THEORETICAL MODELING OF OPTICAL MATURATION OF LUNAR AND MERCURIAN REGOLITH. L. V. Starukhina and Yu. G. Shkuratov, Astronomical Institute of Kharkov University, Kharkov, 61022. Ukraine, starukhina@astron.kharkov.ua

Introduction: Spectral measurements provide information about chemical composition of the surfaces of celestial bodies. For atmosphereless bodies, spectral information is masked by space weathering that occur due to meteoritic bombardment or solar wind irradiation and results in maturation of the surface material (regolith). This emphasizes the importance of theoretical modeling for interpretation of the spectra.

Studies lunar samples showed that space weathering results in formation of grains of reduced metals, mainly iron, in silicate particles [1]. Lunar soils of low maturity contain mostly nanograins of Fe^0 ($npFe^0$), their average diameter being 50-60 Å, located in ~1000-2000 Å-thick amorphous outer zones of regolith particles; in mature soils, especially in impact glass, $npFe^0$, as well as Fe^0-grains of sizes up to a few microns, are distributed all over the particle volume [2].

Spectral effects of these structural changes are (1) darkening, (2) reddening and (3) decrease of the depths of absorption bands of ferrous ions (Fe^{2+}) near 0.95 and 1.85 µm [3,4]. Mercury that is supposed to have silicate composition show no absorption in spectral range from 0.4 µm to 1 µm. This suggested low iron content estimated by different authors from 5.5 wt.% FeO [5] to as low as 1.2 wt.% [6]. Another possible interpretation of the Mercurian spectrum is an extreme maturity of the surface due to processes called overmaturation [7].

To answer the question if Mercurian spectrum can impose any constraints upon its surface composition, theoretical simulation of the effects of space weathering on spectra of silicates, as well as theoretical modeling of the mechanisms of maturation are required. Here we consider both aspects of the problem.

Calculation of spectral effects of space weathering: The model of spectral reflectance of multicomponent regolith-like surfaces [8,9] enables us to calculate the spectrum of mature soil starting from that of immature one (or vise versa). At first, we calculate the spectrum of the imaginary part κ of the refractive index of the soil material using its reflectance spectrum. Next, the κ spectrum is modified by adding the contribution of $npFe^0$ to κ that can be obtained from Mie theory:

$$\kappa_{Fe} = (3/2)(\Delta c)n\cdot\mathrm{Im}[(\varepsilon-1)/(\varepsilon-2)], \qquad (1)$$

where Δc is the difference in average volume concentration of $npFe^0$, n is the real part of silicate refractive index, and ε is the ratio of the complex dielectric constant of Fe^0 to that of silicate. Then reflectance spectrum is calculated for the modified κ spectrum and a new value of particle size.

The theoretical model was tested on spectra of particle size separates of lunar soils [8,9]. An example of theoretical simulation of the effects of space weathering on spectra of a lunar soil is given in Fig.1. The spectrum of bulk soil, which is controlled by particles of sizes <20µm [10], is obtained from the spectrum of less mature 20-40 µm size fraction. As shown in Fig.1, decrease of particle size brightens the spectrum, but only addition of $npFe^0$ in the abundance observed in lunar samples increases the slope enough to reproduce spectrum of more mature soil.

In Fig. 2 simulation of Mercury spectrum starting from that of a lunar sample is presented. The example is chosen among those measured by Lunar Soil Characterization Consortium [10] to show that even high FeO content (15 wt.%) of mare soil can be consistent with Mercury spectrum provided that particle size is small and $\Delta c \sim 10^{-3}$ that corresponds to ~0.1 wt.% of FeO. This is about maximal concentration of $npFe^0$ consistent with rather high albedo of Mercury at reasonable comminution of the surface particles.

However, the spectrum of the lunar sample taken for this example (line 1 in Fig.2) is rather exceptional for lunar soil spectra because of the lack of typical camber at about 0.75 µm (compare Fig.1). Simulation have shown that this upward bent cannot be removed by addition of $npFe^0$. The bent practic-

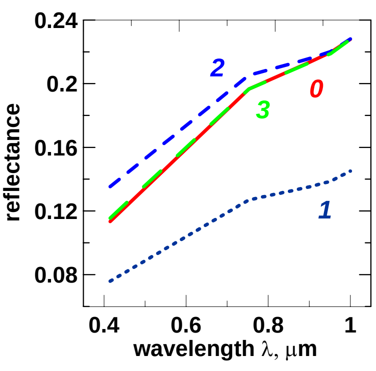

Fig.1. Simulation of bulk spectrum of lunar sample 61141 (*0*) starting from spectrum (*1*) of 20-40 µm size fraction: decrease of the average particle size *l* to 19 µm (*2*) and decrease of *l* to 16 µm combined with adding of $1.5\cdot10^{-2}$ vol.% of nano-Fe^0 grains (*3*).

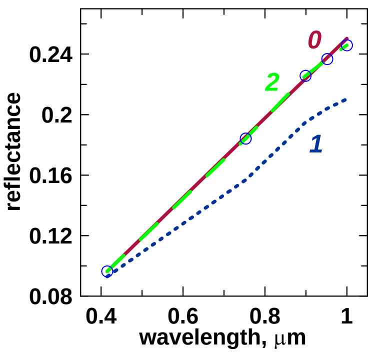

Fig.2. Mercury spectrum (*0*) and its simulation (*2*) starting from spectrum (*1*) of 5-10 µm size fraction of lunar sample 71061 with 15 wt.% of FeO; l is reduced by a factor of 1.7, and 0.1 vol.% of $npFe^0$ is added.

ally disappears at high Δ*c*, but appears again when particle size is decreased enough to compensate darkening yielded by $npFe^0$. The lack of the upward bent for the lunar soil spectrum shown in Fig.2 can be due to the fact that significant part (25%) of FeO is contained in opaque ilmenite mineral, most of the rest iron being in glass component, probably, in large (≥ λ) Fe^0-grains.

