was taken from all individual
MOLA profiles that cross the craters close to their diameters. Some results are presented in Figs. 2 and 3.

**Synthesis.** Below we present a summary of characteristics of several crater modification processes that we infer from the observations.

*Smoothing (mantling) of ejecta.* A few craters (like those in Fig. 1A,C) have rough ejecta, while the majority of them (like that in Fig. 1B) have smooth ejecta (Fig. 2). Thus, smoothing of the ejecta occurs at 10s to 100s Ma time scale, short in comparison to the Amazonian (3.1 Ga according to [7]). High-resolution images show that filling of local lows with mantle material is responsible for smoothing. The roughest ejecta already have well-pronounced mantle in the local lows. This means that the surface smoothing at high latitudes is due to several repeating mantling episodes (ice ages) rather than due to the most recent one.

Several studies have shown that the ejecta volume of the typical craters in Northern plains of Mars is noticeably greater than the cavity volume [e.g. 8]. This has been attributed to deflation of material surrounding the ejecta (an effect similar to pedestal crater formation), which increases the apparent ejecta volume, as well as to crater filling, which decreases the cavity volume. Our morphological observations indicate that emplacement of the mantle material in local lows contributes to the increase of the apparent ejecta volume.

*Erosion of crater walls.* Craters with rough ejecta (that is the youngest craters) have relatively steep walls (25-30°; see Fig. 2). Wall steepness of craters with typical (smoother) ejecta varies in a wide range from ~5° to ~30°. Thus the time scale of wall degrada-

tion is longer than the time scale of ejecta smoothing, on the order of 1 Ga.

Almost all craters with relatively steep walls (>~18°) have recent gullies. Erosion of the walls by gully-forming flows [e.g. 9] is one of the mechanisms of wall degradation. In the equatorial region, where gullies are absent, typical crater walls are 30-40° steep, that is steeper than the steepest slopes in the VBF. Gullies are not observed on gentle slopes (<~18°). This is consistent with gullies formed by debris flow rather than water flow [9]. Some other mechanisms are responsible for further degradation of the walls. Generally the amount of crater wall erosion is small, because all craters in the survey have well preserved rims (intense erosion would remove them). This indicates that the climate conditions favorable for gullies formation occurred not frequently in the Amazonian.

*Crater filling.* Depth / diameter ratio of the VBF craters has been known to be scattered in a very wide range [e.g., 10, 11]. The youngest craters (with rough ejecta) are deep. There is a correlation between wall steepness and relative crater depth (Fig. 3). Thus the decrease of crater depth (crater filling) occurs at the same time scale as wall degradation, that is ~1 Ga.

Erosion of the crater walls is not responsible for crater filling: erosion would remove the rims much earlier than make the cavity shallower. We observe migrating sand trapped by craters with steep walls. This is one of the mechanisms that contribute to crater filling. We also observe indication of thick layers of ice-rich material in the craters (concentric crater fill, large-scale polygons etc.). This ice can be deposited from the atmosphere. Alternatively, this ice can be not completely sublimated residue of frozen water flooded the lowlands during Amazonian-age outflow events.

**Future work:** Study of smaller craters with HRSC stereo images can provide better timing estimates. Comparison of the VBF crater population with craters on other terrains is a clue for understanding of latitudinal, elevational and regional trends.

THE USING OF DIFFERENT FORMS OF DEPENDENCE OF PROJECTIL ENERGY TO CRATER SIZE FOR PREDICTION OF PARAMETERS OF ARTIFICAL CRATER ON COMET 9P/TEMPEL 1.

Kruchynenko V.G.1, Churyumov K.I1, Valter A.A.2, Dobryansky Yu.P.3, Chubko L.S.1

1Astronomical Observatory of Kyiv Shevchenko National University, kruch@observ/univ.kiev.ua
2Applied Physics Institute of the National Academy of Sciences of Ukraine, Sumy-Kiev, avalter@iop.kiev.ua
3Varmin-Mazur University, Olshtyn, Poland, dobr@umw.poland.pl

Introduction. The copper impactor spacecraft of the Deep Impact Mission which has mass of 372 kg, and velocity of 10.2 km/s on July 4, 2005 collided the 6 km in size nuclei of short -periodic comet 9P/Tempe 1 at velocity of 10.2 km/s on July 4, 2005. This at first gave the possibility to compare the calculated parameters of the crater on Small low-density cosmic body with experimental data. Using two theoretical models of collision of the impactor with the comet nucleus we calculated that the possible crater diameter on the nucleus of comet Tempel 1, formed by the artificial impactor Two models were used. First one – the Opik’s model [1-3]. This theory is based on transfer of quantity of impulse, instead of energy as many researchers accepted. 

The estimation of the crater sizes by the methods of the impact-explosive analogy which was developed in a plenty of works and appreciably generalized in [4,5]. was also done.

The calculations by the Opik’s model. According this model The impacting body some time continues to move inside of a target (thus the significant part of energy is carried away outside) and for calculation of such movement it is necessary to use the law of conservation of an impulse, instead of conservation energy.

In a bowl of a crater where there is destruction, coupling between various elements of volume are small and the radial moment of a shock wave is remained. Tangential moving of substance are insignificant and radial velocity of impact is equal to the radial moment on a mass unit, belonging wave front. Destruction and rock outburst from a crater is consequence of the action of radial shock wave. In considered model the basis for the analysis and calculations is hydro dynamical pressure, which defines a value of resistance or braking.

Efficiency of hydro dynamical pressure of moving shock front is defined by expression $Cp\rho^2$. The coefficient $C$ depends on the form of a body; compressibility of medium, a Mach number and for the continuous medium is equal approximately to unit, and on the average close to 0.5. Confirmation of that $C \approx 0.5$ follows from calculations jf Gilvary and Hill,1956/ We accept, that a meteoroid striking in a target has the cylindrical form. The initial radius of the cylinder and its height $2R_o$.

We accept, that a meteoroid striking in a target has the cylindrical form. The initial radius of the cylinder (and radius of a frontal surface) is equal $R_o$, its height $2R_o$. During the moment of contact the powerful shock wave extends both in an impactor and in a target. Its velocity of propagation has the order of velocity of impact. It will make essential destructions (mechanical cracking on fragments) in that area where the crater will be generated.

$R^2h = 2R_0^3$  \(1\)

Where R and h - value of radius and heights of the cylinder during the subsequent moments of time.

Data about compressibility of iron under hydro dynamical pressure $P$ in the interval from $7.7 \times 10^{11}$ N/m² up to $2.9 \times 10^{15}$ N/m² and of silicon under pressure from $2.1 \times 10^{11}$ N/m² up to $2.9 \times 10^{14}$ N/m² are made in the paper (Gilvary and Hill, 1956). They show, that, change of density, as functions of pressure ($\rho \sim P^{\gamma}$), for iron and silicon is insignificant as return value of an exponent of an adiabatic curve $1/\gamma = 0.28$ and 0.30 accordingly. For gases this value more: for one-nuclear gases it is equal 0.60, for two-nuclear one - 0.71.

A velocity $V$ concerns to the center of the impactor mass, $V_1$ - a velocity of motion of a frontal surface. Accepting, that the value of velocity of different parts of an impact or increases linearly from the frontal surface to the back one, the velocity of the back surface, obviously, will be equal $2V - V_1$. Then velocity of change (reduction) of height of the cylin-
der in the course of time \( \frac{dh}{dt} = -2(V - V_1) \). Let \( V_2 = \frac{dR}{dt} \) - a velocity of increase in radius of a cylindrical body. Change of quantity of movement on unit of cross-section of an impactor is represented in the form of (2),

\[
\sigma = \frac{\rho V_2^2}{2} + \sigma_p \tag{3}
\]

Using a known hydrodynamical principle, we shall write down an equation

\[
P_1 - P_2 = \frac{\delta V_2^2}{2} \tag{4}
\]

The analysis of the numerical decision of the equation (4) testifies that the maximal braking is reached on the depth equal \((0.4 - 0.5)h\), i.e. it gives a result close to conclusion of Öpik.

Omitting some intermediate considerations the result for the diameter \( D \) of a crater jne can give as

\[
D/d^{3/2} = 1.20(\text{kV}\delta/h)^{1/2}(\rho\delta)^{1/4}
\]

From this dependences follows, that depth of the formed crater will make \(4.8\ldots5.6\) m, and the size of diameter which will be equal \(22\ldots57\) m, the volume of the destroyed substance (volume of a crater). will make \(810\ldots7080\) m³.

### Model of impact-explosion analogy

From this analogy follows, that the craters which have formed at meteoritic impacts with space velocities are equivalent to the craters made by explosions of little buried charges. Processing of a big number of observed data, and also explosive and impact experiments [Mellosh, 1994] leads for the big range of the sizes to following dependence for diameter of a crater: \( D = LE^{3N} \), where \( E \) - energy of its formation, in this case kinetic energy of an impactor, \( L \) - constant coefficient, \( N = 3.4 \). If \( D \) is expressed in meters, and \( E \) in joules \( L = 0.02 \) For large events can appear dependence [Mellosh, 1994] more preferably:

\[
D = 0.773\left(\frac{g_0}{g}\right)^{0.118}E^{0.294} \tag{18}
\]

Where \( D \) in km, \( E \) in Mr, \( g_0 \) - acceleration of free falling on the Earth, \( g = 0.07 \) sm/c² - on a nucleus of a comet. Calculation using the formula (18) leads to value \( D = 65 \) m.

Depth of craters of such size is accepted [Dence, 1973] as \( h = D/2\sqrt{2} \), that is \( h = 23 \) m.

It seems to be that this estimation will be the closest to experimental result.

### References


APPLICATION OF ADAPTIVE DYNAMIC REGRESSION MODELING FOR PROCESSING AND THE ANALYSIS OF SOME CHANGES OF DURATION OF TERRESTRIAL AVERAGE DAY

S. V. Kurkina  
Ulyanovsk State Technical University, 32, Severny Venez st., 432027, Russia  
sgv@ulstu.ru

Introduction: Models of changes of duration of day allow us to predict not only duration of day, but also changes of average speed of rotation of the Earth. Non-uniformity of rotation of the Earth and movement of poles are caused not only by gravitational influences of environmental heavenly bodies, but by processes proceeding on our planet, and depend on features of the structure and physical properties of terrestrial bowels.

For precision treatment and the analysis of time lines in [1] the approach of adaptive dynamic regression modeling (DRM), as a software package - the automated system DRM is offered.

The set of DRM algorithms includes algorithms of designing of approximations, valuation and structural identification; and also procedures of formation of criteria of quality of approximations, analytical and graphic criteria of performance of conditions of application of classical circuits regression analysis - a method of the least squares (MLS) and computing procedures of modeling of time lines [1] are used.

Construction of models and their analysis: As the initial data the given changes of duration of the day, calculated by International Earth Rotation Service (IERS) with step-type behavior 1 day (eopc04) for 1995-2004 (3653 supervision) have been taken.

At the first analysis stage of the data within the framework of the DRM-APPROACH the checked hypothesis about stationarity lines has been rejected with probability of 95 %.

At the second stage construction of harmonious model or allocation of the trend component were supposed to be made.

The model of a square-law trend with the factor of correlation R=0,81, being optimum of 17 constructed dependences is constructed at a significance value 0,05 on F-statistics: Fтабл = 3,84; Fфакт = 3470,64; σΔ = 0,000604. At each stage of processing the detailed analysis of the rests was carried out. For this purpose modules of check of observance of MLS assumptions and the analysis of quality of the model have been connected with the DRM library, allowing to estimate the degree of adequacy of model to supervision and a degree of suitability of model for approximation in the given selective space.

At the third stage the schedule of autocorrelation was investigated and the factor DW=0,0084 paid off; the conclusion about possible presence of periodic components in the rests was made.

By results of spectral and wavelet analyses of the rests 32 harmonics are allocated. Bearing harmonics with the maximal spectral density appeared to be harmonics with the periods of 365 days; 183; 13,6; 28; 1217; 9,1; 63; 91. Average quadratic deviation (AQD) of the model is equal to 0,00035; AQD on external accuracy (σΔ = 0,000434).

On the fig. 1 there is a schedule of the rests after allocation of a trend component and the schedule on the model received as a result of application to the line of the rests of the spectral analysis. On a correlation matrix of harmonics with the set periods correlation between harmonics is not found out. The carried out wavelet-analysis allowed to reveal a harmonic with the period 422 days. Inclusion in model of a harmonic with this period reduces AQD up to 0,000345 and AQD on external accuracy (σΔ = 0,000392) a little.

Residual fluctuations are smoothed out by autoregression model of the suitable degree, or by methods of martingale approximations (or consecutive application of these two approaches). The model of autoregression of the second degree with AQD 0,00005 is constructed; σΔ = 0,000385.

As a result the time line is submitted by a mathematical model, additively including square-law and periodic trends, and also autoregression. On Fig.2 there is the schedule of complex model of changes of duration of average day for the period since 1995 till 2004.

As a whole for this model the assumption of equality to zero of the mathematical expectation is carried out, model is undetermined, the rests are not allocated under the normal law and homoscedastic, and there is a positive autoregression.
On the schedule (Fig.3) dynamics of changes of duration of day separately for 2004 is shown: characteristic fluctuations with the period half year, year, and also month and half-month are allocated.

It proves to be true the carried out spectral analysis, harmonics with the periods 13,6 are allocated; 28; 9,1; 61 day. Change of duration of day with the period of less then a month is caused by tidal fluctuations [2]. Other allocated harmonics are caused by seasonal changes.

As a result of seasonal changes in speed of rotation of the Earth duration of day within one year can differ from its average duration for a year on ± 0,001. Thus the shortest day fall to July - August, and the longest - for March. The most probable reason of periodic changes of speed is seasonal redistributions of air and water weights on the surface of the Earth [3].

The conclusion: As a result of processing of some changes of the duration of day the model describing changes of duration of terrestrial day is allocated optimum by criterion of minimum AQD and $\sigma_D$. Dynamics of changes of the duration of day for 2004 is analysed.

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SEASONAL VARIATIONS OF THE BOUND WATER CONTENT ON THE MARTIAN SURFACE: GLOBAL MAPPING OF THE 6.1 µm EMISSIVITY BAND BASED ON TES DATA. R. O. Kuzmin1, P. R. Christensen2, M. Yu. Zolotov2, and S. Anwar2. 1Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, 19 Kosygin str., Moscow 119991, Russia, e-mail: roki@geokhi.ru, 2Department of Geological Sciences, Arizona State University, Tempe, AZ 85287.

Introduction: Earth-based spectral observations of Mars [1,2,3,4,8] and results from orbital and landing missions [5,6,7,9,10] indicate the abundant presence H2O, OH-bearing minerals (ferric hydroxides and oxyhydroxides, sulfates, chlorides and phyllosilicates). The salts content in the Martian regolith may vary within 8-25 vol. % [9] and the dominant salts are likely to be presented by Mg and Ca sulfates and chlorides. Last discovery from MER-A and B missions have shown that sulfates content in the Martian rocks varies in the range 15-40 wt.% [11,12]. The mid-IR spectral region of the most hydrated minerals includes distinctive emission peak (near 6.1 µm) attributed to the H-O-H bending fundamental vibration of bound water [13]. The bound water (BW) emission peak is notably observed in ground-based thermal emission spectrum of Mars [4], in TES (on Mars Global Surveyor) spectra [14,15] and in Mini-TES spectra at the landing sites of the MER rovers [16]. This peak serves as a direct indicator for the hydrated minerals in the Martian regolith. Preliminary mapping results of the 6.1 µm emission peak (based on the one Martian year TES dataset) have shown the bound water (BW) content in surface material varies seasonally at the time scale of the half Martian year (spring-summer, fall-winter) and character of its distribution in the northern and southern hemispheres is a notably different [17]. Here we report much more detailed results of the seasonal BW content distribution on Mars, based on the complete dataset of the TES observations accumulated during three Martian years at two spectral resolutions (10 cm⁻¹ and 5 cm⁻¹).

Seasonal variations of the BW content: In addition to our earlier mapping of the BW index we have conducted global mapping of the parameter with more frequent Ls time steps (from 30° to 90°), that allow studying the seasonal variations of the BW content in more details. To eliminate possible influence of the atmospheric water bend (~ 6.3 µm), which is located at lower frequency than the BW emission peak, we made a little modification of the BW index: it has been created using a ratio of the emission maximum near 6.1 µm (averaged value from TES channels 140-142) to averaged value of the emission minimums points (channels 135, 134 and 145, 139) located on each sides from the emission maxima of both the atmospheric water vapor and BW. The examples of the global BW content maps for the different seasons (with Ls time step 60°) and the zonally averaged meridian distribution of the BW content for each of the Martian seasons are shown on the Figures-1 and 2 respectively.

During each seasons, the BW index has strong latitudinal distribution. The highest BW values are located mostly within the peripheral zones of the perennial and seasonal polar caps (cooler areas), while lower BW values are observed at low latitudes (relatively warmer areas). At transition from the North spring (Ls=0°-90°) to the North winter (Ls=270°-360°) the northern maximum of BW values is shifting...

Fig.1. The maps of BW content distribution in different seasons (through the Ls time step 60°). Color bar represents values of the BW index (see text).
from high latitudes to latitudes of 20°-30°N, being mostly disappearing in the period of Ls=130°-230°. At that, during the North spring and summer BW values within low-middle latitudes are four times higher than during the same season in the southern hemisphere. It is notably that the BW index distribution during the North summer is characterized by distinct latitudinal asymmetry (see Fig.2) while during the South summer the asymmetry is much less visible.

Maps of BW can be interpreted in terms of the hydration-dehydration. The maximum of hydration on Mars is taking place in the period of the Northern spring and the first half of summer, while maximum of dehydration is taking place in the period of second half of the Northern summer and fall. Two maximums of the hydration are observed during the Southern summer: one is located at the high latitudes of Southern hemisphere and other is located in the Northern hemisphere within the latitude belt 20°-30°N. The mapping results show that the time scale of the observing transition from hydrated to dehydrated states of surface materials (depending on seasons) corresponds to the Ls range from 10° to 40° (from ~1 to 3 months).

Discussion: Conducted analysis of the seasonal dynamics of the BW content distribution (Fig.3) shows that seasonal regime of the BW index has strong sensitivity to the seasonal course of surface temperature and humidity in the boundary atmospheric layer. For example, the mapped seasonal maximums of the BW content on Mars are correlated with the seasonal maximums of the atmospheric water vapor abundance observed with TES [18] at relatively low temperature.

In addition, the process of minerals hydration on Mars reveals much actively during the aphelion period. Later the atmosphere is warming quite quickly at very abrupt decreasing of the atmospheric water vapor abundance. This decreases relative humidity within boundary layer of the atmosphere. In that time minerals dehydration process on the Martian surface is dominant (see fig.3). In the season of the Southern spring-summer (Ls=240°-330°), new activation of minerals hydration arise at the Southern high latitudes and at the Northern low-middle latitudes due to appearance of two atmospheric water vapor maximums. One of them is associated with sublimation of the perennial southern polar cap and other is related to the active south-north transportation of the atmospheric water to the latitudes via the Hadley cell circulation [19]. Therefore, in the season the relative humidity may become higher due to both increased abundance of the atmospheric water vapor and moderately low surface temperatures resulted in higher dust opacity of the atmosphere, which is typical for the season [18].

Our mapping indicates that seasonal regime of the BW content distribution may serve as qualitative spectral indicator of degree and rates of mineral hydration/dehydration processes on Mars. Since the rates of hydration/dehydration processes strongly vary for different minerals, the observing features of the BW seasonal regime can be used to constrain surface minerals.

DEVELOPMENT OF AN UNUSUAL CRATER IN ARABIA TERRA, MARS

H. Lahtela¹,², C. Popa², J. Korteniemi¹, G. Di Achille², G. G. Ori², G. Neukum³ and the HRSC Co-Investigator Team. ¹Division of Astronomy, Dept. of Physical Sciences, P.O.Box 3000, 90014 University of Oulu, Finland (hlahtela@student.oulu.fi), ²IRSPS, Università d’Annunzio, Pescara, Italy, ³Institute of Geosciences, Freie Universität Berlin, Germany.

Introduction: This still undergoing study discusses the characteristics and evolution of a crater located near the dichotomy boundary in Arabia Terra (36.0ºN/351.8ºE). The crater has a diameter of roughly 25 km and it has undergone several different evolitional phases, which have shaped its appearance (Fig. 1A). Three distinctive different terrain units characterize its floor; relatively smooth terrain, surface cracked by severe fissuring and a low albedo depression with a central bulge (Fig. 1B). Additionally, there are two fluvial channels entering the crater. Together all these units tell us about intense crater floor deformation. We suggest that this has been done by a mixture of aeolian, fluvial with possibly volcanic and lacustrine processes.

The Surroundings: The crater is located near the dichotomy boundary, in an area characterized both by the northern lowlands and the southern highlands. The origin of the boundary is still unknown, but various features speak on behalf of combined fluvial and volcanic deformation of the area [e.g. 1, 2]. There are numerous studies on the dissected crater-like features near the boundary [e.g. 1, 3], often associated with volcanism and hydrothermal environments inside craters. In the regional geologic map [2], made using Mariner 9 and Viking images, dark material inside this crater as well as in other locations was interpreted to be of aeolian origin. Same conclusion has been made also after studying the latest datasets [e.g. 4, 5, 6]. The crater itself is of Noachian age, and degraded, with no apparent rim or traceable ejecta field remnants.

The Water Environment: The crater’s two inflow channels are quite small. Their short and stubby appearances indicate forming by sapping processes. However, there is a fresh impact crater (dash line in Fig. 1A), which superposes the eastern channel. Its ejecta blanket may have covered the channel’s connection to a much larger channel just a few km away. (black arrows in Fig. 1A show connections with possibly buried channel). However the amount of water entering the crater is still unclear and no crucial evidence of past lacustrine environment has been found.

Smooth floor: The western part of the crater is characterized by rather smooth and ‘normal looking’ floor material, i.e. sedimentary materials accumulated and partly eroded after the crater formation. Just next to the rims there are indications of mass wasting deposits on the floor surface.

Cracked terrain and the eastern depression: The inner crater floor has undergone intense fracturing (Fig. 1C). Especially the western part of the

Figure 1: (A) Studied crater (A h1205) in Arabia Terra (HRSC orbit h1205 false-color image). Several processes have contributed to its evolution result is a crater with two inlet channels, low-albedo bulge (arrows 1&2, Fig. 1b) and partly dissect crater floor (rectangle in 1A, Fig. 1c). (B) A 3-D image of the dark bulge seen in the S-E part of the crater floor, clearly shows that it’s composed of two separate peaks (arrows 1&2). The view is towards south (arrow B IN Fig. 1a). (C) The depression surrounding the bulge ends to partly fractured terrain (MOC M0403474). (D) Close-up shows layering (arrows) on the chaos remnants; the crater floor is apparently composed of layered sediments; layered sediments comprise the crater floor.
cracked terrain has a chaotic appearance, indicating collapse by very sudden removal of material from the subsurface. The central-eastern crater floor is about 50 m lower in elevation and significantly lower albedo than the smooth terrain. The contacts with the smooth crater floor reveal distinct layering (Fig. 1D).

The Bulge: The central area of the depression exhibits two separate prominent bulges. The western (Fig 1B, arrow 1) is clearly more massive: it rises 300 m above the depressed area floor and has a total volume of 6 km³, while the smaller (Fig 1B, arrow 2) is 70 m high with a volume of 400 m³. Even though the bulges lie in the depression, they clearly reach higher than all the rest of the crater floor.

The surface material of the bulge and the surrounding depression differs from surrounding materials both in composition and morphology. They have a very low albedo. The darkest part lies in the S-E part of the depression, from where the color diffuses outwards into brighter tones. The dark deposits on Mars are often connected with mafic/ultramafic dust/sand, most probably originating from volcanic sources [e.g. 7,8,9 and the references therein]. If the albedo would be caused by aeolian accumulation of this dust, it should follow the topography and cover mostly the lower parts of the crater. Instead, it covers the huge bulge with ease, indicating that it is closely linked to the characteristics of the bulge itself.

One hypothesis is that the bulge would be the central peak of the crater. However, for that purpose it is extremely large and situated quite far from the crater center for this to be a reasonable explanation.

Instead, we suggest that the bulge is of endogenic origin, built up by a volcanic plume coming near the surface. Volcanic extrusion would also help to understand the intense dissecting of the rest of the crater floor [see e.g. 1, 3, 10]. There have been also other studies from the near by areas concluding that there has been volcanic activity which have built small-scale edifices [e.g. 11].

Mineralogy: A mineralogical survey of the studied area will discriminate between the different units in the crater. Currently, three IR datasets are available: 1) multispectral THEMIS high-res thermal infrared (TIR) (resolution ~100 m/pixel), 2) hyperspectral low-res TES dataset (resolution 3x8 km/pixel; covering TIR part of spectra; suitable for mafic silicate mineral identification) and 3) hyperspectral OMEGA dataset (300 m/pixel at periapsis, covering the near-IR part of the spectrum; suitable for evaporite mineral identification).

Unfortunately the datasets have some shortcomings. TES data has a too small pixel resolution, and it is thus not able to solve small-scale features. Additionally, the OMEGA data covering the target area has not been released to the public yet.

While examining TES spectra, only the usual clay mineral signatures could be proven. The presence of montmorillonite fit rather well to TES spectra. The results from current spectral datasets do not help solving such small-scale features (Fig. 2).

Conclusions: The hypothesis that the bulge is of volcanic origin, related to the endogenic activity of the dichotomy boundary area, seems plausible at this stage of studies. The fractured texture was probably formed due to the rising of the bulge and/or by the same forces, which caused it. There may have been volatile-rich sediments on the crater floor brought by the fluvial channels. Their sudden removal by heating and subsequent evaporation would easily account for the chaotic nature of the cracks.

Future studies: This study is still in its early phases. The major guidelines for the future research are more detailed analysis of the crater’s characteristics with wider areal mapping to identify which of them are due to local phenomena and which part of large-scale the regional evolution. In addition, future higher resolution spectral image systems such as the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) onboard the Mars Reconnaissance Orbiter (MRO) could solve the small-scale mineralogical characteristics of the terrain units and confirm or infirm out hypotheses.

Acknowledgements: We would like to recognize Vilho, Yrjö and Kalle Väisälä Foundation for funding and IRSPS for facilities for the research.


Introduction. Among known planetary materials, enstatite (or E-) chondrites are truly a breed apart. They are highly reduced, with < 1 mol percent FeO in their silicates, in contrast to other chondrites as well as Earth, Mars, and Venus, which have FeO contents in the 10-40 % range [1]. And they contain many unique minerals [2] that imply highly reducing, anhydrous conditions. Like ordinary chondrites, they are classified into petrologic types 3 to 6 (or 7). However, Zhang et al [3] showed that the petrologic types of enstatite chondrites are not always consistent with the geothermometry or mineral chemistry is related to the conditions during brecciation after metamorphism to have determined petrologic types. On the other hand, some melt rocks or melt breccias have recently been discovered in E chondrites [4,5,6]. Abee are characterized as impact melt with ghosts of chondrule-bearing clasts [5]. Its internal structure is a myriad of granulated, metal-rimmed, varying-sized clasts embedded in a dark gray, fine-grained groundmass. Since its initial description [7] Abee has been in subject of more than 31 studies focusing on brecciation, diamonds reportedly of solar nebula origin [8], the oxygen depleted environment where it formed [9] and many parent body studies.

Samples and method. In the present paper the results of elemental abundances in separated grain-sized magnetic and nonmagnetic fractions from Abee 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 subtraction of the matrix element backgrounds [10]. The tables show the average element enrichment factors relative to C1 (11).

Results and discussion. Of 10 grain-sized fractions of Abee EH4 analyzed for siderophile elements, 3 magnetic (metal, schreibersite) fractions have ratios \([(Fe/Ni)A/(Fe/Ni)C1]=0.7\) (mean) less than cosmic and nonmagnetic (sulfides, silicates) fractions – 2.1 (mean) greater then cosmic. This fact supports the opinion that the main process controlling of the composition magnetic phase was sulfurization of metal in protoplanetary nebula. The Abee enstatite chondrite show a typically igneous siderophile element pattern with Ir more depleted than Au and Ni (magnetic fractions- Ir (2.5 -3.0 xC1), Au (5.0 – 6.8 x C1), Ni (4.5 – 5.2 x C1); nonmagnetic fractions Ir (0.08 – 0.2 xC1), Au (0.2 -5.4 xC1), Ni (0.1 – 3.7 x C1). Rare earth elements (REE) measurements in Abee show that all fractions with positive and negative Eu-anomalies are deficient in light REE [Lu (A)/Lu (C1) / La (A) / La (C1)] mean = 2.5 (magnetic) and 1.6 (nonmagnetic fractions). Neither the Eu anomaly nor the light REE depletion can readily explained by nebular condensation at least in solar gas [12]. Perhaps the positive and the negative Eu – anomalies in grain-sized fractions REE patterns are associated with oldhamite. The model of formation of oldhamite, the most REE-enriched phase in EH chondrites, is disputed. Detailed ion probe analyses of individual oldhamite grains revealed various REE patterns, indicating mixing and equilibration of oldhamite
precursors [13]. The Abee EH4 is characterized by enrichment of Zn (1.1 x C1), Na (1.6 x C1) and Ca (1.1 x C1) in fine-grained nonmagnetic fraction (<45 µ). Perhaps, these fraction has niningerite and richterite. The presence in Abee normal zoning in niningerite [14] is due to fast A comparative study cooling in the solar nebula or in the parent body or both.

**Conclusion.** From observed differences of compositions of magnetic and nonmagnetic fractions it follows that our trace element data accord with idea that Abee EH4 reflect main process – sulfurization of metal in protoplanetary nebula and, perhaps, that it may have undergone an igneous event. Abee’s brecciated structure [4,5,6] is a vivid representation of a violent and complex sequence of impacts – large angular clasts of partly melted material with igneous oldhamite-rich dark inclusions, all embedded within a previously melted, but similar, groundmass.


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

<table>
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<tr>
<th>Fractions (µ)</th>
<th>Na</th>
<th>Ca</th>
<th>Sc</th>
<th>Cr</th>
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<th>Ni</th>
<th>Co</th>
<th>Au</th>
<th>Ir</th>
<th>Zn</th>
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<td>&lt;0.3</td>
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Table 2. The average element enrichment factors of nonmagnetic fractions of Abee enstatite meteorite.

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Dissipative Collisions of Asteroid-sized Bodies. M.P. Lazarev, A.V. Vityazev, Institute for Dynamics of Geospheres RAS, 38 Leninsky prosp. (bldg. 1), 119334 Moscow, Russia, avit@idg.chph.ras.ru

Formulation of the problem. Inelastic collisions of bodies which change parameters of orbital motion and internal conditions have been investigated during a long time in dynamic and evolution models of planetary system [1-3]. Recently a considerable attention has been focused on numerical calculations of collisions of small bodies (impactors) with large one (target) for purposes of investigation of cratering processes: fragmentation, heating, melting and so on [4]. It is common to assume in these calculations that all kinetic energy of collision dissipated. In contrast in some papers in planetary cosmogony (see, for example, [5]) it was assumed that all impact energy goes to the change of the energy of orbital motion. The numerous results of well known calculations of the Moon origin model in consequence of the impact of Mars-size bodies to protoEarth are unsuitable for rapid estimation of the importance of a dissipative factor in other collisions. Here simple method and analytic formulae for estimation of influence of dissipative factor in collisions of bodies are present.

Methodology. Let us consider the case when mass of impactor $M^*$, moved with velocity $V^*$, is many fever than target mass $M$. For simplicity assume that larger body before the central impact move in a circular orbit with keplerian velocity $V = \left(\frac{GM^*}{R}\right)^{1/2}$, where $M$ – mass (third) central body (Sun or some planet) around which $M^*$ and $M$ in the same plane are circulated. Recall that for the Earth $V_\oplus \cong 30 \text{ km }/\text{s}$. In noninertial system of the coordinates with origin in centre of $M$ from equations of the momentum conservation we can obtain

$$\Delta V = V^* \alpha (1-\alpha),$$

$\alpha = M^*/M << 1.$

After the collision the quantities characterized the energy of body $M$, their orbital momentum $L$, radial $v_r$ and tangent $v_\phi$ velocities are obtained the small increments. For orbital energy we have $E = E_0 + \mu \Delta E$, where $\mu$, smaller than 1, implies that a part of orbital momentum in a system of collided bodies going to spinning of $M$. Let assume, that $V$ vastly more than escape velocity from $M$ (actual $V^*/V \sim 0.1$). Then we make use the solution of two-body problem (see, for example, [6]).

Results and Discussion. From the integrals of the two-body problem we can obtain for example the expression for eccentricity of new orbit of body $M$ after collision in radial direction:

$$e = \Delta e = \sqrt{\mu V^* \alpha V}.$$  (#)

Firstly we emphasize that increment of eccentricity $\Delta e \propto \mu^{1/2}$. If on the crater formation goes the sizable part ($\mu = 0.1 \div 0.01$) of kinetic energy we must taken it into account for estimations of $\Delta e$. In [5] where the (#) without $\mu$ was obtained the result for the consequences of impact of Mars-size body with protoEarth can been wrong. From (#) one can see that the some anxiety about a dramatic dynamic consequences «deep impact» mission to Tempel 1 (here $\alpha < 10^{-12}$) were also wrong. Some details of calculations and other formulae would be presented on the conference.

Abstract: The analysis of the simplest model shows that an asteroid – rubble pile evolves, depending on the parameter $V^2d$ (where $V$ is the average velocity of fragments and $d$ is the average distance between fragments), either as a conglomerate of "independent mutually gravitating clusters" (when $V^2d < fm$, where $f$ is the gravitational constant and $m$ is the average mass of a fragment) or as a "receding cluster" (when $V^2d > fm$). In the latter case the recession energy is drawn from the gravitational energy of the cluster. Within the framework of the model considered, the characteristic consolidation time in the first ("elliptical") case is estimated to be within $\sim$ ten million years; in the second ("hyperbolic") case, the doubling time for the average distance between the asteroid fragments lies within the limits of several hundred thousand to several million years. It should be noted that the actual consolidation time in the first case may be considerably smaller due to the presence of diffuse matter increasing kinetic energy loss. In the second case, the presence of diffuse matter will result in accelerated exchange of gravitational and kinetic energies and consequently in accelerated "recession" of the cluster of fragments. Thus the mechanism considered enables an asteroid – rubble pile to survive for a long time, and on the other hand, even without tidal effects, it prevents the transformation of the whole Asteroid belt into a structureless "cloud".

Joseph S. Levy1, James W. Head, III1, David R. Marchant2; 1Brown University, Department of Geological Sciences, Providence, RI, USA, 2Boston University, Department of Earth Sciences, Boston, MA, USA

To understand the origin and evolution of polygonally patterned ground on a debris-covered glacier in the Antarctic Dry Valleys, we map the distribution of five morphological units of polygonally patterned ground in Mullins and central Beacon. Mapping is based on surface morphology and morphometric parameters such as trough intersection angle. Where Mullins Valley debouches into Beacon Valley (Figure 1), polygonal patterning transitions from radial orthogonal-intersections to non-oriented hexagonal-intersections, providing a time-series of polygon evolution under uniform microclimate conditions [1]. Based on surface morphology, ice flow rates, cosmogenic nuclide exposure ages, and stress analyses, we conclude that the most likely explanation for the variety of polygon morphologies observed in Mullins and central Beacon Valleys is a model which combines initial thermal contraction-crack formation oriented radially to glacier flow in areas of rapid horizontal ice flow, modified over ky-timescales by sublimation and deformation, and ultimately dominated by non-oriented thermal stresses in areas of stagnant buried ice.

Mullins Valley (77°54'S, 160°35'E) is a tributary valley to Beacon Valley (77°49'S, 160°39'E) and is located in the stable upland zone of the McMurdo Dry Valleys [2]. Climatically, Mullins Valley is hyper-arid, with an annual water-equivalent precipitation of < 10 mm/year (Marchant and Head 2004). Air temperatures range between -48°C and approximately 0°C in adjacent Beacon Valley, with brief hour- to day-scale excursions above 0°C, and a mean annual temperature of -22°C [9].

The floor of Mullins Valley is dominated by Mullins glacier, a debris-covered glacier approximately 3 km long and 0.8 km wide. The glacier is composed of several distinct lobes which are nested axially down-valley. We map five distinct morphological units on Mullins glacier: (Zone 1) nascent, radial crack formation (RC), (Zone 2) radial polygons (RP), (Zone 3) deformed radial polygons (DRP), (Zone 4) non-oriented (“hexagonal”) polygons (NOP), and (Zone 5) no polygons (NP) (Figure 1). For a complete discussion and description of morphological units see Levy et al. (2005, in review). Our mapping program, based on field excavation and mapping,
as well as analysis of high-resolution air-photography, modifies the traditional theory of polygonal patterning—classically described by [5] and [6] as roughly hexagonal networks of thermal contraction cracks. In contrast, our model integrates the morphological characteristics of each zone of polygonal patterning to generate a model which traces the development of polygon morphology as influenced by stress associated with glacier flow, sublimation, and thermal contraction.

Additionally, ice ages give maximum constraints on the time required to produce the observed morphological units. We correlate synthetic aperture radar interferograms from Rignot et al. (2002) with our polygon morphology map of Mullins and central Beacon Valleys. We find that there are distinct spatial relationships between relatively “fast” surface velocities (≥ 5 mm/year) and the formation of radial polygons as well as between slow/stagnant flow rates and areas showing hexagonal non-oriented polygons.

Cosmogenic nuclide exposure ages for the Mullins Valley glacier are reported in clusters of 13 ka, 136 ka, 300 ka, and 730 ka (Marchant, personal communication), which correlate well with polygon morphology and interferogram analysis. The youngest exposure age (13 ka) was measured near the contact between Zones 2 and 3. 136 ka cosmogenic nuclide samples were collected from the Zone 2, near contacts with Zone 5. The 300 ka samples were collected in the Zone 3. The oldest cosmogenic nuclide samples (730 ka) were measured deep in the zone of deformed radial polygons. These observations are consistent with rapidly flowing ice in the up-valley reaches of Mullins Valley and a significant stagnation or slowing of ice as it enters Beacon Valley.

We present the following conceptual model to account for the morphological, age, and flow data observed in Mullins and Beacon Valleys. The troughs of oriented, orthogonal-intersection polygons, which initially arise from cracks that form through a combination of flow-oriented glacial stress and thermal cracking, are modified by sublimation of near-surface ice during down-valley transport. Some polygon troughs re-seal by the time of arrival in Beacon Valley, particularly by slumping of till, resulting in the removal of inherited sites of material weakness and the exposure of new sites for preferred cracking. The near-stagnant ice in Beacon Valley is largely subjected to thermal stress, in the absence of glacial stress, which is adequate to produce contraction cracks that produce non-oriented hexagonal.

This work enhances our ability to interpret polygonal patterning ubiquitously observed on Mars at high latitudes and interpreted as indicators of thermal contraction phenomena (e.g., Mangold et al., 2004; Mellon, 1997). By unraveling the morphological effects of thermal and structural stress in a competent icy layer, we can better understand climate and deformation history recorded in martian polygonally patterned terrain.

References
MODELING OF TURBULENT ACCRETION SATURN'S SUBNEBULA AND FORMATION OF SATELLITES. A. B. Makalkin, V. A. Dorofeeva, and E. L. Ruskol, 1Institute of Physics of the Earth Russian Acad. of Sci., B. Gruzinskaya, 10, 123995 Moscow, Russia, e-mail: makalkin@ifz.ru, 2Vernadsky Institute of Geochemistry and Analytical Chemistry Russia Russian Acad. of Sci., 19, Kosygin Str, 119991 Moscow, Russia.

Introduction: Cassini-Huygens observations of Saturn system, including Titan, provide constraints on models of formation of Saturn and its regular satellites. In this work most of our attention is concentrated on the construction of model of the circumplanetary protosatellite disk of Saturn and consideration of the satellite formation in the disk. This work develops our preceding studies concerning conditions of formation of Galilean satellites in the Jovian subnebula [1, 2, and 3]. Recent models on the Saturn’s subnebula [4] and on Jovian subnebula [5] consider constraints on volatile abundance in the regular satellites of Jupiter and Saturn.

Model and results: A two-dimensional numerical model of the protosatellite disk of Saturn (Saturn’s subnebula) has been constructed. The model includes accretion of the gaseous and solid material from the surrounding region of the solar nebula on the Saturn’s subnebula and accretion from the subnebula onto Saturn. The above accretion processes suggest the subnebula to be turbulent and heated by dissipation of turbulence in addition to radiation of young Saturn. It is shown that the main mechanism of energy transport in the subnebula is radiation. The opacity of the gas-dust medium of the subnebula is defined by composition of the solid component. We take into account temperature dependence of opacity below the water ice condensation temperature as well as at higher temperatures where organic compounds are responsible for opacity. Various degrees of gas depletion (or enrichment of solids) in the subnebula relative to the solar composition are considered in with regard to the observational constraints including the C/H ratio measured by the CIRS instrument aboard Cassini [6] new C/H data by Cassini. The radial and vertical distributions of temperature, pressure and density are obtained and the position of the water-ice condensation front is calculated. The model satisfies the restrictions from the mass of Titan and the present-day data on its chemical composition. The temperatures in the outer part of the subnebula (beginning from the Titan distance) are sufficiently low for stability of clathrate hydrates while their formation in the surrounding region of the solar nebula is more probable.

The rate of replenishment of protosatellite disks of Jupiter and Saturn by solid material due to the capture of planetesimals from the surrounding region of the solar nebula is evaluated. It is shown that in cosmogonically short time intervals $10^4–10^5$ yr several large (10-100 km) bodies could be captured into the subnebula by two-body collisions, independently on gas inflow. The bodies play the role of seeds (embryos) in the process of satellite formation, which proceeds by accretion of subnebula’s solid material onto these seeds. The main income of solid matter into the subnebulae from the feeding zones of the planets presumably resulted from the capture of dust particles and minor planetesimals (~20 m and less) through gas drag. We conclude that regular satellites obtained most of their material from the subnebulae, but satellite formation is impossible without capture of the seed planetesimals.


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**Introduction:** Lunar rings have inner nature.

Old arcs of the Moon and young arcs of the Earth we can see on f.1

Old arcs of the Moon and old arcs of the Earth we can see on f.2, where is African structures [1].

Zero – meridians on the maps f. and f. 2 (lunar and earthen 0) in one position.

**Resume.** The matrix of earthen tectonic forms is drown on the Moon.

Earthen (lines ) and lunar (points) arcs,

COPRATES CHASMA NORTH WALL INTERIOR LAYERED DEPOSIT: LAYER MEASUREMENTS AND COMPARISON WITH JUVENTAE CHASMA ILDS USING MARS EXPRESS HIGH RESOLUTION STEREO CAMERA (HRSC) DERIVED TOPOGRAPHY

Michael, G. 1, E. Hauber 1, K. Gwinner 1, R. Stesky 2, F. Fueten 3, D. Reiss 1, H. Hoffmann 1, R. Jaumann 1, G. Neukum 4, T. Zegers 5, and the HRSC Co-Investigator Team

1Institute of Planetary Research, German Aerospace Center (DLR), Berlin, Germany
2Pangaea Scientific, Brockville, Ontario, Canada
3Department of Earth Sciences, Brock University, St. Catharines, Ontario, Canada
4Remote Sensing of the Earth and Planets, Freie Universitaet, Berlin, Germany
5ESTEC, ESA, Noordwijk, The Netherlands

Introduction: The interior layered deposit (ILD) in the north wall of Coprates Chasma differs from those we have examined in Hebes, Ophir, Candor, Melas, and Juventae Chasmata [1] in that its base apparently occurs some 2700 m above the main chasma floor. Situated in a re-entrant of the chasma wall, it has two main sections, the maximum lateral extent of the larger deposit being about 15 km. Nevertheless, it displays several small-scale features in common with an apparently exhumed ILD in Juventae Chasma (HRSC orbit 1070) of much greater extent (50-150 km, also in two sections): bright material; regular layering with 3-6 m layer thickness; steeper slopes with chutes headed by scalloped alcoves, and ending with dark talus fans; morphologically similar yardangs. Using a digital terrain model derived from HRSC stereo together with projected nadir and colour channels and manually coregistered MOC images, we investigate the structure of the layering, and the present erosional features. As elsewhere, we are interested to make measurements of the strike and dip of the layering, since this information could potentially discriminate between theories of the origin of the layers: whether water emplaced sediments, volcanic ash deposits, or some other airfall deposit.