Large grains of absorbing materials are opaque; so grain growth changes light absorption to scattering and provides brightening of a soil. Consider such a possibility for Mercurian surface.

Maturation mechanisms for silicate soils and extralunar extrapolations: In 70-80th there were discussions as to which of the space weathering agents – protons of solar wind or impact vapor produced in meteoritic bombardment – cause reduction of Fe in maturation. Thermodynamic data for Fe-FeO equilibrium [11] indicate that high temperature and low oxygen pressure are sufficient for reduction of Fe (Fig. 3), i.e., no additional reducing agents are required. This was confirmed in laser heating experiments with silicates (e.g.,[12]). Thus, the lack of protons on Mercurian surface protected from solar wind by magnetic field cannot prevent soil maturation.

As shown in [13], even impact melting is sufficient for $npFe^0$ formation on airless bodies, since $npFe^0$ formation rate in melt is faster than cooling rate. Impacts of submicron scale may be responsible for $npFe^0$ in the rims of regolith particles, whereas micron-scale Fe^0 grains can grow in particle volumes due to submillimeter impactors.

Meteoritic flux and impact velocity on Mercury are greater than for the Moon [5], so much more impact melt and vapor can be expected [5] and hence, much more impact glass and reduced iron on Mercurian surface as compared to the Moon.

Overmaturation of regolith as a mechanism of removal iron spectral features. High temperature on Mercurian surface enhances diffusion of atoms, which favors so called ripening in space-weathered silicate particles. The largest grains among the numerous $npFe^0$ grow at the expense of the smaller ones, their number decreasing by many orders of magnitude, their size increasing up to a few microns. The kinetics of Fe^0 grain growth in typical Fe-bearing silicate was calculated in [7] on the base of ripening theory [14] and diffusion data [15] (Fig.4).

Conclusions: (1) Impact melting can provide the observed characteristics of mature soils without any additional maturation mechanisms. Consequently, the mechanism may cause regolith maturation on bodies shielded from solar wind irradiation, such as Mercury, and on asteroids where collision velocities are sufficient for impact melting only.

(2) Intensive meteoritic bombardment on Mercury can result in reduction of almost all Fe^{2+}, formation of impact glass and $npFe_0$; high surface temperature allows growth of most of the Fe^0 grains to sizes when light absorption changes to light scattering, which brightens the surface.

(3) Optical spectrum of Mercury does not enable us to put any constrains on the abundance of iron on the Mercurian surface. Reduction and growth can mask iron in surface material up to concentrations typical of lunar mare soils. Limits for iron content on the surface of Mercury can be established on the base of geological models.

References: [1] Vinogradov A. P. et al. (1972) *Proc. Lunar Sci. Conf. 3d*, 1421-1427. [2] Morris R. (1977) *Proc. LPSC 8th*, 3719-3747. [3] Housley R. et al. (1973) *Proc. Lunar Sci. Conf. 4th*, 2737-2749. [4] Pieters C. M. et al. *J. Geoph. Res. 98*,. 20817-20824. [5] Cintala M. J. (1992) *J. Geooph. Res. 97*, 947-973. [6] Warell J., and Blewett D. T. (2004) *Icarus 168*, 257-276. [7] Starukhina L. V. and Shkuratov Yu. G.. (2003) *LPSC XXXIV*, Abstr.#1224. [8] Starukhina L. V. et al. (1994) *LPSC XXV*, 1333-1334. [9] Starukhina L. V. and Shkuratov Yu. G.. (1996) *Solar System Res.*, *30*, 258-264. [10] Taylor L. A. et al. (2001) *J. Geoph. Res.*, **106**, 27985-28000. [11] O'Nell H. St. C. (1988) *Amer. Mineral. 73*, 470-486. [12] Moroz L. V. (1996) *Icarus 122*, 366-382. [13] Starukhina L. V. (2006) *LPSC XXXVII*, Abstr. # 1147. [14] Lifshitz I. M., Slyozov V. V. (1961). *J. Phys. Chem. Solids 19*, 35-50. [15] Buening D. K., and Buseck P. R. (1973) *J. Geoph. Res. 78*, 6852-6862.

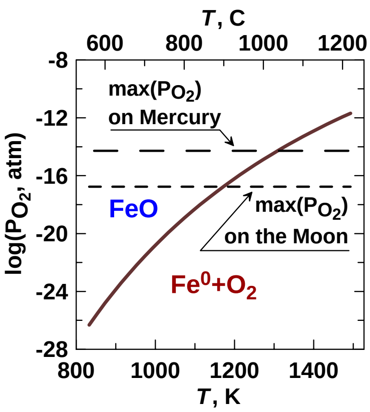

Fig.3. Thermodynamic conditions of iron reduction.

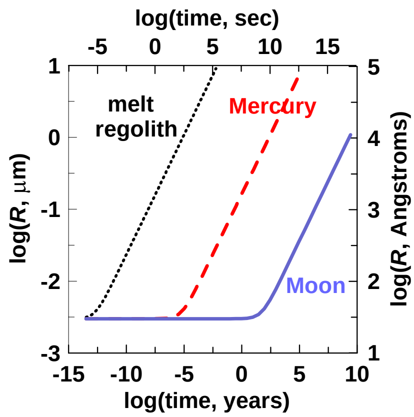

Fig.4. Rate of grain growth for reduced iron in olivine at day temperatures of the Moon and Mercury.