Coprates Chasma north wall ILD (orbit 515):
Small compared to the other ILDs, about 15 km across, the deposit base is around 2700m above the Coprates floor, and has a height of ~3000m. It shows two apparent scales of layering: one producing a step-like relief, the other much finer. Measurements of the fine layers give mean thickness values of 3.2 m (Profile 1: 30 layers in 205m over 1.5km baseline, registered MOC, only dark to light transitions counted) and 5.7 m (Profile 2: 45 layers in 255m over 1.6km baseline). The coarser layers are measured to have a mean thickness of 85m (Profile 3: 14 layers in 1200m over 3 km baseline). Seen in perspective view with MOC overlays, it would appear that the coarse layering is not horizontal, but dips outward from the centre of the ILD. Wind eroded yardangs are seen between two main sections in apparently similar material. Talus deposits originate particularly from one height level in the main section. Some appear dark and fresh; lighter ones more eroded. Their formation leaves distinctive alcoves at the head of a chute. At the south-west of the ILD there is a scarp of 60º inclination with many such chutes, and morphologically similar to that seen in the Juventae North ILD. The lesser section of the ILD is confined by a ridge of the chasma wall.

Juventae ILD N1 (orbit 1070/1):
This ILD is 70 km north-south in extent, with a wind-eroded top-surface. The yardangs show fine layering, with a typical mean thickness of 2.7m (Profile 1: 60 layers in 160m over 2 km baseline) and 6.5m (Profile 2: 35 layers in 230m over 2 km baseline). There are chutes
(horizontally, up to 1 km) to talus fans headed by distinct scalloped alcoves, originating in particular from at least two specific height levels. There is a strongly degraded concavity on the west side, where significantly more talus is seen, and the chutes are deeper. The dune material to this side of the ILD is particularly abundant. Traceable layers can be observed somewhat outside bright mound of the ILD to the west, over 8 km distance. These are interrupted by a 500 m crater. There is a layer of a different type from the fine layers to the south of the ILD, extending over more than 10 km. There is also a talus flow (1.5 km) from a smooth top-surface, but with no alcoves.

Regular fine layering suggests cyclic formation process.

We hope to get more insight into the origin of the ILDs in the future with the use of results from the OMEGA instrument [2].

**References:**


Disturbances in the Earth’s ionosphere and magnetosphere caused by impacts of asteroids and comets are studied. The 2D hydrodynamic numerical simulations of a cosmic body passage through the atmosphere with allowance for deceleration, deformation and disruption due to aerodynamic loading and formation of the wake behind the body are performed. A plume (a mixture of the air and the products of “explosion” after the impact onto the land or into the ocean) is formed. Rising plume reaches high altitudes, operating as a MHD generator. Field-aligned currents heat the lower layers of the ionosphere and change their conductivity. A part of the plume moves at higher than escape velocity and may pierce the ionosphere and magnetosphere.

For a 1-km body the energy of the high-velocity part of the plume is comparable to that of the Earth’s magnetic field (~200 Mt TNT). The magnetic field cannot stop the plume. The magnetosphere is severely distorted, Van Allen belts disrupted.

For a 30-60 m cosmic body (Tunguska-like object) the plume rises up to a height about 400-500 km and falls back due to gravity, heating the atmosphere at altitudes above ~100 km, and change conductivity. The magnetic storm lasting several hours with amplitude of about 70 nT was observed in Irkutsk (at the distance of about 900 km from the impact point). Impacts of bodies with sizes of 1-2 km are rather rare events. Consequences of the Tunguska-live impacts are not very severe, at least at the regional scale. So we studied impacts of small cosmic bodies with sizes from ~0.1 km up to 1 km.

The MHD numerical simulation of the motion of the plume and its interaction with the geomagnetic field are performed. Excitation of MHD waves is demonstrated with amplitudes of $10^2$-$10^3$ nT. These disturbances are capable of triggering precipitation of particles from Van Allen radiation belts, increase ionization at lower altitudes, produce intense electromagnetic noise.

Lower parts of the plume increase the air density at large distances from the impact point. For an example, a 400-m stony body, with the initial velocity of 17 km/s impacting the ocean for the moments of 200-600 s increase the density in the cylinder with radius of 1600-1800 km and altitude of ~2000 km up to the density of $10^{-14}$ g/cm$^3$, which corresponds to the normal density at the altitude of ~350 km. That changes the recombination rates and increases the ionization due to solar radiation and cosmic rays. This effect resembles rising the F2 layer of the ionosphere up to altitudes of 1000-2000 km. Radii of the density disturbances at higher altitudes are much larger – they reach 5000-7000 km at an altitude of 1200 km at the same moments of time.

Large mass of the plume falls back due to gravity and produced intense oscillations of the ionospheric conducting layers propagating to very large distances from the impact point. So the disturbances may have global character.

We note that recently found asteroid 2004 MN4 has the size of about 300-400 m. It will pass the Earth in 2029 at a rather small distance – about 36000 km. There is some possibility that trajectory of the asteroid will change and it may hit the Earth in 2035-2039, most probably into the ocean. Its energy is about $3 \times 10^4$ Mt TNT. Tsunami created by the impact will provide devastating effects. But in addition severe magnetospheric and ionospheric disturbances may make the normal work of some technical systems of the modern civilization not possible: they will disrupt radio communications, hinder TV broadcasts and radiolocation and produce great errors in the location by GPS system and so on.

Determination of the time necessary to restore normal state of the ionosphere and magnetosphere is the goal of the future research.
THE EARLIEST NAMES ON TITAN: NOMENCLATURE SYSTEM FOR ONE MORE WORLD.

T. C. Owen1, K. Aksnes2, R. Beebe3, J. Blue4, A. Brahic5, G. A. Burba6, B. A. Smith7, V. G. Teifel8

1Institute for Astronomy, 2680 Woodlawn Dr, Honolulu, HI 96822-1897, USA owen@ifa.hawaii.edu.
2Institute of Theoretical Astrophysics, P.B. 1029 Blindern, 0315 Oslo, Norway kaare.aksnes@astro.uio.no.
3New Mexico State University, MSC 4500, P.O. Box 30001, Las Cruces, NM 88003, USA rbeebe@nmsu.edu.
4US Geological Survey, 2255 N Gemini Dr, Flagstaff, AZ 86001, USA jblue@usgs.gov.
5Centre d'Etudes de Saclay, Universite Paris VII, L'Orme des Merisiers, 91191 Gif-sur-Yvette Cedex, France brahic@discovery.saclay.cea.fr.
6Vernadsky Institute, 19 Kosygin St, Moscow 119991, Russia gburba@gmail.com.
7P.O. Box 70, Tesuque, NM 87574-0649, USA brad.smith@starband.net.
8Fessenkov Astrophysical Institute, 480020 Almaty, Kazakhstan tejf@hotmail.com.

Introduction: A rich harvest of new results has been obtained in 2005 by the Cassini-Huygens mission to the Saturnian system, a cooperative project of NASA, the European Space Agency and the Italian Space Agency. One of the main achievements is a unique possibility to obtain the first look onto the surface of Saturn's largest Moon, Titan. The success of the Huygens probe into Titan's thick nitrogen-methane atmosphere has revealed a new world, strangely Earth-like, with methane playing the role of water, low temperature ice substituting for rock, and organic aerosols precipitated from the atmosphere taking the place of soil. Streams of liquid methane course over the icy bedrock of a world nearly frozen in time shortly after its formation.

Titan surface structure: Prior to the Cassini-Huygens mission there was almost no data on the surface of Titan, despite the fact that it is one of the largest moons in the solar system, with a diameter of 5150 km. Titan's atmospheric haze hid the surface from Voyager and Earth-based observations in visible light. The Cassini spacecraft during its flybys close to Titan revealed the surface features with its visual and IR imaging instrument, and with the first strips from an imaging radar survey. The resolution of these images is still coarse to moderate, and Cassini has still seen only a fraction of the surface. As a result the description of the feature types on Titan that we can make today is just as first approach.

The current synoptic map covers about 80% of Titan – segments of the area south of 35°N. Images for this map were obtained in a near-IR band (938 nm) at which light can penetrate Titan's atmosphere. The map shows only brightness variations on Titan's surface (the illumination is such that there are no shadows and no shading from topographic variations). The dark terrains are presumably lowlands. The Huygens probe landed in such a region. The bright regions of Titan are thought to consist of upland terrain that is relatively uncontaminated by the dark material that fills the lowland regions. In the south polar region there is a dark feature with sharp boundaries identified as the best candidate so far for a past or present hydrocarbon lake on Titan. In a bright terrain area just north of the Huygens landing site, there are numerous channels with a dendritic pattern. They have been formed by a running fluid and could carry liquid methane in past and present times.

Creation of the Titan nomenclature: The International Astronomical Union's (IAU) Working Group on Planetary System Nomenclature (WGPSN) established 10 feature types and Latin terms for naming on Titan, as well as 7 categories of proper names. The initial work has been done by the authors of this abstract – the members of the Outer Solar System Task Group, which is a subdivision of the WGPSN, and Jennifer Blue, who is a geographer at the USGS/Flagstaff Astrogeology Research Program. Further development was made by the WGPSN, which approved the names.

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**Feature type definitions:**

*Albedo feature*: Geographic area distinguished by amount of reflected light.

*Arcus*: Arc-shaped feature.

*Crater*: A circular depression.

*Facula*: Bright spot(s).

*Flumen*: Channel(s) on Titan that might carry liquid.

*Lacus*: Small plain; on Titan – a "lake" or small, dark plain with discrete, sharp boundaries.

*Macula*: Dark spot.

*Regio*: A large area marked by reflectivity or color distinctions from adjacent areas, or a broad geographic region.

*Ringed feature*: Cryptic ringed feature.

*Virga*: A streak or stripe of color.

**Proper name categories:**

*Albedo features*: Sacred or enchanted places, paradise, celestial realms from legends, myths, stories, and poems of cultures from around the world.

*Crater and ringed features*: Deities of wisdom.

*Facula*: Islands on the Earth that are not politically independent.

*Flumen*: Mythical or imaginary rivers.

*Lacus*: Lakes on Earth, preferably with a shape similar to the lacus on Titan.

*Other features*: Deities of happiness, peace, and harmony from world cultures.

*Virga*: Deities of rain.
Feature names on Titan: The 43 names listed below can be used in papers with a note “provisional name” until their final official approval by the forthcoming IAU General Assembly (August 2006).

Coordinates are rounded to 1°. Longitudes are W.

ALBEDO FEATURES (13)

**Aaru** 10 N 340 Egyptian abode of the blessed dead.

**Adiri** 10 S 210 Melanesian afterworld where life is easier than on Earth.

**Aztlan** 10 S 20 Mythical land from which the Aztecs believed they migrated.

**Belet** 5 S 255 Malay afterworld reached by a flower-lined bridge.

**Ching-tu** 30 S 20 Chinese Buddhist paradise where those who attain salvation will live in unalloyed happiness.

**Dilmun** 15 N 175 Sumerian garden of paradise, primeval land of bliss.

**Fensal** 5 N 30 Magnificent mansion of the Norse goddess Frigga, to which she invited all married couples who had led virtuous lives on Earth to enjoy each other's company forever.

**Mezzoramia** 70 S 0 Oasis of happiness in the African desert, from an Italian legend.

**Quivira** 0 N/S 15 Legendary city of a fabulous treasure sought by Coronado and other explorers (SW USA).

**Senkyo** 5 S 320 Japanese ideal realm of aloofness and serenity, freedom from worldly cares and death.

**Shangri-la** 10 S 165 Tibetan mythical land of eternal youth.

**Tsegihi** 40 S 10 Navajo sacred place.

**Xanadu** 15 S 100 An imaginary country, in the poem “Kubla Khan” by English author Samuel Coleridge (1772 – 1834).

ARCUS (1)

**Hotei Arcus** 28 S 79 One of the seven gods of happiness in Japanese Buddhism. He is the god of contentment, good fortune, cheerfulness, and he is always smiling.

CRATER (0)

No feature named so far with this term.

FACULA / FACULAE (15)

**Antilia Faculae** 11 S 187 Archipelago corresponding to the mythical island of Antilia, once thought to lie midway between Europe and the Americas.

**Bazaruto Facula** 12 N 16 Mozambique isl.

**Coats Facula** 11 S 29 Canadian island.

**Creté Facula** 9 N 150 Greek island.

**Elba Facula** 11 S 1 Italian island.

**Kerguelen Facula** 5 S 151 French island.

**Mindanao Faculae** 7 S 174 Philippine isl.

**Nicobar Faculae** 2 N 159 Indian archipel.

**Oahu Facula** 5 N 167 Hawaiian island.

**Santorini Facula** 2 N 146 Greek island.

**Shikoku Facula** 10 S 164 Japanese island.

**Sotra Facula** 12 S 40 Norwegian isl.

**Texel Facula** 11 S 183 Dutch island.

**Tortola Facula** 9 N 143 One of the British Virgin Islands.

**Vis Facula** 7 N 138 Croatian island.

FLUMEN / FLUMINA (0)

No feature named so far with this term.

**LACUS** (1)

**Ontario Lacus** 72 S 183 Lake between Canada and USA.

**MACULA** (4)

**Eir Macula** 24 S 115 Norse goddess of healing and peace.

**Elpis Macula** 31 N 27 Greek goddess of happiness and hope.

**Ganesa Macula** 50 N 87 Hindu elephant-headed god of good fortune and wisdom.

**Omacatl Macula** 18 N 37 Aztec god of good cheer and lord of banquets.

REGIO (1)

**Tui Regio** 20 S 130 Chinese goddess of happiness, joy and water.

RINGED FEATURES (3)

**Guabonito** 11 S 151 Taino Indian (Antilles) sea goddess who taught the use of amulets.

**Nath** 30 S 8 Irish goddess of wisdom.

**Veles** 2 N 137 Slavic god of housekeeping wisdom.

**VIRGA** / **VIRGAE** (5)

**Bacab Virgae** 19 S 151 Mayan rain god.

**Hobal Virga** 35 S 166 Arabian rain god.

**Kalsru Virga** 36 S 137 NW Australian rainbow serpent, bringer of rain.

**Perkunas Virgae** 27 S 162 Lithuanian god of rain, thunder and lightning.

**Shiwanni Virgae** 25 S 32 Zuni (SW USA) rain god.

Diameter of ringed features (km):

**Guabonito** 55

**Nath** 95

**Veles** 45

Introduction: Voyager and Galileo images of Europa have shown its surface to be highly deformed by tectonic features. The formation of these features has been attributed to deformation of the outer ice shell from stresses induced by diurnal tides, nonsynchronous rotation, and/or polar wander [1]. The effects of these stress fields (acting independently or in concert) are inferred to fracture the ice shell into plates that subsequently translate and/or rotate with respect to one another (e.g., [2]).

Using Voyager image data, [3] showed that the dark material that composed the interior of some features could be removed, allowing the margins of the features to fit closely together and surrounding preexisting offset lineaments to be reconstructed. The successful reconstruction of these preexisting features indicates that, at the scale of the observations, these plates behaved rigidly during deformation (i.e. deformation was accommodated at the margins of the plates). A number of successful reconstructions involving external extensional and strike-slip features have reinforced this result (e.g., [4-6]) and led to the general assumption that tectonic plates behave rigidly on Europa.

Exceptions to this assumption have been identified by [7] and [8], using tectonic reconstructions of Europa’s surface. The former involved analysis of the extensional band Thynia Linea. A reconstruction of the band, using the displacement azimuths of preexisting features, indicated that displacement magnitude decreased toward either end of the extensional band. It was suggested that this behavior was akin to a “tear” in a nonrigid Europan lithosphere. The latter involved the reconstruction of a set of ridges and a band-like complex surrounding Castalia Macula, using an inverse modeling technique. The application of this technique indicated that deformation associated with the formation and/or modification of these features had been accommodated within one or more tectonic plates in the region. Both reconstructions suggested nonrigid behavior on the kilometer-scale.

The identification of nonrigid behavior associated with tectonism on Europa raises several questions. Are these two regions unique or is nonrigid behavior more common then has been previously recognized? Is there a scale dependence for the occurrence of nonrigid behavior? Can we quantify nonrigidity and how it affects the mechanical behavior of Europa’s lithosphere? To search for answers to these questions, we examine a tensile feature on Europa’s equatorial trailing hemisphere that defines a number of tectonic plates that have translated/rotated with respect to each other (Fig. 1a).

Study Area: The morphology of the feature we analyze and its relationships with the surrounding terrain suggest it is an extensional band [6,9,10]. For the purposes of this analysis, we refer to it hereafter as band A. It trends E-W and diverges into two bands in the western portion of the image (A	extsubscript{1} and A	extsubscript{2} in Fig. 1b). The width of the band ranges from 5-15 km. Bands A	extsubscript{1} and A	extsubscript{2} interact with each other three times within the available image coverage. These interactions separate the region between bands A	extsubscript{1} and A	extsubscript{2} into three plates (Fig. 1b). Plates 2 and 3 have surface areas of ~6000 km	extsuperscript{2} and plate 4 has a surface area of ~1500 km	extsuperscript{2}. The boundaries of plates 1 and 5 extend beyond the image coverage of the region and therefore cannot be defined strictly (Fig. 1).

Only six lineaments post-date the formation of band A (Fig. 1b), indicating that it is a relatively young feature. There are a number of preexisting lineaments that have been offset by the formation of the band (Fig. 1b). These can be used to reconstruct the deformation history of the region during the formation of band A. Assuming that the band formed by extension, our expectation is that the only modification of any reconstructable preexisting lineaments along band A are the offsets created by the addition of new material within the boundaries of the band. This suggests that, in a rigid-plate environment, a reconstruction of any of the plates defined by band A should realign preexisting offset lineaments along the boundaries of the plates with comparable accuracy.

Results: When reconstructing the region modified by the formation of band A	extsubscript{1}, we hold plate 1 fixed and translate plates 2-4 with respect to it (Fig. 1c). The realignment of preexisting offset lineaments along band A	extsubscript{1} in the reconstructions of plates 3 and 4 suggest that these plates behaved rigidly. However, while preexisting features offset by band A	extsubscript{1} along the boundary of plate 2 can be reconstructed such that they have a negligible strike-slip offset, a significant amount of band A	extsubscript{1} remains in the reconstruction (Fig. 1c). This suggests that some component of nonrigid behavior could be associated with plate 1 and/or 2.

For the reconstruction of plate 5, we hold plates 1-4 fixed. Preexisting lineaments offset by band A	extsubscript{2} along the boundary between plates 3 and 4 and plate 5 are realigned with negligible offsets. However, the offsets of preexisting lineaments along the boundary between plate 2 and plate 5 retain a significant strike-slip component. This suggests that some component...
of nonrigid behavior could be associated with plate 2 and/or 5.

These preliminary results suggest that nonrigid behavior of Europa’s lithosphere is not unique to the two instances reported previously [7,8]. They also suggest that the occurrence of nonrigid behavior is not scale dependent (plates 2 and 3 are similar in scale yet one displays evidence for nonrigid behavior and the other does not). Quantifying the amount of nonrigid behavior present here would be difficult using this analysis. However, this region is an excellent candidate for use with the inverse modeling technique developed by [8]. The application of the technique to this region offers a means of quantitatively verifying the occurrence of nonrigid behavior in the region, as well as possibly quantifying the amount of nonrigid behavior if it were present.


Fig. 1 a) Mercator projection of an image mosaic (Europa_pole_2_pole01.cub) produced by the University of Arizona’s Planetary Image Research Laboratory centered at ~0° lat and 227° lon and at a resolution of 220 m/pix. b) Sketch map of the region with the band that is reconstructed shown with a gray fill and plates defined by this feature numbered 1-5. Prominent features that post-date the formation of the band are shown in gray and offset preexisting features are shown in black. c) Reconstruction of the region with previously offset features shown in white.
**GROWTH OF CONDENSATIONS IN THE PREPLANETARY DISK AND THE PLANETS FORMATION.** G.V. Pechernikova and A.V. Vityazev, Institute for Dynamics of Geospheres RAS, 38 Leninsky prosp. (bldg. 1), 119334 Moscow, Russia, pechernikova@idg.chph.ras.ru

**Formulation of the problem.** Two approaches for explanation of planetary systems were developed in the last two centuries. According to the first of them, planets were accumulated from the solid bodies (Chladni, 1794, Multon-Chamberlin, 1905, Schmidt, 1945, Safronov, 1950-80, Wetherill, 1980-1990, et al). According to the second one, planets were originated from large gas-dust condensations (Kant-Laplace, XVIII c., Berlage-Weiszsekker, 1930-50, Eneevo-Kozlov, 1975, et al). In the standard scenario of the Solar System formation it is assumed that terrestrial planets were formed from the solid bodies [1-4] after a short condensation stage, but giant-planets were formed from condensations. Recently [5, 6] some geochemical arguments in support for prolonged stage of condensation for terrestrial planets were proposed. Here we attack such a possibility from a dynamical point of view.

**Model: Growth of mass of the largest condensations in terrestrial zone.** A formal description of evolution of an individual condensation is given in [1, 2]. Here we outline a constitutive part of it. The internal gravitational force of a primary condensation is greater than external forces. Therefore, the condensation starts to contract, until gravitation is balanced by centrifugal force, increasing during contraction. Its velocity of rotation is close to the Keplerian velocity. It has long been noticed that non-central collisions were the only mechanism capable to maintain the planetesimals in rarefied state. The combining of condensations leads to an efficient compression, on the average, irrespective of whether it is a cloud of dust or rotating swarm of debris. Thus, in the case of the combining of two condensations of comparable masses colliding centrally, its mass doubles practically while the angular momentum remains as before: the radius of the condensation decreases twofold while the density increases 16-fold. At tangent collisions the change of angular momentum can even exceed considerably the angular momentum before the collision, maintaining the newly formed condensation in rarefied state, and even leading to the scattering of a substantial part of the matter. Thus, in the condensation system there is a need to describe simultaneously change in number, masses, sizes (densities) and angular moments of colliding, combining and disintegrating condensations in the framework of a self-consisted problem. The description of the set of equations can be found in [2]. Below, a brief account of the corresponding model and results of new calculations are presented.

The mass distribution of preplanetary bodies is usually represented in the form of a simple power law

\[ n(m, t)dm = c(t) m^{-q}dm, \quad 1 < q < 2, \]  

where the exponent \( q \) does not depend on time and is determined by a coagulation equation of the type

\[ \frac{dn(m, t)}{dt} = \int_{m_0}^{m_2} A(m', m - m') n(m', t)n(m - m', t)dm' - \int_{m_0}^M A(m, m') n(m', t)dm'. \]  

Here, \( A(m, m') \) is the coagulation coefficient and \( m_0 \) and \( M \) are the lower and upper limits of the distribution. The coagulation coefficient \( A(m, m') \) is proportional to the frequency of collisions between bodies with masses \( m \) and \( m' \), having a relative velocity \( V(m, m') \) before encounter, and is written as:

\[ A = \pi(r + r')^2 \left[ 1 + \frac{2G(m + m')}{V^2(r + r')} \right] V, \]

\[ V = \sqrt{V^2(m) + V^2(m')}, \]  

where the first three factors determine the cross section for the collision within the framework of the two-body problem. For \( A(m, m') \propto m^\alpha + m'^\alpha \) with \( 0 < \alpha < 2 \) a solution of the equation (2) was obtained as \( q = 1 + \alpha/2 \) [2]. On the assumption that the momentum of a condensation rotation is determined only by its mass, \( K \propto m^4 \), it was shown in [2] that \( q = \gamma \) for the massive condensations, the escape velocity from which exceeds \( V \). The rate of growth of a largest condensation is described by the equation [1, 2]:

\[ \frac{dm}{dt} = \pi l_0^2 \rho_0 v \approx \frac{8\pi}{3} (1 + 2\theta) \zeta r_0^2 \left( \frac{m_0}{m} \right)^{\theta - \gamma} \frac{\sigma_0}{P_k}, \]  

where \( l_0 \) is the impact parameter, \( \rho_0 \) and \( \sigma_0 \) are the density and the surface density of the solid matter in the feeding zone of the planet respectively, \( v = (Gm/\theta r)^{1/2} \) is the average relative velocity of the condensations, \( r_0 \) and \( m_0 \) are their initial radius and mass, \( P_k \) is the period of revolution around the sun, \( \zeta \) is the coefficient of initial contraction of the condensations (\( \zeta \approx 10^{-1} \)), \( \theta \) is a parameter of the order of first units for bodies in the absence of the gas and of the order of ten for bodies moving in the gas (Safronov' parameter) [1, 2].

The density \( \delta \) of the condensations during their collisional evolution increases as

\[ \delta = (m/m_0)^{10 - 6 \gamma} \delta_0 \]  

from the density \( \delta_0 \) of the condensation just after its initial contraction. When \( \delta \) increases up to \( \sim 1 \) g/cm\(^3\) condensations become solid bodies. Denoting \( m/M_0 = z^2 \), where \( M_0 \) is the mass of the Earth, and taking into account that \( \sigma_0(t) \) decreases from its initial value \( \sigma_0 \) due to depletion of the material as

\[ \sigma_0(t) = \sigma_0 \left[ 1 - (m/t_M_0)^{2}\right], \]  

we pass to variable \( z \) in Eqs. (4-6). Then, an increase of \( \delta \) and the time of growth of the condensation from
the mass $m_0$ to $m = z^3 M_0$ are described by the following expressions:

$$\delta = \left(\frac{z}{z_0}\right)^{\delta_0}$$

$$t = \frac{1}{B \delta_0} \int_{z_0}^{z} \left(1 - z^2 \right)^{-\delta} dz$$

$$B = \frac{0.2 \pi (1 + 2 \theta)}{9} \left(\frac{M_0}{m_0}\right)^{\gamma-7}$$

We set $z_0 = (m_0/M_0)^{1/3}$ and $\delta_0 = 10^{-5} \text{g/cm}^3$, $m_0 = 5 \times 10^{16} \text{g}$ for the terrestrial zone. From (7) we find the value $z_{\text{max}}$ of conversion of condensations into solid bodies for the values of $\gamma$ determined for massive condensations. Then, inserting $z_{\text{max}}$ in Eq. (8), we estimate the time $t_{\text{max}}$ for corresponding increases of mass from $m_0$ to $m_{\text{max}} = z_{\text{max}}^3 M_0$.

**Results.** In Figure and Table we illustrate our estimates of the duration of the evolution of condensations and the corresponding increase in their mass, first obtained with taking into consideration of depletion effects.

Growth of the mass $m(t)$ of a massive condensation in the terrestrial zone depending on parameters $\gamma$ and $\theta$ (see Table). The curves calculated at $\theta = 10$ are marked by *, ones calculated at $\theta = 5$ are marked by **.

<table>
<thead>
<tr>
<th>Curve</th>
<th>1**</th>
<th>1*</th>
<th>2**</th>
<th>2*</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\theta$</td>
<td>5</td>
<td>10</td>
<td>5</td>
<td>10</td>
</tr>
<tr>
<td>$t_{\text{max}}$, $10^3$ yrs</td>
<td>15</td>
<td>8</td>
<td>44</td>
<td>23</td>
</tr>
<tr>
<td>$m_{\text{max}}$, g</td>
<td>1.45</td>
<td>1.50</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\gamma$</td>
<td>3.5 $10^{20}$</td>
<td>5 $10^{21}$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**References:**


**SPACE TIME LOCALIZATION OF UNDISCOVERED PLANETARY SATELLITES.** N.I.Perov\(^1\) and A.A.Nahodneva\(^2\), \(^*\)State Pedagogical University, Astronomical Observatory, Respublikanskaya, 108. 150000, Yaroslavl, Russia, e-mail: perov@yspu.yar.ru, \(^*\)State Pedagogical University, Respublikanskaya, 108. 150000, Yaroslavl, Russia.

**Introduction:** Dynamical evolution of the Solar planetary system [2], [6], [3], [4], and since 1995 evolution of the exoplanets [9], is one of the main problems of celestial mechanics. Discovery, based on the observations, of extrasolar planetary systems (at May 2005 there had been found 150 planets like Jupiter) was one of the important results of the astronomy of the end of XX century and stimulated development of astronomy. The general direction of theoretical works, devoted to extrasolar planets, is improving of the cosmogonical models. The modern cosmogonists consider the discoveries of planets Earth-type and **minor bodies**, formed new planetary systems, will be in observing astronomy in the first and the second decades of the XXI century [7], [8].

We determine of regular, irregular, coorbital satellites of planets as natural celestial bodies, diameters of which are no less 1 km, and which are revolving around the planets (distance between the planet and the satellite is smaller by a factor of several orders in comparison with the distance between the planet and the Sun). In this case the <planet centric> force predominates over forces, are due the influence of the Sun, others planets and secondary satellites, oblateness of the planet, though the letters may set up significant perturbations of satellites orbits. It should be noted for the solar planets satellites the perturbations from the others planets are small, in comparison with perturbations of the Sun, and so others planets perturbations do not determine the motion of the satellites [2], [6].

In 1999-2005 many new small satellites of Jupiter (51), Saturn (28), Uranus (13) and Neptune (6) has been detected [4], [10], [11]. Opening this unusual collection of the satellites, discovered for the short interval of time, is an outcome of application of special methods for searching these objects and using of modern equipments, including 8.3 m telescope "SUBARU", permitting to scan across the great regions of sky near the giant planets [10].

In accordance with aforesaid it is very important and interesting to estimate by **theoretical way** numbers of unknown satellites of Saturn, Uranus, Neptune (the number of known satellites of Neptune remains almost constant for the decade and a half, after the flight of "Voyager-2" near this planet [5]) and satellites of the exosolar planets basing on the known parameters of these planets.

**Oblateness of the solar planets and number of theirs satellites:** In the table 1, made up with due account of observed data [4], [10], [11] connection between number (N) of the secondary satellites and geometric oblateness (α) of the Solar system planet is set up. N\(_o\) is known number of the observed planetary satellites and N\(_t\) (α) is a number of satellites, calculated by a formula (1)

\[
N_t(\alpha) = -0.0213 + 312.4235 \cdot \alpha + 10836.650 \cdot \alpha^2
\]

Table 1. Oblateness (α) of the Solar system planets and number (N) of the planetary satellites (19.08.05).

<table>
<thead>
<tr>
<th>Planet</th>
<th>Oblateness, α</th>
<th>Number of observed satellites, N(_o)</th>
<th>Theoretical number of satellites, N(_t) (α)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mercury</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Venus</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Earth</td>
<td>0.0034</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Mars</td>
<td>0.0052</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>Jupiter</td>
<td>0.062</td>
<td>63+ rings</td>
<td>63</td>
</tr>
<tr>
<td>Saturn</td>
<td>0.103</td>
<td>47+ rings</td>
<td>147</td>
</tr>
<tr>
<td>Uranus</td>
<td>0.06</td>
<td>28+ rings</td>
<td>58</td>
</tr>
<tr>
<td>Neptune</td>
<td>0.02</td>
<td>13+ rings</td>
<td>13</td>
</tr>
<tr>
<td>Pluto</td>
<td>?</td>
<td>1</td>
<td>?</td>
</tr>
</tbody>
</table>

The equation (1) is derived with help of the least square method (the planets and the satellites up Jupiter including are considered). The great number of the significant digits underscores the negligible influence of computers errors (32 significant figures with help of system REDUCE are kept up). Criterion of Fisher - Snedecor [1] for the (1) equations gives \(F=74184.8>>F_{0.01; 2;2}=999.5\) (\(F_{0.01; 2;1}=4999.5\)), that is evidence of significance of regression equation (1). Moreover, coefficient of determination is \(R^2_{\text{num}}=0.999986\) and variance is \(S_i^2=0.0196\). It is clear, \(N_i(\alpha) = 0\) for \(\alpha_{10}=0.028915\) and \(\alpha_{20}=0.00008414\). The minimum of the function of \(N_i(\alpha)\) is \(-N_i(-0.014415) < 0\), but usually only positive values of oblateness (α>0) are dealt with.

In the frame of the restricted three body problems it is proved the great value of oblateness of the planet interfere with falling dawn of the nonecliptical satellites on surface of the central body [2, 6]; Poincare’s and Cruvedy’s theorems impose restrictions on angular velocity of rotation and geometric oblateness of gravitating liquid in a state of relative equilibrium.

Basing on the table like table 1 [13] we had suggest in 2003 the hypothesis: with help of cosmic mission “Cassini” 116 satellites of Saturn would be discovered.
since July 1, 2004 and the geometrical oblate nesses of Mercury and Venus are about of 0.00008 and oblateness of Pluto equals approximately 0.003. (Since 2003, 15 satellites of Saturn have been revealed).

Table (1) illustrates good agreement $N_0$ and $N_t(\alpha)$ for the planets nearest to the Earth (and space of near which is better investigated). Since for Neptune $N_0 - N_t(\alpha) = 0$, we should wait the satellite system of Neptune are not so developed as the satellite systems of Jupiter, Saturn and Uranus.

For determination of space-time position of unknown satellites of Saturn in paper [14] new quadrature is obtained analytically. It shows dependence eccentricity $(e)$ of osculating orbit against time $(P)$.

$$2 \int \frac{dw}{\sqrt{\vartheta}} = P$$

$$\vartheta = \vartheta_1 \vartheta_2$$

$$\vartheta_1 = 4\varphi c_1 w^7 - \frac{4}{3} \varphi w^5 + w^2 (-c_2 + 2 + 2c_1) - 4$$

$$\vartheta_2 = -4\varphi c_1 w^7 + \frac{4}{3} \varphi w^5 - 10c_1 w^4 + w^2 (8 + 8c_1 + c_2) - 6$$

$$w = \frac{1}{\sqrt{1 - e^2}},$$

$$c_1 = (1-e^2)\cos^2 i,$$

$$c_2 = \frac{2\gamma}{\sqrt{1-e^2}} \left(1 + \cos 2\gamma\right) + 2(e^2 - \sin^2 i) + e^2 \sin^2 i (5 \cos 2\omega - 3).$$

$$\gamma = \alpha/(2\beta),$$

$$\alpha = -\frac{3}{8} c_{20} \left(\frac{a_0}{a}\right)^3,$$

$$\beta = \frac{3}{16} \mu r \left(\frac{a}{a_1}\right)^3,$$

$$\tau = \beta (t - t_0) \left(\frac{\mu}{a}\right)^3,$$

where $a, \beta, \gamma, c_1, c_2, a_1, a_0, \mu$ - are parameters of restricted twice averaged problem of 3 bodies, making into account oblateness of the planet and perturbation from the Sun; $\tau$ - modified time of motion of argument of pericentre $(i)$ in the interval $0<\omega<\pi/2$ [15].

For determination of space-time region in which Saturnine satellites may exist put the following conditions:

1. Pericentric distance is greater then the radius $R_S$ of the planet;
2. Apocentric distance is lesser then the sphere of action of Saturn;
3. Lifetime of satellites is greater 1 billion years.

Analyzing formula (2) the following results are drawn up. If initial meanings of $e_0=0.423009; \varphi_0=35^\circ 60404; a=20R_S$, then lifetime equals 38,726 years (satellite falls down on the planet). (See fig.1)

If initial values of $e_0=0.275076; \varphi_0=8^\circ 65344; a=60R_S$, then lifetime equals 18 billion years (such satellite may exist in the given model). (See fig.2)

**Conclusion:** The exact quantitative relationships between numbers of natural satellites of the planets of the Solar system and parameters of these planets, and theirs orbital semi major axis would make it possible to discover and investigate satellite systems of exosolar planets, because the Solar system is not unique in the Galaxy [9]. On the contrary of works M.L. Lidov and M.A. Vashkovyjak [15] our results based on the investigation of quadrature [2] derived from the system of differential equations of motion of satellite in the frame of the restricted twice-averaged 3-body problem. It is very interesting to make up a method of space-time localization of undiscovered planetary satellites based on new model problems of celestial mechanics integrated in a final form.

**References:**
MODELING POLARIZATION PROPERTIES OF LUNAR REGOLITH WITH T-MATRIX APPROACH. D.V. Petrov, E.S. Zubko, E.N. Synelnyk, Yu.G. Shkuratov. Astronomical Institute of V.N. Karazin Kharkov National University, 35 Sumskaya Street. Kharkov. 61022. Ukraine. e-mail: petrov@astron.kharkov.ua.

Introduction: The lunar regolith has well-studied dependence of the linear polarization degree on the phase angle $\alpha$. The main features of this dependence are the negative polarization branch at small phase angles and the positive polarization branch with maximum near 105°.

The negative polarization of the Moon is demonstrated in Fig. 1 that presents the integral polarimetric data [1]. The negative polarization branch for different lunar regions is almost the same; it has a parabolic shape with the inversion angle (near 23°) and the depth (near 1%). The average albedo of the lunar surface is rather low, near 12 % in the visible spectral range. This means that single particle scatter can dominate the lunar surface polarization response. Thus it is reasonable in first approximation to use theories of single particle scatter by particles with irregular shapes to model the polarization features. In this work we exploit the T-matrix method for calculations of scattering by wavelength-size scatterers to study a role of shape of scatterer in formation of the observed polarization.

Model of regolith particles: The shape of particle can be described with an angular dependence of the distance from the particle center to the particle surface, $R(\theta, \varphi)$, where $\theta$ and $\varphi$ are polar and azimuth angles, respectively. As a model of lunar regolith particles we used randomly shaped grains with a Gaussian distribution of radii. The Gaussian shape can be generated by expansion of $R(\theta, \varphi)$ into series over spherical harmonics and Legendre polynomials. Thus the shape can be described by the following equation [2]:

$$R(\theta, \varphi) = \frac{e^{i(\theta, \varphi)}}{\sqrt{1 + \sigma^2}}, \quad (1)$$

where

$$s(\theta, \varphi) = \sum_{l=0}^{\infty} \sum_{m=-l}^{l} P_l(\sin \theta) [\alpha_{lm} \cos m \varphi + \beta_{lm} \sin m \varphi] \quad (2)$$

where the coefficients $\alpha_{lm}$ and $\beta_{lm}$ are independent Gaussian random variables with zero mean and equal variances:

$$\beta_{lm}^2 = (2 - \delta_{lm}) \frac{(l - m)}{(l + m)} c_l \beta^2 \quad (3)$$

$$\beta^2 = \ln(1 + \sigma^2), \quad (4)$$

$$c_l = (2l + 1) \exp[-\kappa] I_l(\kappa), \quad (5)$$

$$\kappa = \frac{1}{4} \left( \sin \frac{\Gamma}{2} \right)^2, \quad (6)$$

where $\sigma^2$ is the radii variance, $\Gamma$ is the correlation angle, $I_l(z)$ is the modified spherical Bessel function, $P_l(z)$ is the Legendre polynomials.

Fig. 1. Phase dependence of the linear polarization degree of the Moon (line with points) [1]. To model the lunar regolith particles, a Gaussian random shape with $\sigma = 0.3$, $\Gamma = 15^\circ$, and the refractive index $m = 1.5$ were used. Curves shown correspond to particles with the size parameter $x = 15$ (solid curve) and for averaging over sizes $x = 10.5 \div 15.9$ (dashed curve).
This method allows us to generate irregular particles. An example of such particles is given in Fig. 2. As can be seen the correlation angle strongly affects the shape of irregular particles.

**Method of calculations:** We used the T-matrix method to calculate scattering properties of particles. This method allows us calculations of the scattered field in any point of space. The main idea of this method is the expansion of the incident and scattered fields in spherical wave functions [3]. The T-matrix gives a relationship between the expansion coefficients of incident and scattered fields. The T-matrix method allows one to carry out averaging over orientations [4] and sizes [5].

**Results and discussion.** Using this method we have calculated the phase dependence of the linear polarization degree of model particles (Fig. 2); we have found scattering properties of all types of the particles shown. We used the refractive index $m = 1.5+i0$. The most appropriate type is the particle (d) with $\sigma = 0.3$ and $\Gamma = 15^\circ$. Curves shown in Fig. 1 correspond to particles with the size parameter $x = 15$ (solid curve) and for the case of averaging over sizes $x = 10.5 \div 15.9$ (dashed curve). In all cases averaging over orientations was carried out. For comparison we show also the linear polarization degree of the lunar regolith (line with points). As can be seen, the curve corresponding to the lunar regolith cannot be reproduced with the only size parameter. However, if one uses the averaging over sizes, the model phase function, at least qualitatively can be fitted to the observed curve.

![Fig. 2. Examples of random Gaussian shapes. $\sigma = 0.3$; (a) $\Gamma = 60^\circ$; (b) $45^\circ$; (c) $30^\circ$; (d) $15^\circ$.](image)

**References:**

MOON MINERALOGY MAPPER (M3): SCIENCE AND EXPLORATION OPPORTUNITIES. Carle M. Pieters1 and the M3 Team, 1Department of Geological Sciences, Brown University, Providence, RI 02912 (Carle_Pieters@brown.edu).

The Moon Mineralogy Mapper (M3, or "m-cube") is a state-of-the-art imaging spectrometer that will characterize and map the mineral composition of the Moon. M3 will be flown on Chandrayaan-1, the Indian Space Research Organization (ISRO) mission that is scheduled for launch in late 2007. The Moon is a cornerstone to understanding early solar system processes, and M3 high-resolution compositional maps will dramatically improve our understanding of the early evolution of a differentiated planetary body and provide a high-resolution assessment of lunar resources.

M3 is one of several foreign instruments chosen by ISRO to be flown on Chandrayaan-1 to complement the already strong ISRO payload package. After a detailed NASA peer-review process, M3 was selected for funding through NASA’s Discovery Program as a Mission of Opportunity. M3 is under the overall oversight of PI Carle Pieters at Brown University. It is being built by a highly talented and committed JPL team led by Tom Glavich as Project Manager and Rob Green as Instrument Scientist. Each member of the M3 Science Team is uniquely experienced and has a specific responsibility for data calibration, analysis and/or interpretation. The rest of the Science Team includes: J. Boardman, B. Buratti, R. Clark, JW. Head, T. McCord, J. Mustard, C. Runyon, M. Staid, J. Sunshine, LA Taylor, and S. Tompkins.

The primary science goal of M3 is to characterize and map lunar surface mineralogy in the context of lunar geologic evolution. This translates into several sub-topics relating to understanding the highland crust, basaltic volcanism, and potential volatiles. The primary exploration goal is to assess and map lunar mineral resources at high spatial resolution to support planning for future, targeted missions. These goals translate directly into requirements for accurate measurement of diagnostic absorption features of rocks and minerals, with sufficient spectral resolution for deconvolution and sufficient spatial resolution for context. These requirements are met by M3’s design: visible to near-infrared imaging spectrometer with high signal to noise, and excellent spatial and spectral uniformity.

M3 spectral requirements are for a 0.7 to 3.0 μm range (optional to 0.43 μm is the baseline). Measurement are obtained for 640 cross track spatial elements and 261 spectral elements. This translates to 70 m/pixel spatial resolution and 10 nm spectral resolution (continuous) from a nominal 100 km polar orbit for Chandrayaan-1. Spectra of lunar soils and minerals sampled to the full resolution of M3 are shown in Figure 1. The M3 FOV is 40 km in order to allow contiguous orbit-to-orbit measurements at the equator that will minimize lighting condition variations.

Over the two-year mission lifetime, there are four periods of optimal lighting conditions for spectroscopic measurements (two 2-month periods/year). One period will be devoted to global assessment at reduced resolution (320 spatial elements, 87 spectral) and the other three will be devoted to obtaining full resolution data for prioritized targets (10-50% of the surface). The nominal mission relies on India’s new Bangalore DSN facility for data downlink. This implies that data for only 6 of 12 orbits/day will be transmitted when the Moon is in sight of Bangalore DSN. The nominal measurement timeline and data collection sequence for M3 is shown in Figure 2.

Figure 1. Example reflectance spectra of lunar minerals and soils sampled to M3 full resolution. [The weak feature near 2900 nm is due to trace amounts of terrestrial water remaining on the samples in a purged environment.]

Figure 2. M3 data acquisition periods during a nominal Chandrayaan-1 mission timeline.

**Mission Status:** M3 has had a Preliminary Design Review (PDR) and is scheduled for Critical Design Review (CDR) in early 2006. International agreements are proceeding in parallel with instrument design and implementation.
**Introduction:** Contrast enhancement is a common operation for enhancing detail in planetary image data. The enhancement exposes subtle patterns already present in the image by mapping a range of values in the input image to a larger range in the displayed image. A drawback of this technique is that if an image has regions with widely varying albedo, stretching the gray-scale levels for the entire image will typically saturate levels of high or low albedo. Having one or more saturated regions dramatically reduces the useful information presented in the stretched image. Here we present a technique to interactively (and independently) enhance contrast stretching of subregions within a single image.

**Related Work:** Traditionally, contrast enhancement involves analyzing the histogram of grayscale values and mapping a subset of the range to the entire 0 to 255 spectrum. Adaptive Histogram Equalization [2] is a more sophisticated approach that can spatially enhance details and provide good results. In its implementation, Matlab provides a number of parameters that can control different settings of contrast enhancement. However, there is no notion of identifying and processing similar regions in the image separately. In this work, we are interested in interactively segmenting the image into regions and performing contrast enhancement on the derived regions.

For region extraction, we rely heavily on the work of Levin et al. [1] for determining colored regions based on a few sparse marks. Their work tries to map the chrominance channels of a grayscale image to the luminance intensities based on sample strokes. A linear system of equations is setup which propagates the colors to the unmarked regions of the image. The result is a colored image. Since the system is interactive, as opposed to a fully automatic system, the user can iteratively refine the strokes to improve the quality of the colorized image.

**Our approach:** Our framework provides the user with an interface to mark colored strokes on an image, utilizing a different color for each distinct region of the image. Levin’s algorithm is then invoked to calculate the colorized image. Note that we work in LUV representation of colors rather than RGB, where the L channel corresponds to the intensity and UV channels correspond to the chromaticity of the pixel color. Once the UV channels are obtained for the colorized image, we need to segment the image into regions. The number of regions is equal to the number of unique marked colors and we classify each pixel as belonging to one of the colors. We use a simple L2 norm between the pixel UV values and the UV values of the marked colors. The marked color with the minimum distance is then assigned as the label of the pixel.

Once the classification has been performed, the user is provided with individual histograms for each region. The user can adjust settings on the histogram to perform interactive contrast enhancement.

Fig 1 shows a sample interactive session. The user scribbles strokes to indicate the glacier and the valley floor. Note that the strokes are relatively sparse and the user does not need to precisely mark out boundaries. Levin’s algorithm then computes the colorized image, which is then thresholded to give the masks. Contrast is enhanced separately on these sub-regions to bring out details that were not apparent with a single stretch for the entire image. Fig 3 and 4 show results obtained by our technique in comparison to other methods.

**Limitations:** Automatic segmentation is a hard problem in the field of image processing. Although Levin’s algorithm is interactive and the quality of the segmentation can be iteratively refined, the algorithm will fail to produce plausible results if the user maps similar neighboring intensities with different scribbles. In general strong changes in intensities improve the quality of the segmentation. Another problem with our approach is the post-stretch appearance of boundary pixels. Since the contrast enhancement is performed independently for the regions, the boundaries adjacent to the regions will look unnatural. Blending the boundary pixels to a plausible range remains future work.

**Applications:** This method of contrast enhancement is ideal for initial analysis of many different types of image data. It is especially useful for images with low solar incidence angles, where extreme bright and dark regions are associated with topography (i.e., crater walls, scarps, etc). A few potential applications involve stretching regions covered by shadows or separately stretching distinct units within the same image.

Fig. 1: Steps of our interactive session for the Antarctica dataset. The user marks some strokes to indicate the three regions (top right). A colored image is calculated (bottom left) which is then thresholded to give the masks (bottom right).

Fig. 2: The same process applied to a different dataset. Here the user is trying to separate brightly lit areas, shadowed areas and the rest of the terrain.

Fig. 3: Results from naïve contrast stretch (left), adaptive histogram equalization (center left). Comparable results using region based contrast enhancement (center right) and a different stretch (right).

Fig. 4: Results from naive stretch (left) adaptive histogram equalization (center) and region based contrast enhancement (right).
INITIAL RESULTS FROM USING CUBIC AND VARIATIONAL IMPLICIT INTERPOLATION FOR COMBINING HRSC STEREO TOPOGRAPHY WITH MOLA LASER ALTIMETRY. Prabhat\textsuperscript{1} and C. I. Fassett\textsuperscript{2}, \textsuperscript{1}Center for Computation and Visualization (prabhat@cs.brown.edu), \textsuperscript{2}Department of Geological Sciences (Caleb_Fassett@Brown.Edu), Brown University, Providence RI 02912, USA.

**Introduction:** Combining MOLA and HRSC topography information is useful because the two datasets give us knowledge about the Martian surface in very different ways. MOLA was a high-precision instrument but returned sparse data, and HRSC is somewhat less-precise but allows for creation of complete topographic models. In this work, we present some preliminary results from using simple interpolation functions (cubic and variational implicit) to combine the two sources of information. Our preliminary results look promising and warrant further investigation.

**Data sources:** \textsuperscript{MOLA:} The MOLA instrument was a laser-ranging device; it operated by sending very short pulses of infrared radiation to the surface and measuring the time it took before it was reflected back to the spacecraft [1]. Before the laser ceased operation, MOLA made \(\sim\)660 million individual measurements of the Martian surface. Shots are separated by roughly 300 m along orbit, the footprint of MOLA shots is \(\sim\)100 m, and the geolocation of shots is known to better than 100 m. The vertical precision of MOLA is on the order of \(<\) 0.5 m, and analysis of crossover shots of multiple orbits suggests the repeatability or drift in measurements is \(<\) 1 m [2]. MOLA-derived DTMs are derived by gridding of MOLA shot data, but obviously such grids are interpolated across regions where MOLA did not make measurements. Although the high-precision measurements made by MOLA have given us valuable information about Martian topography, the sparse coverage presents challenges, particularly in the more equatorial latitudes.

\textsuperscript{HRSC:} The HRSC instrument has nine data channels for nadir imaging, photometry, color (red/green/blue/IR), and stereo imagery [3]. The nadir channel has resolution up to \(\sim\)12 m/pixel over a large, continuous image swath, which is ideal for interpreting surface geology. Using the HRSC stereo channels, it is possible to derive a scene-wide topographic model with spatial resolution of \(\sim\)50-200 m/pixel, depending on the observing conditions [3]. Unfortunately, the orbit of the Mars Express mission is not circular, so that the resolution of HRSC images and the derived DTM changes dramatically depending on the distance from the Martian surface. Derivation of stereo models requires knowledge of the camera optics (which are assumed stable) and tie points between image locations on the two stereo images. A surface topographic model can then be derived using the parallax between observations. An automated processing routine has been developed for HRSC data [7], but in practice the derivation of stereo models is computationally intensive and difficult. DLR produced \textquote{standard} 200-m resolution DTMs have been produced and released to the HRSC team, which we use as the reference HRSC model here.

There is a misfit of the HRSC DTMs to the MOLA points of tens to thousands of meters. There is thus a need to incorporate MOLA data into the HRSC processing to improve the DTM product. Recent work by Spiegel et al. [4] has begun to consider this problem, but their solution differs substantially from ours, in that they attempt to adjust the HRSC DTM to incorporate information derived from the MOLA gridded data. As outlined below, we attempt to combine the HRSC DTM with MOLA data using a constraint imposed by the MOLA shots themselves.

**Our approach:** Assuming that the input MOLA and HRSC data is perfectly registered, we would like to displace the HRSC grid, so that the ideal surface satisfies the following two properties:

1) it should pass through all the MOLA points

2) it should introduce minimal distortion to the dense HRSC grid.

We first calculate a displacement grid by taking the difference between MOLA points and the corresponding HRSC points. Since this grid is sparse, we can rephrase the goal as trying to fit a smooth surface to the displacement grid.

We have tried two approaches to solve this problem. We use MATLAB’s griddata function [5] with cubic interpolation and a variational implicit function [6]. For the results reported in this work, we use a basis function of the form \(f(x) = x^3/2\). The cubic interpolation functions maintain \(C^0, C^1\) and \(C^2\) continuity. The implicit function passes through the specified MOLA points while minimizing the aggregate squared curvature. While this is a desirable feature, this does not completely address the second property. Formulating an optimization function that would attain a compromise between both properties needs to be explored further in future work.

We tested our technique on two subsets of HRSC DTM data from the Arsia Graben region (Fig 3) and Nili Fosae region (Fig 1,2). The Arsia Graben dataset has 7324 MOLA points and 501x258 HRSC grid. The
Nili Fosae dataset has 8216 MOLA points and 298x241 HRSC grid. A summary of the quantitative results obtained on a Linux machine (2GHz, 4GB RAM) is presented below.

<table>
<thead>
<tr>
<th>Cubic interpolation</th>
<th>Implicit function</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time (s)</td>
<td>MSE</td>
</tr>
<tr>
<td>Arsia</td>
<td>0.93</td>
</tr>
<tr>
<td>Nili</td>
<td>0.51</td>
</tr>
</tbody>
</table>

This work was performed under a NASA grant to J.W. Head for participation on the HRSC team.

References:

Fig. 1: Results for the Nili Fosae region. The raw MOLA data (left) and HRSC grid (center) is shown. Resulting mesh from variational implicit function is shown on the right. Note that the MOLA mesh on the left is generated using cubic interpolation for visualization purposes alone.

Fig. 2: Central figure shows cubic interpolation results. Figure on left shows the displacement applied to the HRSC grid under variational implicit interpolation; the figure on the right is the difference between the two methods.

Fig. 3: Results from Arsia Graben region. MOLA (left), HRSC(center) and implicit interpolation (right) meshes.
THE PARAMETERS INVOLVED IN HAPKE’S MODEL FOR ESTIMATION OF THE COMPOSITION OF THE EJECTA LUNAR TERRAINS. S.G. Pugacheva, V.V. Shevchenko. Sternberg State Astronomical Institute, Moscow University, 13 Universitetsky pr., 119992 Moscow, Russia, pugach@sai.msu.ru.

**Introduction.** In the previous papers we were estimated the surface roughness of the ejecta lunar terrains by means comparison of the local phase function and the average integrated lunar indicatrix. The difference between the modeled and observed phase functions was obtained for surface having various degree of the surface roughness. We were investigated the surface of the lunar landing sites. The great difference between the modeled and observed phase functions demonstrates for phase angle in range about 18° and corresponds a high degree of the surface roughness [1]. The value of this difference of intensities was used as a photometric parameter of the surface roughness \( \Delta I \). In this article we compared the values photometric parameter of the roughness \( \Delta I \) with the parameters the Hapke’s model.

**The Hapke’s theoretical model of the lunar surface reflection.** The Hapke’s formula is well known model for the estimation of the surface roughness [4, 5, 6]. The model allows to define of the physical and chemical characteristics of the lunar surface by means phase function. This model is the only existing one, which accurately represents the reflection properties of the Moon. The photometric function by Hapke of the bidirectional reflectance (R) may be written in the general form:

\[
R = \frac{w}{4\pi(\mu_o + \mu)} \left[ 1 + B(g)P(g) + H(\mu) - 1 \right] S(\theta),
\]

where \( \mu_o, \mu \) are cosines of incidence (\( \mu_o \)) and emergence (\( \mu \)) angles, \( g \) is the phase angle, i, e are angles of incidence and emergence, \( w \) is the single scattering albedo, \( B(g) \) is the opposition effect function, \( P(g) \) is the phase function, \( H(\mu) \) is the multiple scattering function, \( S(\theta) \) is the function for macroscopic roughness.

The function \( B(g) \) describes the brightness of a surface near zero phase angle. The parameters of the function \( B(g) \) describe the backscatter due to blocking and shadowing within the soil.

\[
B(g) = B_o \left[ 1 + \frac{1}{h} \tan \left( \frac{g}{2} \right) \right].
\]

The parameter \( h \) characterizes compaction of the regolith and size of the particle. The parameter \( B_o \) defines amplitude of the opposition effect.

The function \( P(g) \) includes two parameters \( b \) and \( c \), which determines the phase function form and the nature of scattering (\( c<0.5 \) corresponds to forward scattering and \( c>0.5 \) to backward scattering).

\[
P(g) = \frac{(1-c)[(1-b^2)/(1+2bcos(g)+b^2)^{3/2}]+c[(1-b^2)/(1-2bcos(g)+b^2)^{3/2}]}{1+B(g)P(g)},
\]

where \( g \) is the phase angle, \( b \) and \( c \) are the parameters with the material properties of the lunar regolith.

The function of the isotropic multiple scattering \( H(x) \) includes angles incidence and emergence and the single scattering albedo \( w \).

\[
H(x) \approx \frac{(1+2x)/(1+2x(1-w)^{1/2})}{1+2x(1-w)^{1/2}},
\]

where \( x \) is either \( \mu_o \) or \( \mu \).

The equation \( S(\theta) \) allows to calculate the effects of macroscopic roughness on light scattered by a surface having an arbitrary diffuse-reflectance function [3, 6]. The parameter \( \theta \) is a mean topographic slope angle of the surface. The effects of macroscopic roughness will modify the reflectance. The geometry of the reflectance is shown in figure 1. The signal \( r(i,e,g) \) is interpreted as if it came from a smooth, horizontal area, the bidirectional reflectance from sloping area is \( r(i_e,e,g) \). The expressions for \( r(i,e,g) \), \( r(i_e,e,g) \) and \( S(\theta) \) will make the following equation true:

\[
R(i,e,g) = r(i_e,e,g)S(\theta).
\]

The mean slope angle \( \theta \) equal

\[
A = 1/(\pi \tan^2 \theta) \quad \text{and} \quad B = 1/(\pi \tan \theta),
\]

where \( A \) and \( B \) are empirical coefficients of the equation described by a Gaussian [6].

The Hapke’s theoretical integral phase function involves six parameters: \( w, B_o, h, \theta, \) and two parameters to describe \( P(g) \). For
a low-albedo lunar surface parameter $B_0 = 1$. The parameters have been used in a number of analyses to describe the integral scattering properties of lunar objects.

Figure 1. The parameter $\theta$ is the mean slope angle, which indicates the rough-surface bidirectional reflectance. (From Hapke, 1993 [6])

The bi-directional reflectance of the lunar scatters of landing sites. We used the Monte Carlo method for calculation of the model parameters of the Hapke’s function. The algorithm of the calculation is minimizing the difference between measured and modeled reflectance spectra. The parameter $\theta$ was calculation from the empirical expression [6]. The Saari and Shorthill catalogue were used as observed phase functions [2]. The separate points represent of number of landing sites (Surveyor I, III, V, VI, VII, Apollo 11 and 12, Lunokhod 1 and 2). Each group can be corresponded to the ejecta with individual character of the surface distribution of KREEP materials. Figure 2 represents the diagram of relationship between the photometric roughness parameter ($\Delta I$) and parameters of the Hapke ($w, h, b, c, \theta$) in different lunar landing sites. The reflectance of real, high- and low-albedo surfaces contains a similarly shaped crater. The changes of brightness of the surface are altered by increasing values roughness and albedo. The physical albedo and integral phase function of the Moon may be corrected for effects of macroscopic roughness.

Conclusions. The present work allows study optical characteristics of surface and to determinate grain sizes and material composition of the ejecta of the lunar terrains. The texture roughness in the submillimetric and centimetric range is the most representative than a high macroscopic roughness with large slopes at the hundred meter scale. According to these results, there may be a possibility to describe the real case of natural regolithic surface, and to investigate specific anomalous of the local KREEP assimilation.


Figure 2. Hapke’s parameters plotted against the photometric roughness parameter ($\Delta I$).

Introduction. High-resolution topographic information is important in different applications for the study of planetary surfaces: their evolution, state, and properties.

The most precise topographic data are obtained through direct measurements with different altimeters on board of spacecrafts (e.g., [1]). Indirect techniques for surface height estimates involve photogrammetry (e.g., [2]), photoclinometry (e.g., [3]), and interferometry (e.g., [4]).

Illumination-viewing conditions for observations of planetary surface are important for relief reconstruction with indirect methods. These conditions are usually limited during real missions by spacecraft orbit geometry, particular time of observations, spacecraft life time and scientific program which lead to the limited number of observations of the same surface areas.

In the present work we analysis relationship between illumination-viewing conditions of surface observations and spatial error distribution arose during surface relief reconstruction with photometric method proposed in [5]. The method is the most mathematically correct and can be classified as a photoclinometry method since it uses the fact that observed brightness of the surface depends on the surface orientation.

The study was made using simulated observations of the test surface in the optical wavelength range and for radar experiments.

Photometric method for the planet relief determination. The photometric method used here is based on the statistical approach to a problem of surface relief determination from a set of images taken under different viewing conditions. The brightness $J(x,y)$ of an element with coordinates $(x,y)$ in the $j$-th image can be written as

$$J_j(x,y) = \sum_{x'=x}^{x+\Delta x} \sum_{y'=y}^{y+\Delta y} g_j(x-x', y-y') I_j(x', y') dx' dy' + N_j(x,y),$$

(1)

$N(x,y)$ is a realization of recording noise, $g()$ describes the blurring of the true brightness distribution $I(x,y)$ due to recording system transfer function, phase distortions of the radiation during the propagation through a medium, and so on.

For small surface tilts, the true surface brightness $I(x,y)$ can be expanded into a Taylor series, and with two series terms it can be presented as

$$I_j(x,y) = I_{0j} + c_{xj} \frac{\partial H(x,y)}{\partial x} + c_{yj} \frac{\partial H(x,y)}{\partial y},$$

(2)

where the pair $(\partial H/\partial x, \partial H/\partial y)$ is a two-dimensional height gradient vector, and the pair $(c_{xj}, c_{yj})$ is the first derivative of the photometric function. In the frame of our approach, the photometric function has to be known a priori.

For an ideal case with no noise, the set of equations (1) and (2) gives exact solution for the surface relief $H(x,y)$ (the height deviation above a mean plane). For actual experimental data accompanied by the noise, the same set may allow no solution at all.

The best way to solve the problem with real observations is the statistical approach. This approach allows determination of the height distribution that is the most probable for the given set of images (Eq. 1).

$$\tilde{H}_m(k_x, k_y) = \sum_j D_j^* \tilde{g}_j(k_x, k_y) J_j(k_x, k_y) \alpha(k_x, k_y) + \sum_j (D_j^*)^2 W_j,$$

(3)

where $W_j = g_j^*(k_x, k_y) \tilde{g}_j(k_x, k_y) J_j(k_x, k_y)$, and $D_j = k_x c_{xj} k_y c_{yj}$. Here $\tilde{g}_j(k_x, k_y)$ and $\tilde{J}_j(k_x, k_y)$ are Fourier transforms of $g_j(x,y)$ and $J_j(x,y)$, respectively; “*” means complex conjugation. $\alpha(k_x, k_y)$ and $\beta_j(k_x, k_y)$ are reciprocal of height and noise spectral density, respectively.

Eq. (3) specifies an optimum filter for derivation of the most probable relief from the set of images. This filter depends strongly on a-priori known factors that distort received signals. This approach can correctly account for statistically independent additive gaussian noise and blurring factor.

Viewing conditions and error fields for the determined relief. For the present study we used test
lunar-like cratered surface. Slopes of the test surface shown in Fig. 1 are varied between 0 and 31° as for real planetary surfaces. The highest slopes are located at the largest crater rim (brightest shadows in Fig. 1).

We obtained sets of images of the test surface with two photometric functions to simulate optical and radar observations. Lambert law was chosen as a photometric function described light scattering in the optical wavelength range, and Muhleman law [7] specified for radar observation of the Venus surface [8] was used to simulate radar experiment. Examples of initial optical and radar images used for the relief reconstruction are shown in Fig. 2a and Fig. 3a, respectively. We chose the geometry of experiments when direction to the source of the flux is coincided with direction to the receiver as for radar experiments. Incidence angle was 35.5°. In Fig. 2a, 3a the direction of surface illumination is shown with arrows. Illumination-viewing azimuth angle (A) here is 90° (anticlockwise from the bottom). Images in Fig. 2a, 3a are differed rather high from each other in brightness contrast. Brightness contrast is higher for radar observations (Fig. 3a).

We reconstructed surface topography with Eq. 3 using one image and two images with different illumination-viewing azimuths. Spatial distributions of differences between calculated and test relief are shown in Fig. 2, 3 for: (b) – one initial image (A=90°); (c) – two initial images (A=90°, -90°); (d) – two initial images (A=0, 90°). Spatial distributions of height errors are similar when relief was reconstructed using one image (Fig. 2b) and two images observed from opposite to each other sides (Fig. 2c) for simulated optical observations. These distributions depend strongly on image coordinate that is normal to the viewing direction. For radar experiments such distributions are different (Fig. 3b, 3c). Absolute errors of heights estimates were lower with the use of two initial images.

Errors distributions of heights reconstructed using two images obtained at normal to each other illumination-viewing directions depend on both coordinates (Fig. 2d, 3d). Errors values in this case are the lowest.

**Conclusion.** Thus, model calculations confirmed that images obtained at normal to each other illumination (viewing) directions are the most preferable for relief reconstruction. The most unfavorable for such purposes is the use of images obtained from opposite to each other directions.

Calculation also shows that relief cannot be reliable reconstructed using one image. But for some cases the use of one image is enough for identification of relief shapes.
PRINCIPAL COMPONENT ANALYSIS OF LUNAR SOIL CHARACTERIZATION CONSORTIUM DATA. D. Stankevich, Y. Shkuratov, C. Pieters, Astron. Institute of Kharkov National Univ. 35 Sumskaya St., Kharkov, 61022, Ukraine. Geological Sciences, Brown Univ., Providence, RI 02912, USA.

Summary: We analyzed spectra of lunar soils in the range 0.35–2.50 μm. Using the Lunar Soil Characterization Consortium (LSCC) database we show the principal component method (PCA) to be a useful technique to find a correlation between spectral properties of the samples and chemical composition. The number of spectral eigenfunctions important to approximate high-informative RELAB spectra is quite high, near 20.

Introduction: Interpretation of spectral properties of lunar materials relies on fundamental characteristics of diagnostic absorptions that are based on principals derived from mineral physics [1]. Such applications, however, require high precision visible to near-infrared spectra (0.3 – 2.6 μm) of high spectral resolution. Spectroscopic analysis has been highly successful for limited targets on the lunar nearside using instruments developed in the late 1970’s [2]. It was shown that “red slope” is controlled by Fe2+ and Ti 4+ ions in lunar mineral; 1 μm band is formed by pyroxenes and olivines; pyroxenes also are responsible for the 2 μm band.

The majority of high spatial resolution data currently available for the Moon comes from multi-spectral imagers that typically consist of a digital framing camera equipped with several filters. The Chandrayaan-1 spacecraft (Indian lunar mission) will be equipped with a scanning imager spectrometer providing high spatial as well as high spectral resolution. For interpretation of future data it is necessary to study relationships between the general spectral properties of lunar soils and their measured compositional characteristics [3]. To probe information in high spectral resolution spectra of the lunar samples obtained in the laboratory over the range 0.35–2.50 μm, a principal-component analysis was used to develop a statistical link between spectral properties and composition.

Initial data. We use the LSCC data on mare and highland soils [4]. The samples are representative of the compositional diversity. We used bulks (<45 μm) and 3 size-particle separates <10, 10-20, 20-45 μm for each sample. High spectral resolution bi-directional spectra were acquired in the RELAB at Brown Univ. Altogether we used 78 independent RELAB spectra, each includes 441 wavelengths. All spectra and composition data are available at: http://www.planetary.brown.edu/pds/LSCCsoil.html. For the analysis presented here, we focus on two compositional parameters that directly influence spectral characteristics of the lunar regolith, these are FeO and TiO2 contents.

It is important to note that our analysis is strictly statistical by nature. The optical properties of exceptionally well characterized soils have been accurately measured, and we simply identify and use the most highly correlated relations.

PCA spectral analysis. We present the results of linear statistical analysis of the relation between lunar sample albedo with regolith parameters and composition. We used the PCA method. The main goal of this method is to find the system of functions, allowing to describe any given spectrum as a series with minimal number of terms. These terms are an eigenfunctions for the correlation matrix of spectra. Eigenvalues show the relative weight of the eigenfunctions. Figure 1 shows eigenvalues for the studied 78 spectra. One can see that these values fall down very quickly with number increase. This means that a relatively small number of the eigenfunctions enables to describe the spectra. The maximal number of the founded eigenfunctions is smaller than the number of spectra due to computational limitations (stability of the inverse matrix). In our case we determined 40 eigenfunctions.

Figures 2 and 3 present spectral curves of the first two eigenfunctions. The first component describes general features of the spectral set. This characterizes the general slope of spectra and the NIR bands centered near 1 and 2 μm. The second component demonstrates more subtle structural features: the spectral bend in visible range and the fine structure of the NIR bands.

The eigenfunctions of higher orders characterize more subtle structure of the spectra.
An important question is how many eigenfunctions should be taken into account for spectra approximation. This number is closely related to the noise of spectra. We developed a new procedure to estimate the number. The main idea is to make the PCA twice. In the first case we used the initial data set. In the second case we used the set with added random noise. Than we compare these two sets keeping the eigenfunctions that remain almost the same. The stable eigenfunctions are important for further analysis, and unstable ones are noise-dependent. In Fig. 4 we plot the correlation coefficients between eigenfunctions of the two sets. The axes are numbered with the eigenfunction numbers. The scale from blue to red colors corresponds to increasing correlation coefficient (from 0 to 1). As can be seen in Fig. 4 the first 18 eigenfunctions remain the same when a random noise with 0.01% amplitude is added. This noise is typical for RELAB data. We note also that increasing the noise amplitude by the factor 3 makes the number of stable eigenfunctions of about 10, i.e. the number is still rather high.

Thus any spectrum from the spectral suite can be expanded to series on eigenfunctions: \( S_k(\lambda) = w_1 \phi_1(\lambda) + w_2 \phi_2(\lambda) + \ldots \), where \( \phi_i \) is the \( i \)-th eigenfunction, \( w_k \) is the \( l \)-th loading coefficient for \( k \)-th spectrum. We use the loading coefficients as a characteristic of each sample. Then, we find the closest correlations between the coefficients and the composition of the samples using a linear combination, e.g., for FeO the we have the following: \( \log(\text{FeO}%) = a_0 + a_1 w_1 + a_2 w_2 + \ldots \), where \( a_m \) is the coefficients providing such a correlation \((m=18)\). Figures 5 and 6 demonstrate correlations between measured and predicted content of FeO and TiO\(_2\), respectively, for the LSCC samples. The correlations are rather high; in both the cases it is 0.896.

**Conclusion:** Thus lunar soil spectra in the range 0.3 – 2.5 \( \mu \)m can be presented with a limited set of spectral functions (on the order of 10) whose a linear combination provides a high accuracy fit to the original spectra. The number of the significant eigenfunctions is higher then in the case of Clementine filter data that also were used to map chemical and mineral composition of lunar soils [5].

Introduction: The mechanisms of maturation of the regolith on the surfaces of atmosphereless celestial bodies were discussed from the very beginning of the studies of lunar soils. The most puzzling feature of a mature soil is the presence of 1000 Å-thick amorphous rims containing nanometer size grains of reduced iron (npFe⁰) on the surfaces of regolith particles [1]. One group of researchers (e.g., [2]) suggests that the rims with npFe⁰ are due to solar wind (SW), whereas the alternative hypothesis was condensation of impact vapor in meteoritic bombardment [3]. Here a new approach to rim formation is presented, with an emphasis on the importance of impact melting of regolith particles, especially, melting of their surfaces in the impacts of submicron scale. The contributions of solar wind, impact evaporation, and impact melting to rim formation are compared.

Experimental basis: There are experiments indicating that heating of Fe-bearing silicate materials to melting or submelting temperatures is enough for formation of npFe⁰, provided that oxygen pressure is low enough. Subsolidus reduction of Fe²⁺ ions in olivine Fo89 and pyroxene En88 was observed in [4]. An indication to Fe reduction in silicate melt may be derived from laser heating experiments [5]. On the irradiated olivine particles amorphous rims with npFe⁰ were found, whereas in pyroxene powder neither amorphous rims nor npFe⁰ were detected on the irradiated particles. Yet a small number of large (~100μm in size) dark amorphous enstatite particles that contained npFe⁰ distributed over all the particle volume were found. Such particles were obviously formed from melt due to poor thermal contact with the neighbors, and not from condensed vapor that covers continually the surfaces of all particles and contribute to rim formation.

The important resemblance of both types of experiments [4,5] is that formation of Fe⁰ grains occurred only at the sites of enhanced diffusion: at crystal surface, planar defects (grain boundaries, cleavage planes, [4]) or in liquid phase or solidifying glass [5]. I. e., even at high submelting temperature, mobility of atoms is too low to form Fe⁰ grains in crystalline phase. This is consistent with the observation that in lunar regolith npFe⁰ occur in amorphous rims and in glasses.

The importance of impact melting: Even at the sites of enhanced diffusion, such as grain boundaries or amorphous material, elevated temperature is required to provide mobility of the atoms high enough for association of isolated Fe⁰-atoms into nanometer grains. The duration of the laboratory experiments on subsolidus reduction (e.g., [4]) was a few hours, which corresponds to cooling times of material heated in impact events of meter scale and higher. However, most frequent heating of the upper regolith layers is due to micrometeorites for which cooling times of heated material is typically <<1 s. Thus, though at negligible oxygen pressure on the lunar surface, reduction of Fe may take place at subsolidus temperature, high temperature lasts too little time for association of Fe atoms into grains in solid silicates. Below we show that fast association is possible in impact melt due to much higher diffusion coefficients D in melts as compared to solids.

Is formation of npFe⁰ possible in impact melting of particle rims?: npFe⁰ formation times. Compare the time required for formation of npFe⁰ grains of the observed average radius r₁ ≈ 25 Å [6] and solidification time of impact melt created on the surface or all over the volume of a regolith particle. Kinetics of npFe⁰ growth has been considered in [7] on the base of ripening theory [8]. The formation time tᵢ for a population of npFe⁰ grains of a radius r₁ is tᵢ ≈ r₁²/2Dξ₀ [8], where D is diffusion coefficient of Fe in silicate matrix and ξ₀ ≈ 0.6 [7]. In melts, D ≈ 10⁻⁵ cm²/s, so tᵢ ≈ 5·10⁻⁹ s.

Rim solidification times. The shortest time of solidification is typical of impact melt formed by the tiniest (submicron) projectiles on the surfaces of the regolith particles. Note that the thickness of the melted zone formed by submicron projectiles is the same as that of amorphous particle rim h ≈ 1000 Å. In this case, cooling occurs by heat conduction to the solid particle material of heat conduction coefficient η ≈ 10⁴ erg·cm⁻¹·s⁻¹·K⁻¹ and heat capacity per unit volume cᵥ ≈ 10⁷ erg·cm⁻³. Solidification time is tₛ ≈ h²/cᵥ/2η ≈ 5·10⁹ s. Thus, even in the impact events of the smallest (submicron) scale, impact melt survive long enough for elementary act of maturation (formation of npFe⁰ grains).

Surface melting times. Taking the mass spectra of submicron micrometeorites [9], and assuming melted mass ≈10 projectile masses, we obtain that more than 95% of the exposed surface of lunar regolith is melted by submicron projectiles in 3·10⁵ to 6·10⁵ years. This is about the average duration of each of 30 exposures of a particle on regolith surface, i. e., more than an order of magnitude shorter than its total exposure age 1.5·10⁷ years [10]. Most traces of microcraters are removed due to fast surface diffusion at submicron scale [11].
However, next cycles of impact melting during the 30 particle exposures do not result in considerable growth of nFe0 grains (from $\approx$50 to $\approx$60 Å only) because their growth is much slower than their formation [7,8]. On the other hand, the cooling time is short enough for solidification of the melted silicate material in amorphous state even on a crystalline substrate. This is not the case for nonsilicate materials, such as ilmenite, that are not apt to form glasses and are not observed in amorphous state either as glass particles or as rims [12].

Thus impact melting and subsequent cooling, being fast enough for amorphization of the melted material, are slow enough for formation of nFe0 grains of the average diameter 50 Å.

Melting of the bulk of regolith particles. According to [9], the largest contribution into micrometeorite mass (and, consequently, into impact melt) is made by projectiles of sizes from $\approx$10 μm to $\approx$1 mm that cause melting of one or a few regolith particles as a whole. The probability of total melting of a regolith particle during surface exposure is $\approx$10%, the rest of the glass particles observed in lunar soils being due to subsurface melting. Because of poor interparticle contacts, the cooling mechanism for an entirely melt regolith particle is irradiation, and solidification time is $\approx$10⁻⁷ s for a typical single particle (60μm in diameter) and $\approx$0.1 s for a 1 mm droplet of silicate material. Estimates on the base of [7,8] show that this time is enough for the formation of nFe0 grains up to $\approx$1000 Å in diameter in the bulk of the glass particles.

Rates of different maturation processes in space: Let us compare the formation time of amorphous rims due to submicron impact melting to such a time for SW and condensation of impact vapor.

Solar wind. Amorphization time is the shortest for this mechanism. Indeed, $\tau_{SW} = \frac{4h \rho_0}{\eta D_j}$, where $n_0 \approx 10^{23}$ cm⁻³ is the number of atoms per unit volume of a solid material, $h \approx 10^{-5}$ cm is the maximum penetration depth of SW, $N_0 \approx 2$ is the average number of atomic displacements per SW particle, and $j$ is the normal flux of SW, factor 4 being due to rotation of a celestial body. Then $\tau_{SW} \approx 8 \times 10^6$ s $\approx$ 300 years near the Earth orbit ($j = 2.4 \times 10^6$ cm⁻²s⁻¹) and $\tau_{SW} \approx 1500$ years in the Main asteroid belt ($j = 4.5 \times 10^5$ cm⁻²s⁻¹). However, this fast amorphization occurs on cold surface, so impact heating is required to form nFe0 grains out of isolated Fe0-atoms.

Condensation of impact vapor. On the base of the data about micrometeorite flux [9], in the assumption that the mass evaporated in an impact is about the projectile mass, we obtain the evaporation rate of $\approx$0.1 Å/year. The vapor is distributed in a few upper particulate layers of regolith and condenses on the surfaces of the particles there, so the 1000 Å-thick layer of condensate covers the particles in 10 upper layers in $\approx$10⁴ years. This shows that condensation of impact vapor contributes to the composition of particle rim, but rim formation by this mechanism is much slower than by melting in submicron impacts. Besides, condensate may be deposited in multiple events by layers of thickness $<100$ Å, whereas melting occurs each time to a depth 500–2000 Å, which provides cooling slow enough for nFe0 formation in each impact event.

Conclusions:

(1) Impact melting of the upper zones of regolith particles by submicron projectiles enables formation of nanophase Fe0 grains of the observed average diameter 50 Å in melt, subsequent cooling being fast enough for preservation of the amorphous structure of particle rim formed in solar wind bombardment.

(2) Impact melting of the bulk of regolith particles favors growth of nFe0 grains up to $\approx$1000 Å observed in agglutinatic glasses.

(3) Impact melting involves an order of magnitude larger volumes than the other mechanisms; in particular, melting of the particle rims occurs much faster than condensation of a film of impact vapor of the same thickness. This makes impact melting the most effective mechanism of formation and growth of nFe0 on the surface and volume of regolith particles.

Thus, impact melting can provide the observed characteristics of mature soils without any additional maturation mechanisms. Consequently, the mechanism may cause regolith maturation both on bodies shielded from solar wind irradiation, such as Mercury, and on asteroids, where collision velocities do not provide impact evaporation but is enough for impact melting.

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VENUS EXPRESS SCIENCE GOALS, PAYLOAD, AND PLANNED OBSERVATIONS D. V. Titov1, H. Svedhem2, and L. Marinangeli3. 1Max-Planck Institute for Solar System Studies (Max-Planck-Str. 2, Katlenburg-Lindau, 37191 Germany, titov@mps.mpg.de), 2ESA/ESTEC (Keplerlaan 1, 2200 AG Noordwijk, The Netherlands, Hakan.Svedhem@esa.int), 3 IRPS, Universita d’Annunzio, Pescara, Italy.

Introduction: The first phase of Venus spacecraft exploration by the Venera, Pioneer Venus, and Vega missions in 1960-90 established a basic description of the physical and chemical conditions prevailing in the atmosphere and at the surface of the planet. At the same time, they raised many questions on the physical processes sustaining these conditions, most of which remain unsolved to this day. The fundamental mysteries of Venus are related to the global atmospheric circulation, the atmospheric chemical composition and its variations, the surface-atmosphere physical and chemical interactions including volcanism, the physics and chemistry of the cloud layer, the thermal balance and role of trace gases in the greenhouse effect, the origin and evolution of the atmosphere, and the plasma environment and its interaction with solar wind. Besides, the key issues of history of Venusian volcanism, global tectonic structure of Venus, and important characteristics of the planets surface are still unresolved. Re-use the Mars Express spacecraft and existing instruments gave Europe an excellent chance to have an almost fully equipped orbiter mission and to make a breakthrough in Venus exploration in a very short time frame.

Science Goals and Payload: Venus Express will aim at a global investigation of the Venus’ atmosphere and plasma environment from orbit, and will address important aspects of the geology and surface physics. The instruments inherited from the Mars Express and Rosetta missions form the core of the Venus Express payload. They are: SPICAV/for IR – a versatile UV-IR spectrometer for solar and stellar occultation and nadir observations, PFS – a high-resolution IR Fourier spectrometer, ASPERA – a combined energetic neutral atom imager, electron, and ion spectrometer, VIRTIS – a sensitive visible and near infrared spectro-imager, a radio science experiment VeRa. This payload set is complemented by two newly developed instruments: the miniature four-channel digital camera VMC and the magnetometer MAG [1].

Compared with the earlier spacecraft missions, Venus Express will make a breakthrough by fully exploiting the existence of spectral “windows” in the near-infrared spectrum of Venus’ nightside, discovered in the late ‘80’-s, in which radiation from the lower atmosphere and even the surface escapes to space and can be measured. A combination of spectrometers/spectro-imagers covering the near-UV to mid-IR range and plasma instruments onboard an orbiter can provide global study of the Venus surface and atmosphere up to about 200 km. The whole set of the Venus Express investigation tools is such that observing the same target at the same time with different instruments provides a comprehensive, versatile view of the phenomena taking place at Venus.

Mission Scenario and Operations: The Venus Express spacecraft will be launched between 26 October and 25 November 2005 by the Russian Soyuz-Fregat launcher from the Baikonur cosmodrome in Kazakhstan. After about five months of cruise Venus Express will be inserted in a polar elliptical orbit around Venus with apocentre distance of about 66,000 km, pericentre of 250-350 km, and revolution period of 24 hours. The nominal mission will last for two Venus sidereal days (~500 Earth days) with possible prolongation for another 500 days. The science operations will include high-resolution observations from the pericentre, global views from the apocentre, limb sessions, stellar, solar, and radio-occultation [2].

Surface studies: Several open questions on the geological evolution of Venus have been raised by the Pioneer-Venus, Venera, and Magellan investigations. The complex stratigraphical sequence observed on Magellan SAR images implies an internally active planet with extensive volcanism and tectonic activity, even in geologically recent time. There is still a debate whether Tessera highlands are different in composition from the Planitia. Moreover, the peculiar environmental characteristics at the surface have produced unique weathering processes of the primary rocks, processes which are still not well understood. Venus Express will contribute to the surface studies in several ways: bi-static radar and gravity experiments by the radio-science (VeRa) instrument and the night side thermal sounding in the near-infrared spectral windows by the spectrometers and imagers [2]. The spectrometers and imagers onboard the orbiter will obtain critical measurements of the gas species and their distribution in the lower atmosphere providing clues on the redox state in equilibrium with surface minerals and eventually detect anomalies due to volcanic emission. Recent modeling demonstrates the capabilities of mapping the FeO distribution on the surface of Venus from the emissivity in the NIR range. The classification map derived from NIR observations will complement the radar reflectivity and radiothermal emissivity derived from Magellan and Pioneer datasets. Additional information on the nature of the surface material, will be provided by the
bistatic radar VeRA and the dielectric measurements will be compared with previous results from the Magellan bistatic experiment. The figure shows the surface targets and area that would be covered by these investigations.

Figure. The Magellan map of Venus with the surface targets and area that would be covered by the Venus Express observations: red circles – bi-static radar, magenta ovals – gravity experiment, hatched area – sounding of the lower atmosphere and the surface in the near-IR transparency “windows” on the nightside.

Thanks to the complementarities between the Venus Express and Magellan observations, some critical aspects of the evolution of the planet will be better assessed in the coming years.

MORPHOLOGICAL AND TOPOGRAPHICAL CHARACTERISTICS OF EPONA CORONA: CASE STUDY OF A MULTIPLE CORONA ON VENUS. T. Törmänen1, V.-P. Kostama1, M. Aittola1, and J. Raitala1,  
1Planetology Group, Div. of Astronomy, Dept. of Physical Sciences, P.O. Box 3000, FIN-90014 University of Oulu, Finland (terhi.tormanen@oulu.fi)

Introduction: Coronae are large volcano-tectonic structures with concentric and/or radial structures and associated volcanic features [e.g. 1-4]. Coronae are proposed to form as a result of buoyant mantle diapirs deforming overlying lithosphere and several competing geophysical models have been proposed [e.g. 3-8]. We have conducted a survey [9] of multiple coronae (coronae with at least 2 linked structures with a common annulus [3,5]) from Magellan data. Currently we include 67 coronae into the multiple coronae population (3 features were left out after closer examination compared to our original survey). Of the multiple coronae 47 are Type 1 and 20 Type 2 coronae (Type 1 and 2 as defined in [10]).

Here we report results from geologic mapping of Epona Corona (28°S, 208.5°E) as part of our mapping of a sample of the multiple coronae. Mapping was based on full-resolution Magellan F-MAPs and topographic data.

Epona Corona (355x225 km; Fig. 1) is a Type 1 two-part structure located on Wawalag Planitia on a regional slope from W and N towards E and S. Epona is a located on a S-N trending chain of 2 coronae and 3 arachnoids [11]. We classify it as Class A multiple corona [9]. The western and eastern parts of Epona Corona are here called Epona West (EpW) and Epona East (EpE), respectively. The shape of EpW is ovoidal with a more linear southern rim, which has the same strike as the linear NW and SE sides of EpE. There is also a circular corona located just NE of EpE (center at 26.6°S, 212°E; CC in Fig. 1a), which we think is not part of Epona. It has a low interior and part of a topographic rim to the east. The rim and associated structures of Epona appear to overlie and deform this corona indicating that it is older than Epona, or that at least major part of its evolution predated formation of Epona.

Topography: Topographically Epona belongs to group 3b [7] (rim surrounding interior dome), and both EpW and EpE have roughly the same characteristics. EpW has a ~150-250 m-high rim along its W/NW annulus (all heights are relative to adjacent plains unless otherwise stated), which continues south around the EpW becoming lower and then rising into a ~400-600 m high rim along the southern edge of EpW. Interior to the rim there is a topographically lower band, which also defines the topography of the N side of EpW. Along the southern rim this lower band becomes a 100 m low trough, which extends into the southern EpE. The interior of EpW is at about the same level as the plains to north of EpW and ~200 m higher than adjacent plains to S. There is also a higher domal rise (DR in Fig. 1a) superposed on the N low, which appears to be a volcanic center. Between the W and E parts of Epona there is 100-400 m high broad ridge, which continues as a 150-400 m high rim around the NW, NE and most of the SE side of EpE. The southeastern topographic rim decreases in height from NE to SW. The interior of the EpE is located 200-300 m lower than adjacent plains but has a broad ~200 m central high with superposed volcanic domes.

Structures and Units: EpW has concentric fractures on its rim (a in Fig. 1a) formed probably due to tension of the upwarping crust when rim started to form. There are some radial fractures (b), which cross-cut the W/SW section of the annulus. These fractures seem to result from tensional stretching of surface. There are also some very narrow and long fissures (c), which may be surface expressions of dikes [e.g. 2,8,12,13]. Sub-parallel fractures (d) flank the linear trough paralleling the southern topographic rim of EpW. These fractures are interpreted to be graben formed due to extension. The interior of the EpW is characterized by two types of plain-like surfaces with stratigraphically younger radar-dark smooth inner unit (e) characterized by small volcanic domes and pits.

The stratigraphically oldest unit within EpE is densely fractured terrain (f) located near the base of the rim and as scattered inliers within EpE. The fractures in this terrain are probably surface expressions of dikes [8]. This terrain is covered by at least two material units (g, h), which are interpreted to be volcanic in origin (there are several volcanic cones and domes in the area). Rigdes (i) in the NE side of the EpE are interpreted to be contractional folds formed at the same time as the topographic rim.

At the southern slope of the southeastern rim of EpE there are narrow fractures (j) arranged in a left-stepping en echelon formation. The fractures form a belt that fans out into a wider area of more parallel structures around 28.62°S, 210.6°E (k). These fractures, at least the en echelon ones, may be tension gashes formed in a right-lateral shear zone.

There is a set of regional N-S trending wrinkle ridges (l), which deform the plains S of Epona Corona and continue through Epona to N. They wrap
around the SE topographic rim of EpE but do not seem to be disturbed by the topographic rim or trough of the EpW, although they appear to be cut by the widest and apparently youngest fractures at the sides of the southern trough. This indicates that when the wrinkle ridges started forming due to roughly E-W oriented compressional stress, the SE rim of EpE existed (as evidenced by the interaction with topography [14]), but the southern rim and trough of EpW may have been less pronounced than now or that their topography has formed after the formation of the wrinkle ridges (or partly at the same time).

**Sequence of Events:** The inferred sequence of events from observable units and structures at Epona appear to have been: 1) formation of the oldest radial fractures of EpW, 2) formation of densely fractured terrain in EpE, 3) formation of W/NW rim and its fractures and the southern topographic rim of EpW, 4) material covering interior of EpW and possibly concurrent formation of the volcanic unit embaying fractured terrain within EpE, 5) continued formation of radial to sub-parallel fractures cutting part of the W rim of EpW, 6) formation of the topographic rim of EpE and ridges at the eastern end of EpE, 7) formation of the wrinkle ridges, 8) fracturing along outer edge of the SE rim of EpE producing also the *en echelon* fractures perhaps due to extension and shear linked to 9) formation of the S trough of the EpW with faulting along its northern wall and on the crest of the S rim (due also to perhaps further uplift of the rim), and fracturing along the northern side of the NW topographic rim of EpE (m), 10) flow units on the lowest parts of the EpE interior and volcanism in the EpW. Phases 7) to 9) may have partly overlapped in time, because changes in the strike of some wrinkle ridges appear to be controlled by pre-existing fractures within EpW and N of central Epona Corona.

**Conclusions and Future Work:** The sequence of events does in general agree with the diapir model of corona formation [e.g. 2-6]. Whether the interior dome and trough/rim formation require delamination [7] is not certain in this case. Dome topography may be produced by accumulation of volcanic materials and edifice building rather than as a consequence of delamination [7]. The two parts of Epona Corona appear to be produced by two close diapirs, which may have started deforming the crust at about the same time or the eastern one slightly later. It appears that the EpW has either relaxed further or did not produce as pronounced topography as the EpE. The formation of the S trough and associated faulting appears to be a late stage event due to changes of stress field either regionally or more locally (i.e. principally related to corona evolution).

We are continuing mapping multiple coronae of different morphologic and topographic types to constrain the relative timing of formation of the multiple corona parts and models for their formation.

**References:**


**Figure 1.** a) Left-looking Magellan image of Epona Corona (from C1-MIDR 30S207:1). Letters refer to structures and features discussed in text. b) Perspective view of Epona from SW. Vertical exaggeration is 25.
POSSIBLE ROLE OF THE TYPE Ia SUPERNOVA EXPLOSION IN FORMATION OF THE SOLAR SYSTEM; G. K. Ustinova, Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, Moscow 119991 Russia; e-mail: ustinova@dubna.net.ru

Introduction: The solar system is a valuable, rich astrophysical laboratory because it represents the final results of wide-ranging nebular processes leading to formation of stars with planetary systems. The main and the most important link of investigations is the isotopic and elemental composition of the primordial matter. The contemporary level of knowledge is based on the conception that the primordial matter is founded on the matter of the giant molecular gas-dust nebula, which was distributed homogeneously with the products of the nucleosynthesis of about ten of supernovae during ~10 Ma of its existence before turbulent dissipation. However, the numerous isotopic anomalies in meteorites, which are conditioned by the decay of some extinct radionuclides, suggest that, at least, one of the supernovae had been exploded shortly before the collapse of the protosolar nebula. The matter of the last supernova cannot yet be mixed evenly with the matter of the molecular cloud: it rather flowed round, surrounding the cloud gradually [1]. The intervals of formation of the short-lived extinct radionuclides of $^{41}$Ca, $^{26}$Al and $^{53}$Mn testify to the last supernova explosion about 1 Ma before the formation of the solid phases of the primordial matter [2]. The absence of heavier extinct radionuclides (the products of $r$-process) with the similar intervals of formation indicates that the last supernova was the type Ia supernova (the so called, carbon-detonation supernova), which could not survive the carbon explosive burning and was fully disrupted [2,3]

Type Ia Supernovae (Sne Ia) turned out to be the ideal objects for estimating the Hubble constant, and so, the properties of these supernovae are being studied intensively over last years [4]. For the most part, Sne Ia are the explosions of white dwarfs, consisting of carbon and oxygen, that have approached the Chandrasekhar mass, $M_{\text{Chan}} \approx 1.39 M_\odot$. Just because of about equal mass and composition, the Sne Ia are considered as “standardizable candles”: their calibrated light curves have become a major tool to determine the cosmological expansion rate, its variation with look-back time and also the geometrical structure of the universe [4,5 et al.]. At the explosive burning of carbon and oxygen, releasing the energy of ~10^{51} erg, all the intermediate-mass elements like Mg, Si, S, Ca are synthesized, but the doubly-magic nucleus $^{56}$Ni is produced with the highest probability. During 6.1 days it decays to $^{56}$Co, which, in its turn, decays to $^{56}$Fe over 77 days. Thus, depending on the initial C/O composition of white dwarfs, the Sne Ia explosion produces 0.4-0.6 $M_\odot$ of $^{56}$Fe, substantial amounts of the intermediate-mass elements, and, besides, some unburned amounts of C+O can be ejected [6,7 et al.]. The peculiarities of the Sne Ia may be suggested to specify the uniqueness of the solar system among the other forecasting planetary systems. At least, three important aspects may be pointed out.

Heterogeneity of Accretion: The current concept is that the chemical and isotopic composition of the primordial matter was similar to the contemporary solar one [8] in many respects. It is based on the supposed absence of essential fractionation at the collapse stage of the molecular gas-dust clouds during the formation of the planetary systems [9]. However, the Sne Ia explosion just before the origin of the solar system, especially, if its matter surrounded the collapsing nebula, led to rather a heterogeneous accretion: the specific matter of Sne Ia (e.g., the ratio of Fe to Si-Ca is by about a factor of 2-3 higher in the solar composition [7]) had accreted mainly at the concluding stage of the formation of the solar system, enriching the latter with iron, intermediate elements, as well as with unburned C and O. It looks tempting that just the enrichment of the upper mantle of the earth with the unburned C and O has provided the basis for some pre-biological background and subsequent origin of life on the earth.

Rigidity of the Energy Spectrum of Particles: The Sne Ia explosion had established peculiar radiation conditions in the early solar system [10]. The tremendous explosive shock wave and supersonic turbulence had resulted in acceleration of particles in the cosmic plasma with forming a power-law energy spectrum $F (\gamma > E_0) \sim E^{-\gamma}$ of very high rigidity ($\gamma \rightarrow 1$) [11]. Shock waves pick up new particles from the background plasma and pump over the particles from the low energy range of the spectrum to the high energy one. That leads to the enhancement of fluxes of nuclear-active particles (and, therefore, of spallation production rates of isotopes) above the energy $E_0$ (e.g. above the threshold energy of nuclear reaction) by one-two orders of magnitude [12]. That strongly increases the share of the spallation processes in the last nucleosynthesis event of the primordial matter of the solar system. For instance, the origin of light elements Li, Be and B in spallation reactions, as suggested by Fowler in the middle of the last century [13], can in no way be achieved under the average proton fluxes of ~10^{19} cm^{-2}, forecasted for the early solar system [14]. However, the consideration of the problem in the high radiation conditions of the supernova explosion not only ensures the observed abundances of the light elements, but it also allows us to understand why $^7$Li had survived better than other isotopes [15].

The increasing rigidity of the energy spectrum of nuclear-active particles changes the weighted
of their condensation: the matter of carbonaceous chondrites was condensed under the rigid ($\gamma \sim 1.2$) radiation conditions, i.e. from the matter of Sn Ia ($\gamma$), whereas the matter of ordinary chondrites was condensed under the soft ($\gamma \sim 4.2$) ones, i.e. inside the gas-dust molecular cloud ($\gamma$) [12]. In differential meteorites some minerals, condensed under quite different radiation conditions, are observed: olivines - at $\gamma \sim 4.2$, but sulfides and phosphates in the same meteorites, as well as in the Willey iron meteorite, - at $\gamma \sim 1.2-2.5$. It is possible if accretion of the differential meteorites took place in the regions where the Sn Ia matter was blended with it of the molecular cloud, i.e. in their interface regions [12]. It is in accordance with the scenario [1] that the supernova matter surrounded the cloud.

Enrichment of the Spectrum with Heavy Ions: Another remarkable feature of shock wave acceleration of particles is enrichment of their spectrum with heavier nuclei. Indeed, the free path of multiply charged ions is an increasing function of energy: $R = p/Z e$ (where $p$ is the momentum of particles proportional to $A$, and $Ze$ is the ion charge), and the effect of acceleration depends on the ratio of $A/Z$: ions with higher $A/Z$ enter the preshock area from farther distances, and so, they are accelerated more frequently [16]. Study of air showers, induced by charged cosmic rays, shows that at ultrahigh energies ($\geq 10^7$ GeV) the energy spectrum of the particles consists of iron nuclei up to $\sim 100\%$ [17,18]. This is interpreted as an evidence for the acceleration of the primary cosmic-ray particles as fully ionized nuclei in turbulent magnetic fields (e.g. in supernova remnants). This effect is also well known in the contemporary solar cosmic rays: their SEP component (solar energetic particles) is enriched with heavy ions proportionally to $A/Q$ or $A/Q^2$ (where $Q < Z$ is the ion charge) depending on their possible acceleration in the corona (before injection into the heliosphere) or/and in the interplanetary magnetic fields [19, 20 et al]. It means that such a fractionation of particles depends also on number of the shock-wave acceleration acts. Therefore, the degree of fractionation of the matter in different isotopic reservoirs of the solar system is determined by the degree of its shock-wave reprocessing before and during its accretion in different regions of the protoplanetary nebula. For instance, Xe of the Earth and Mars atmospheres may be considered as the primordial Xe of the solar composition which had undergone five acts of shock-wave acceleration in some processes during the formation of the solar system [21]. A lot of other isotopic evidence might be found.

Possible scenario: Apart from isotopic effects in primordial matter, Sn Ia explosion opens, apparently, a clue to the origin of iron meteorites: the large quantity of synthesized and accelerated iron nuclei were the first that penetrated into the collapsing protoplanetary cloud and, being captured by supersonic turbulence, they became centers of iron parent bodies. In some cases of especially huge explosion the captured iron nuclei underlay the metallic cores of some planets, which further were built up due to magmatic differentiation. The intermediate and light nuclei of Sn Ia explosion reached the accreting system rather later. Because of the turbulence drawing into the central plane of the accretion disk, they had played a key role in formation of the earth group planets under the reducing conditions being typical for such heliocentric distances. But the most part of the unburned C of the exploded Sn Ia had accreted at the conclusive stage of accretion in the conditions of low temperatures and free gravitation that had provided the formation of carbonaceous chondrites.

**Introduction:** As it is known, earthquake arises at sudden energy clearing which have been collecting for a long time as a result of tectonic processes in located zones of the earth's crust and the top cloak. Thus occurs break of rocks, sometimes on many tens kilometers [1].

A line of correlations is known that connects seismicity with parameters of the Earth: in the speed of rotation of the Earth, a magnetic field, etc. [2]. Considering seismicity of the Earth is possible only together with its model of formation, evolution and the internal device [3].

The scale of magnitudes defines standard earthquake and estimates other earthquakes on their maximal amplitudes concerning this standard scale under identical conditions of supervision.

The magnitude of earthquakes $M$ by Richter's definition:

$$ M = \lg \left( \frac{A}{A_0} \right), $$

where $A_0$ and $A$ are the maximal amplitudes of record on a certain seismograph for standard and measured event accordingly.

The magnitude is connected with the energy of an earthquake. Change of the magnitude on a unit is equivalent to increasing (downturning) of the energy of an earthquake in 32 times.

There are 11-years cycles of seismic activity of the Earth, which have essential negative correlation with the cycles of solar activity (with the cycles of solar spots) [1]. During the 11-year solar cycle seismic activity grows during the minimal solar activity and during large solar flashes. With duration cyclic changes of geomagnetic and seismic activity are allocated into three solar cycles also.

Under influence of powerful solar proton events there is a transition of power processes inside the Earth from one condition in another which is kept on an extent approximately three solar cycles before the next large flashes. This level of seismic activity of the Earth defines values of the energy clearing at earthquakes for all this period [4].

In the given work for precision processing of some seismic activity the approach of dynamic regression modeling (DRM-APPROACH), presented as a software - modified in comparison with [5, 6] the automated system DRM, was used.

Thus construction of complex model of the earthquakes including trend, harmonious or autoregression components and providing smoothing of noise by a method of martingale approximations [5] was supposed to be made.

**Construction of models and their analysis:** As the initial data 108 values of the magnitude, calculated by the International seismological center (ISC) with monthly averaging for 1996-2004 (http://www.isc.uk) are taken.

At the first stage of application of device DRM for the analysis of time lines check on stationarity was carried out therefore by nonparametric criterion of shift with confidential probability 0,95 the hypothesis about shift of average downwards was accepted; by criterion of dispersion the hypothesis about equality of dispersions of the first and second group is accepted. That’s why, it is possible to count the line approximately stationary.

At the following stage the trend of kind $Y=A\cdot B^t$ was allocated; $Q_{AQD}=0,6363$ have been distributed; $Q_{AQD}$ on external accuracy $\sigma = 6,48$; then the analysis of the rests from a trend is carried out with the purpose of revealing the periodic component.

Autocorrelation function (Fig.1), Darbin-Watson factor (equal to 0,0083), the spectral analysis of the rests (Fig. 2) and the wavelet-analysis of the rests specify presence of autocorrelation that assumes allocation of harmonious or autoregression components.

![Fig. 1. Autocorrelation function of some seismic activity of the Earth](image1)

As a result of the spectral analysis the following data were allocated: a harmonic '1' with period $T = 2$, 

![Fig. 2. The spectral analysis of seismic activity of the Earth](image2)
an amplitude \( A = 5.9812 \times 10^{15} \), a phase \( \phi = -2.5916 \times 10^{-16} \); a harmonic \( t'2 \) with period \( T = 3 \), amplitude \( A = 0.037045 \), a phase \( \phi = 124.83 \); a harmonic \( t'3 \) with \( T = 4 \), \( A = 0.085097 \), \( \phi = 158.22 \); a harmonic \( t'4 \) with \( T = 5 \), \( A = 0.11285 \), \( \phi = 24.046 \); a harmonic \( t'5 \) with \( T = 8 \), \( A = 0.085484 \), \( \phi = 172.94 \); a harmonic \( t'6 \) with \( T = 10 \), \( A = 0.21216 \), \( \phi = 37.722 \); a harmonic \( t'7 \) with \( T = 14 \), \( A = 0.075976 \), \( \phi = 148.09 \); a harmonic \( t'8 \) with \( T = 22 \), \( A = 0.086736 \), \( \phi = 46.363 \). AQD=0.3535, \( \sigma_\Delta = 6.43 \).

On a correlation matrix of harmonics with the set periods some dependence between harmonics with the periods of 2 and 8, 2 and 4, 2 and 22 months has been revealed. Exception of a harmonic with the period 2 months has insignificantly improved quality of model on AQD =0.3517; \( \sigma_\Delta = 6.41 \).

The filtration of lines was carried out by a method of martingale approximations of the second degree; in result the model with AQD =0.049 is received; \( \sigma_\Delta = 6.3 \).

Thus, the model describing behaviour of some seismic process for 1996-2004 is received; its graphic realization is submitted on Fig. 3, and the forecast of seismic activity of the Earth for 24 months - on Fig. 4.

![Fig. 3. Schedules of some seismic activity and the received model](image)

![Fig. 4.](image)

While analysing seismic activity of the Earth it is not difficult to notice the connection with solar activity. The considered line has essential negative correlation with the line of solar activity. The maximal earthquakes are observed during the minimum of solar activity or during the periods close to a minimum, and, on the contrary, in maxima of solar activity seismicity of the Earth accepts the least values. In the examined period three maxima of seismic activity are found out. The first, the strongest, falls at years of a minimum of solar activity, and others two - on a growth phase and recession of solar activity accordingly, that is for the period when there is the greatest number of large solar proton flashes.

Thus, in time dependence of annual values of earthquakes there are the periods caused by geomagnetic activity of the Earth, and the periods of smaller duration caused by solar flashes.

**The conclusion:** As a result of processing lines the model describing dynamic of changes of seismic activity is allocated optimum on measures of quality. The forecast of lines for 24 months is constructed.

Works on the further updating of DRM [7] are conducted. A number of lines of geophysical and geophysical characteristics are processed. Data treatment of dynamics of baricentre of the systems the Ground - moon is planned with the purpose of an establishment of dependence, modeling of dynamics of change of lines and the forecast.

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**References:**

GEOLGY AND STRATIGRAPHY OF SATURN'S SATELLITE DIONE OBSERVED BY THE CASSINI ISS CAMERA. R. Wagner (1), G. Neukum (2), T. Denk (2), B. Giese (1), T. Roatsch (1), and the Cassini ISS Team. (1) Institute of Planetary Research, German Aerospace Center (DLR), D-12489 Berlin, Germany (E-mail: Roland.Wagner@dlr.de), (2) Institute of Geosciences, Freie Universitaet Berlin, D-12249 Berlin, Germany.

Introduction: Over two decades ago, the cameras aboard the two Voyager spacecraft imaged Saturn's satellite Dione (1124 km in diameter) at spatial resolutions of at least 1 km/pixel. Its surface is characterized by (1) cratered plains with varying crater frequencies and ages, (2) smooth plains which were believed to be volcanic extrusions of an H2O-NH3 eutectic melt, and (3) tectonic features such as scarps, troughs and ridges [1][2][3]. These features which are indicative of extension as well as compression were believed to be a consequence of episodes of global expansion and contraction through time [3][4]. The trailing hemisphere, imaged only at low resolution (> 5 km/pixel) by Voyager, shows a system of very bright, filament-like linear markings termed wispy material [2]. It was interpreted as a surficial deposit associated with volcanic exhalations along cracks [1][2].

Image data base: Since the Cassini Orbiter has been inserted into an eccentric orbit around Saturn on July 1, 2004, image data at resolutions between 1.4 km/pixel and 430 m/pixel were obtained by Cassini ISS (Narrow (NAC) and Wide Angle Cameras (WAC)) in two non-targeted flybys (Dec. 2004 and June 2005). Areas not very well covered by Voyager were imaged in these encounters showing more or less unknown terrain in much better detail (e.g. Fig. 1).

Main topics and procedure: The basic questions addressed in this study are: (1) What are the geologic units seen in the areas imaged by Cassini ISS, what is their stratigraphic sequence, and how do they compare to the units mapped on Voyager data? (2) What are the ages of these units, obtained from crater size-frequency measurements and from application of impact chronology models? Is there a Crater Population I & II as has been suggested [1]? (3) Is there evidence for past cryovolcanic activity? (4) What is the nature of the wispy terrain on the trailing hemisphere? Ages are assigned by impact cratering chronology models. Two such models are currently used: (a) Model I with a lunar-like (exponential) decay in impact cratering with time (Late Heavy Bombardment (LHB)), and with a more or less constant cratering rate since about 3 b.y. (billion years) [5][6][7], and (b) Model II with a constant cratering throughout most of solar system history [8].

Results: (1) heavily cratered plains are the spatially most extensive and oldest units on Dione, confirming Voyager results [1][2][3]. Degraded tectonic features indicate early episodes of tectonism. Unexpectedly, there are no old, degraded impact basins as seen on other icy satellites, such as Rhea, Iapetus, or the Galilean satellite Callisto [9][10][11]. Model ages of this oldest unit are either higher than 4 b.y. (Model I) or higher than 2.5 b.y. (Model II). (2) Resurfacing has been caused by tectonism rather than by cryovolcanism. There is no evidence for flows or pyroclastic deposits on Cassini ISS data so far. (3) Also, the wispy material has been shown to be of tectonic rather than of cryovolcanic origin. Light ist scattered from numerous fault scarps. Several episodes of extensional tectonism (with an unknown contribution of shear) can be recognized. According to crater size-frequency measurements and model ages, tectonic episodes may date back to > 3.7 b.y. (Model I) or > 1 b.y. (Model II). (4) Less densely cratered plains in many cases are associated with younger craters, basin(s) and their ejecta. Since there are stratigraphically younger large craters, there is no conclusive evidence for two different crater populations I (heliocentric projectiles during LHB creating craters > 20 km in diameter) and II (smaller, post-LHB Saturno-centric projectiles) as suggested by [1]. (4) Only one (unnamed) basin with a diameter of about 400 km was discovered so far (see Figs. 1, 2). This basin is stratigraphically young. Model ages are on the order of ~3.2 b.y. (Model I) or only ~0.33 b.y. (Model II). (5) The youngest units on satellite surfaces are generally associated with bright ray craters. Such features have not yet been observed on Dione. One feature named Cassandra was presumed to be a ray crater but turned out to be actually a set of radial scarps radiating away from a point source, exposing bright ice on the scarp slopes.

Future work: Upcoming targeted flybys during the Cassini Mission (one targeted flyby will take place on Oct. 11, 2005) will provide extensive high-resolution and stereo image coverage which will help to investigate small-scale surface details and tectonic landforms. Also, a comparative investigation of Dione and its outer neighbour satellite Rhea is of importance since Rhea also shows wispy markings on its surface which could not been seen in detail so far. These features most likely also originate from tectonic stress rather than from cryovolcanism.


**Fig. 1:** Part of the trailing and southern hemisphere of Dione. Image shows cratered plains, degraded craters, tectonic features, and a large basin near the south pole with a diameter of 400 km. The south pole is located to the west of the basin. The single trough going from top to bottom approximately follows the 270° (West) longitude. False-color representation using the hue-saturation-intensity color transformation.

**Fig. 2:** Detailed view of the unnamed basin in the south polar region. The basin possibly has two rings, but the inner ring is very poorly seen.

Introduction: Explosive volcanic eruptions are predicted to have occurred on Mars during its geologic history. We focus attention on the formation of a distinctive type of explosive volcanic product, accretionary lapilli, whose recognition in the geologic record during detailed surface exploration could provide important information about the nature of the volcanic history of Mars. Most treatments of the formation of fall deposits from volcanic eruption clouds on Mars have assumed that individual clasts were transported through, and fell from, such clouds as discrete objects with negligible mutual interactions. However, it is well established that the formation of accretionary lapilli, a mechanism that allows small particles to fall much more rapidly as members of clusters than as individuals, is an important factor in determining the spatial thickness and grainsize variations in fall deposits on Earth. We therefore explore the nature of this process in the martian environment.

Gilbert and Lane [1] showed that accretionary lapilli form in explosive eruption clouds when initially small particles grow by becoming coated with water condensing in the eruption cloud and accreting other small particles which they overtake as they fall. The overtaken particles become trapped in the water layer, and chemical processes bond the accreting particles to one another and ensure the stability of the growing structure. The critical factors controlling the growth of a lapillus to a final size \( D \) (assumed to be much greater than its initial size) are the vertical distance through which it falls, \( Z \), and the mass loading in the eruption cloud, i.e. the total mass of pyroclasts per unit volume, \( \nu \). Other important factors are the density of individual ash particles, \( \rho_p \), the porosity of the aggregates, \( \phi \), the density of the liquid or gas in the pore spaces, \( \rho_g \), and the aggregation coefficient, \( E \), combining the efficiency of sticking and the probability of collision, which accounts for the extent to which a small particle attempts to follow streamlines in the gas and be swept around any larger clast that overtakes it. The relationship is \( D = (0.5 \nu E Z) / \rho_p \) (1), where the lapillus bulk density \( \rho_b \) is equal to \( \rho_p \phi (\rho_p - \rho_g) \). Two conditions are implicit in the assumptions behind this model. The first is that large pyroclasts fall through an eruption cloud faster than smaller ones. This will be true as long as the atmospheric pressure is large enough to ensure that particles interact with the gas according to the gas laws, falling with terminal velocities that are a function of their size. A comparison of the mean free paths of molecules in the martian atmosphere with the ranges of lapilli sizes considered here shows that it is only at heights greater than ~50 km on Mars that mean free paths exceed particle sizes so that the Knudsen regime becomes operative and all particles move at the same speed. This is not a limitation, because we assume that martian eruption clouds rarely rise to heights much greater than ~20 km. More important is the fact that a liquid water film is required on the colliding clasts to aid adhesion. This is common over a wide range of heights in eruption clouds on Earth, both where clasts are rising above the vent and where they are falling at greater lateral distances. However, it is more problematic on Mars because of the low (everywhere <240 K; [2]) atmospheric temperatures. Once a significant amount of atmosphere has been entrained into an eruption cloud on Mars any water is likely to be frozen to ice. However, we show that temperatures will be above the freezing point in much of the rising part of an eruption cloud, and we infer that it is here that most lapilli formation occurs. The fact that the relative motion of large and small clasts is superimposed on the high-speed turbulent motion of the rising part of the eruption cloud, rather than on the lower-speed, more nearly laminar motion of the atmosphere outside the rising core of the cloud, as on Earth, does not hinder the formation process and, indeed, may enhance it by providing an effectively greater path length over which particle collisions can occur.

Experimental data [1] show that \( E \) in equation (1) is ~1 for very small adhering particles and decreases from ~1 to ~0.1 as the adhering particle size increases from ~10 to ~100 \( \mu \)m. \( \rho_p \) is taken as 1500 kg m\(^{-3}\) on Earth and Mars, and is ~50% greater than \( \rho_g \) when water fills pore spaces and is very much greater than \( \rho_g \) when gas fills them. As a compromise we assume that pore spaces are half filled with water. \( \phi \) is generally ~0.4 for lapilli on Earth and there is no reason to expect values to be greatly different on Mars, so that \( \rho_b = \phi (\rho_p - \rho_g) \approx 1100 \text{ kg m}^{-3} \). Application of eruption cloud models developed for the Earth to Mars suggests that such clouds should rise much higher in the atmosphere of Mars for a given mass eruption rate [3-5]. However, Glaze [6] pointed out that some of the assumptions made in these models about the mechanism of entrainment of atmospheric gases are not justified above ~20 km in the atmosphere of Mars, so that high discharge-rate eruptions generating clouds that might have been expected to convect to much greater heights may instead produce structures more analogous to the umbrella-shaped plumes on Io, where gas-particle interactions are minimal except near the surface. We therefore adopt 20 km as the maximum likely value of \( Z \) on Mars. Well above the vent in eruption clouds on Earth, \( \nu \) ranges from ~2 \times 10^{-3} to 10 \times 10^{-3} \text{ kg m}^{-3}. The major difference between explosive eruptions on Mars and Earth is that the volatiles exsolved from the magma must eventually decompress by a much greater factor to reach equilibrium with the atmosphere on Mars than on Earth. Enhanced gas expansion implies that a given mass, and hence a given number, of pyroclasts will occupy a greater volume on Mars than Earth, leading to a smaller value of \( \nu \), which we now estimate.

It is likely that the vast majority of eruptions on Mars take place under conditions where the vent cannot flare outward toward the surface sufficiently rapidly to allow the pressure in the erupting jet of volcanic gas and entrained pyroclasts to decrease to the atmospheric pressure. Instead the eruption is choked, with the pressure in the vent being the value at which the velocity of the gas-pyroclast mixture is equal to the speed of sound in that mixture. Above the vent the pressure decreases through a complex series of shocks to reach the ambient atmospheric value; there is some lateral expansion of the jet, and the upward velocity of the gas and clasts increases before beginning to decrease again as the upward momentum of the erupted mixture is shared with entrained atmospheric gases.
The speed of sound, $U_s$, in the gas-pyroclast mixture is given by $U_s^2 = \left[ \frac{n Q T}{m} \right] \left[ 1 + \frac{\left( (1 - n) m P_s \right)}{\left( n Q T \rho_m \right)} \right] (2)$, where $n$ is the exsolved mass fraction of the major volatile in the magma (here assumed to be water vapor with molecular weight $m$ equal to 18.02 kg/kmol), $Q$ is the universal gas constant, 8314 J kmol$^{-1}$ K$^{-1}$, $T$ is the temperature of the eruption products, taken as 1450 K for a mafic magma, $\rho_m$ is the density of the magmatic liquid, taken as 2700 kg m$^{-3}$, and $P_s$ is the pressure in the choked flow in the vent.

The speed of the erupting mixture is determined by the amount of expansion of the magmatic gas between the level at which the magma disrupts into pyroclasts and the vent. Using the common assumption that little gas exsolution occurs between the fragmentation level and the surface, and that fragmentation takes place when the volume fraction of gas bubbles in the magma exceeds a critical value of order 0.75, we find that the pressure at the fragmentation level is $P_f$ where $P_f = \frac{[n Q T \rho_m]}{[3 (1 - n) m]} (3)$. Equating the energy obtained from the decompression of the erupting mixture between the pressures $P_f$ and $P_c$ to the kinetic energy of the eruption products gives $[7] U_c^2 = \frac{2 \left[ n Q T \right]}{m} \ln \left( \frac{P_f}{P_c} \right) (4)$. The choked condition requires equating $U_c$ to $U_s$ at the vent, so that $\left[ 1 + \frac{\left( (1 - n) m P_s \right)}{\left( n Q T \rho_m \right)} \right]^2 = 2 \ln \left( \frac{P_f/P_c}{P_s/m (1 - n) (P_f - P_c)} / \left( n Q T \rho_m \right) \right) (5)$. Given any choice of $n$ and hence $P_c$, this equation can be solved recursively by inserting an initial estimate of the value of $P_c$ (one half of $P_f$ is appropriate) into the right-hand side and solving the equation to obtain an improved estimate of $P_c$. After an adequate level of convergence has been obtained, either of equations (2) or (4) can be used to obtain the mean eruption speed $U_c$.

Above the vent, a series of shocks and expansion waves within the volcanic jet allows the pressure in the vent, $P_v$, to relax to the atmospheric pressure, $P_a$, over a vertical distance of at least several vent diameters. This process has been explored theoretically for expansions into a vacuum or into a crater-like structure around the vent, but is not well-studied for other geometries, especially where eruptions take place into an atmosphere which must eventually begin to be expanded into the jet. Nevertheless, a reasonable approximation to the amount of energy available from the decompression from $P_v$ to $P_a$ is $\Delta E$ given by $\Delta E = \frac{\gamma}{(\gamma - 1)} \frac{\left[ n Q T \right]}{m} \left[ 1 - (P_f/P_s)^{\frac{\gamma - 1}{\gamma}} \right] (6)$, where it has been assumed that the expansion and cooling of the gas-pyroclast mixture from its eruption temperature $T$ can be treated as the adiabatic expansion of a pseudo-gas having a ratio of specific heats $\gamma$ given by $\gamma = (s_{sp} + K_s) / (s_{sv} + K_s) (7)$ where $s_{sp}$ and $s_{sv}$ are the specific heats at constant pressure and constant volume, $\sim 3900$ and $\sim 2800$ J kg$^{-1}$ K$^{-1}$, respectively, of steam, $s_c$ is the specific heat at constant volume of the silicate matrix, $\sim 1000$ J kg$^{-1}$ K$^{-1}$, and $K_s$ is defined as $K_s = \left( 1 - n \right) / n (8)$. The energy $\Delta E$ must be added to the kinetic energy of the eruption products at the vent level to obtain their final velocity, $U_{vc}$, after reaching atmospheric pressure from $0.5 U_s^2 = \Delta E + 0.5 U_{vc}^2 (9)$.

Equations (2) to (5) involve only the magma water content, the magma temperature, and the physical properties of the water vapor, and so for any given magma, the values of $P_v, P_c$, and $U_c$ will be the same for choked eruptions on both Mars and Earth. However, the atmospheric pressure is involved in calculating the final velocity via equations (6) to (9) and so it is here that a significant difference appears between eruptions on Mars and Earth.

The value of $w$ is determined by the amount of expansion of the volatile phase, in this case water vapor, in the decompression region above the vent. The ratio $R$ of the volatile volumes before and after adiabatic expansion is given by $R = \frac{P_v/P_a}{\gamma - 1} (10)$ and is thus different on Mars and Earth, again because of the differing atmospheric pressures. The decompression is partly accommodated by the increase in gas velocity from $U_c$ to $U_{vc}$ and partly by the increase of the cross-sectional area of the erupting jet, and it is this area change that is directly reflected by $w$. The factor $A$ by which the area increases, and hence $w$ decreases, is given by $A = \frac{R}{(U_{vc}/U_c)} (11)$. The factor by which $w$ is smaller on Mars than on Earth is equal to $A_{3sp}/A_{3sv}$, and since the largest value of $w$ for terrestrial eruption clouds is $\sim 10 \times 10^3$ km s$^{-1}$, we divide this value by $(A_{3sv}/A_{3sv})$ to obtain the values of $w_m$ for Mars for each value of $n$. Finally, inserting these values of $w_m$ into equation (1), together with the values of the other parameters discussed earlier as being relevant to Mars, gives the values of the largest likely lapilli sizes, $D$. Values vary between $0.7$ and 0.9 mm. The smallest lapilli sizes are likely to be smaller than these values by at least a factor of 10, and we note that all of the lapilli sizes are about one order of magnitude smaller than those commonly found on Earth.

The dispersal of accretionary lapilli with sizes that grow from $\sim 0.1$ to $\sim 1$ mm while falling through the martian atmosphere from heights of $\sim 20$ km will depend on the wind regimes that they encounter while falling and on their terminal velocities. Wind speeds (dominated by zonal winds) are given for northern and southern summer conditions and for low and high dust loading [2], and typical conditions can be represented adequately by a wind profile that decreases roughly linearly from $\sim 40$ m s$^{-1}$ at 20 km to zero at ground level. Clast terminal velocities depend on clast size ($d$, the diameter of an equivalent sphere) and bulk density $\rho_m$, and on the density $\rho_s$ and viscosity $\eta$ of the atmosphere, both functions of height. Using the model in [2] we find the pressure, temperature, density and viscosity values, and the terminal fall velocities $U_t$ for several sizes of particles with density $1100$ kg m$^{-3}$. These terminal velocities are calculated for the relevant regime (laminar or turbulent) in which the particles fall by taking $U_t$ to be the smaller of $U_t = \frac{A_d D g / \left( 18 \eta \right)}{\left( 12 \right)}$, $U_t = \frac{\left( 4 \pi \rho_s D \right) / \left( 3 \eta \rho_m \right)}{\left( 12 \right)}$ (13), where $g$ is the acceleration due to gravity, $\sim 3.72$ m s$^{-2}$, and $C_d$ is a dimensionless drag coefficient that depends on the shapes of particles but is of order unity. Using these values, incremental fall times between successive heights in the atmosphere are calculated and the windspeed $W$ is used to find the lateral displacement while traversing each height increment. These distances are then summed to give the lateral displacements while falling from 20 km. It is clear that the extent of dispersal of an accretionary lapilli fall deposit from an eruption cloud on Mars could range from several tens to many hundreds of km depending on the final particle size.

POSSIBLE REDUCTION OF SLIGHTLY SIDEROPHILE ELEMENTS IN IMPACT PROCESS  |  O. I. Yakovlev1, Yu. P. Dikov2, M. V. Gerasimov3. 1Vernadsky Institute of Geochemistry and Analytical Chemistry, RAS, Moscow 117975, GSP-1, Kosygin St., 19, yakovlev@geokhi.ru , 2Institute of Ore Deposits, Petrography, Mineralogy and Geochemistry, RAS, Moscow 109017, Staromonetny per., 35, dikov@igem.msk.su , 3Space Research Institute, RAS, Moscow 117810, Profsoyuznaya st., 84/32, mgerasimov@mx.i.ki.rssi.ru

Introduction: V and Cr are lithophile elements. Nevertheless, the depletion of these elements in the crust by possible sink into the core infers their siderophile behavior during the Earth’s accretion and provides classification of V and Cr as slightly siderophile elements. The problem of geochemical behavior of these elements is tightly connected with the definition of specific conditions during accretion which had induced metallization of elements. In a series of our former experimental works [1, 2] we have formulated an idea that the main reason of metallization of iron and other siderophile elements was their thermal reduction during impacts. Such a mechanism could provide reduction of elements in both impact melts and a vapor phase in every planetesimal impact during accretion. Here we present new experimental results which prove the possibility of impact-induced reduction of slightly siderophile elements during high-temperature impact related conditions. In a case that these reduced phases were accumulated in a core siliceous mantle of the Earth and Moon have to have depletion in these elements compared to the composition of source material.

Experimental technique: The experiments were carried out in a pulse-laser setup in the regime of a free generation of laser radiation [3]. The Nd glass laser had the following parameters: a wavelength of λ=1.06 μm, a pulse energy of ~600 J, a power density of radiation of ~10^6-10^7 W/cm^2, and a pulse time of ~10^{-3} s. Typical temperature under such condition were 3000-4000 K. Vaporization of samples was performed in helium at 1 atm. and room temperature. The laser beam was focused to a diameter of ∼3 mm. It melted and vaporized a few tens of milligrams of a sample. A metal screen was installed in the spreading path of the vapor at a distance ∼7cm from the sample. Glass tiny spherical particles (1-20 μm in diameter), which were frown out from the melted sample by expanding vapor, were found on the film of the condensate.

Starting samples in this set of experiments were prepared as pressed tablets of carefully mixed powders of peridotite with V_2O_3 and Cr_2O_3 oxides. The compositions of starting samples were (mol %): 1) Si 11.6; Fe 5.8; Mg 9.3; Ca 3.9; V 10.8; O 58.6; 2) Si 14.4; Fe 3.8; Mg 16.0; Ca 1.9; Cr 5.4; O 58.5. It was important that all elements in starting samples were in oxidized state.

Chemical analyses of glass spherules were made using PHI 660 Scanning AUGER Microprobe. Chemical analyses of the condensates were made with by XPS technique.

Results: a) Condensates. Analyses of the condensates have shown that together with metallic component of iron there are sufficient proportions of metallic V and Cr. The mean compositions of condensates were: (sample 1) Si 19.1; Fe^{2+} 3.1; Fe^0 1.9; Mg 11.8; Ca 1.5; V^{4+} 0.7; V^{3+} 3.1; V^0 0.7; O 58.2; and (sample 2) Si 18.4; Fe^{2+} 3.2; Fe^0 1.6; Mg 10.4; Ca 1.8; Cr^{3+} 4.5; Cr^0 1.2; O 59.0.

The average degree of reduction of iron in the condensate was ∼35%, and for chromium and vanadium that was ∼20% and ∼15% respectively.

b) Spherules. Analyses of spherules shows a wide diversity of their compositions, what is indicative of individual thermal history of melted droplets. Auger analyses show a pronounced deficit of oxygen in droplets. The quantity of oxygen was depleted compared to its starting value 10 to 20% in average. To estimate the degree of reduction of samples we have calculated the reduction index (RI), which was the subtraction from unity of a ratio of measured concentration of oxygen in droplets to that in starting sample. RI shows the proportion of oxygen, which is lacking for the total oxidation of elements in a spherule. This index is varying between 25 and 70% for sample 1 and between 13 and 50 for sample 2 (Fig. 1 and 2). Such high deficit of oxygen is the result of the presence of elements in metallic state in the melt. Taking into account analyses of condensed films, we consider that metallic component is mainly presented by iron, with some quantity of chromium and vanadium. Analytic evaluation of the state of silicon and magnesium in some spherules has shown that up to 15% of silicon and up to 15% of magnesium also can be present as Si^0 and Mg^0.

Conclusions: Experimental results prove the efficient thermally-induced reduction of iron and slightly siderophile elements (Cr and V) during high-temperature impact-related heating of silicates. It could be an efficient mechanism for metallization of siderophile and slightly siderophile elements during a period of impact accretion of the Earth and the Moon. Metallic components of slightly siderophile elements could be partially removed from the mantle material by their dissolution in metallic iron phase and consequent sink into the forming core providing a certain depletion of these elements.

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Fig. 1

Cr$_2$O$_3$ + peridotite

Fig. 2

V$_2$O$_3$ + peridotite

Reduction index (oxygen deficit), %

O, at.%
Influence of Surface Roughness on Scattering Properties of Wavelength-Size Particles Simulating Regolith Grains. E. Zubko¹², K. Muinonen¹, T. Nousiainen¹, Yu. Shkuratov², and G. Videen³, ¹Observatory, P.O. Box 14, FIN-00014, University of Helsinki, Finland, ²Astronomical Institute of Kharkov National University, 35 Sumskaya St., Kharkov, 61022, Ukraine, ³Department of Physical Sciences, P.O. Box 64, FIN-00014, University of Helsinki, Finland, ⁴Army Research Laboratory AMSRL-CI-EM, 2800 Powder Mill Road, Adelphi Maryland 20783, USA

Introduction: Most if not all atmosphereless celestial bodies are covered by a regolith layer consisting of dust particles with complex irregular structure. The particle size varies widely. Apparently, small submicron-sized dust particles do not exist independently of larger particles but adhere to them. Variation of the number of small and large dust particles in the regolith results in differing surface structures on the large particles. Small grains on the large particles can be interpreted as surface roughness. Here we investigate the photopolarimetric properties of irregular particles comparable to wavelength with different kinds of roughness. Note that earlier investigations on the influence of surface roughness on the light-scattering properties of spherical particles comparable to wavelength have been carried out [e.g. 1].

Initial irregular particles and generation of surface roughness: As the basic irregular particle we have chosen the so-called Gaussian random particle. The algorithm for the generation of sample particles was described in [2]. As statistical parameters, we used the relative radius standard deviation $\sigma = 0.245$ and the power law index $\nu = 4$ in the covariance function of the logarithmic radius. For such parameters, the sample particles have a smoothly undulating surface (Fig. 1, on the left).

In order to compute the light-scattering properties, we use our own implementation [3] of the Discrete-Dipole Approximation (DDA) technique [4]. The main advantage of DDA over other approaches is the absence of any restriction on the shape and internal structure of scatterers. According to the general idea of DDA, we replace each initial irregular particle with an array of dipoles which are located in cubic lattice points.

In order to generate additional roughness on the Gaussian particle surface we divided all dipoles into two classes: those belonging to the surface layer and those inside the particle. A given dipole belongs to the surface layer if the number of neighboring dipoles is less than 26. Among the surface-dipole sites, a certain number of sites were randomly chosen. Half of them were marked as seed sites for material, whereas the rest were seed sites for free space. Each of the remaining surface-dipole sites obtained the same class as the nearest seed site. The varying ratio of seed-site number to the total number of sites provides us with different kinds of roughness.

In the current investigation, 100 Gaussian sample particles have been studied. The mean numbers of dipoles inside the particle and within the surface layer are 19133 and 5483, respectively. Two sets of seed sites were used: 3000, i.e., 1500 for material and 1500 for free space (Fig. 1, center) and 300 (Fig. 1, on the right). The first and second sets of seed sites result in small-scale and large-scale roughness on the particle surfaces, respectively. All examples look realistic in comparison to real planetary regolith particles.

Results of computations and discussion: Here we present results of phase curves of intensity and degree of linear polarization for Gaussian particles with and without surface roughness. Computations were carried out for the refractive index $m = 1.6 + 0.0005i$ which is close to the refractive index of pyroxene glasses in visible light [5]. The size parameter $x = 2\pi r_{\text{eq}}/\lambda$ was varied from 2 to 12 with the step of 2 ($r_{\text{eq}}$ is the radius of the circumscribing sphere of the largest sample particle and $\lambda$ is the wavelength). On the average the Gaussian sample particles occupy only about 0.14 of the volume of the circumscribing sphere, thus the relationship between the size parameter of the circumscribing sphere and size parameter of the equal-volume sphere is given as $x_{\text{eq}} = 0.52 \cdot x$. Note that at $\lambda = 0.5 \mu m$ the actual size of the particles considered varies from 0.32 to 1.9 \mu m. Here we show the curves calculated for size parameters $x = 6, 8, 10$.

The light-scattering properties of each sample particle were averaged over orientations using no less than 12 different orientations. In some cases, this number reached the value of 60. The choice for the number of orientations was based on the necessity to obtain a statistically reliable result. For a given set of parameters, the same number of orientations was applied to all particle samples. Thus, the total number of sample particles with differing orientations was between 1200 and 6000.
In Figure 2, phase curves of normalized intensity (upper panel) and degree of linear polarization (lower panel) for unpolarized incident light are shown for \( x = 6 \). One can see rather similar intensity behaviors independently of the type of particle surface, though both kinds of roughened particles produce intensity curves more similar to one another than to those for the initial irregular particles. Contrary to the intensity, the linear polarization is more sensitive to features on the particle surfaces: the curves differ noticeably. We note that, for small phase angles, all polarization curves have significant negative polarization branches (NPB). The NPBs as well as full polarization phase curves depend on the type of particle surface—in the present case, NPBs are neutralized with increasing surface roughness.

Figure 3 presents the same as Figure 2 for \( x = 8 \). As earlier, the intensity phase curves are similar; whereas, the polarization curves strongly depend on surface roughness. In this case, however, the NPB becomes more pronounced with increasing surface roughness. Note also that, generally, both kinds of roughened particles produce polarization phase curves that are closer to one another than to that for the initial irregular particles.

In Figure 4 results are shown for \( x = 10 \). Again, only a small difference in the intensity curves is seen. Nevertheless, the curves of the roughened particles are still more similar to one another. The same similarity is observed for the polarization phase curves. Contrary to the previous cases, the NPB is practically independent of surface roughness.

**Conclusion:** Our simulations show:

1. Polarization phase curves are more sensitive to surface roughness than intensity phase curves for irregular particles having sizes comparable to the wavelength;
2. Photopolarimetric properties resulting from the two types of surface roughness are closer to one another than to those of the initial irregular particles; and
3. The behaviour of the NPB is complex. Increasing the surface roughness can make the NPB more or less pronounced.

All the obtained results are potentially useful for the development of a remote sensing technique to estimate variations of morphology of planetary regolith particles.

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**References:**