INTERACTION OF LARGE IMPACT CRATERS WITH THE MARTIAN AQUIFIER. E. Pierazzo[1] and B. A. Ivanov[2], [1]Planetary Science Institute (1700 E. Ft. Lowell Rd., Suite 106, Tucson, AZ 85719, U.S.A.; betty@psi.edu), [2]Institute for Dynamics of Geospheres, RAS (119334 Moscow Russia, Leninsky Prospect 38, baivanov@idg.chph.ras.ru).

Introduction: The presence of ancient footprints of water flows at the surface of Mars together with instrumental evidences of the presence of underground water/ice justify our interest on impact cratering as a cause for local (but widespread) hydrothermal activity in impact craters on Mars. Our previous work [1-3] focused on the starting conditions for impact-induced hydrothermal activity in the case of a model crater with diameter around 25 km. Here we discuss an extension of previously published results to the case of craters about 4 times larger.

Mars as a Target: We model the Clifford's standard model of porous space in the upper Martian crust [4] with an exponentially decreasing porosity, resulting in 10 to 12 km of the upper crust potentially saturated with water/brines. While this model may be oversimplified, it provides geologically reasonable estimates of the volume of the Martian underground aquifer. In our previous work [3] we use a target with decreasing initial ice fraction with depth (from 20 vol. % of ice near surface to zero at a depth of ~10 km). Assuming a Martian thermal gradient of 15 K/km we predict that water is present as ground ice above a depth of about 4 km, continuing as liquid ground water below. At a depth of about 10 km fracturing porosity becomes negligible due to the lithostatic pressure.

For the impact crater size modeled in [1-3] the maximum transient cavity depth is of the order of 6 km. Here the entire crater formation zone is imbedded into the potential aquifer layer and special attention must be devoted to a detailed description of the behavior of a H_2O/rock mixture [3]. However, even for such a small crater a comparison with our earlier approach [1], that used a dry rock target, shows that the temperature field under the crater is mostly controlled by the rock EOS. Therefore, the temperature distribution in a dry target is a good proxy for the more complicated case of a H_2O/rock mixture. Here we use this result to make a simplified analysis of the thermal field under craters with much larger diameters, using as additional support data from terrestrial impact craters.

Central Uplift Amplitude: In the terrestrial environment the initial depth of rocks present in the central uplift (*SU*="structural uplift") of complex impact craters is known from geological data [5]. Grieve et al. [5] propose the formula $SU=0.06\ D_a^{1.1}$ where D_a is the apparent (visible at the surface) diameter of the crater. This expression provides (in the limits of the accuracy of the available data) the same estimates of SU as the earlier, simpler relation: $SU \approx 0.1\ D_a$ [6]. Recent numerical modeling [7] suggest even larger values, using $SU \approx 0.15\ D_a$ for Vredefort (where the modeled SU is ~22 km for a model diameter D_a ~140 km). A direct transfer of these estimates to Mars suggests that craters with diameters larger than 100 km should uplift rocks from a depth beyond 10 km. Consequently, just after crater formation the central uplift in these large craters consists of initially non-porous dry hot rocks from below the assumed Martian aquifer. To refine this conclusion we need to model the complex crater formation in the Martian gravity field. This was previously done for Gusev crater, that has a diameter D_a ~150 km [8].

Numerical Modeling: The numerical modeling in question uses the 2D hydrococe SALEB with tabular versions of the basalt, granite, and dunite ANEOS equations of state to describe properties of the Martian crust and mantle. Mechanical description of materials include strength and dry friction for damaged rocks and the Acoustic Fluidization (AF) model to simulate the temporary dry friction reduction around the growing crater [9]. Several model runs were done with various choices of the AF parameters. The most sensible parameter – the oscillation decay time - varies from 60 to 120 seconds. The impact is modeled by a spherical projectile 14 to 16 km in diameter with impact velocity of 10 km/s. The target is initially balanced in the Martian gravity field. The crust thickness is assumed to vary from 50 to 60 km with a near surface thermal gradient of about 13 K/km. While the hypothetical porous water/ice bearing layer is not explicitly included into this model (all runs use a dry target), a separate material (with a granite EOS) is used to outline the deformations of the upper crust layer.

All model runs resulted in the formation of a complex crater with rim-to rim diameters of 160 to 170 km and a central uplift. Depending on the AF model parameters used, the summit of the central uplift may or may not reach the pre-impact surface. The maximum model crater depth is in the annular trough surrounding the central uplift, and reaches 4 to 7 km below the pre-impact level.

The most appropriate modeled crater is shown in Fig. 1. The crater has a rim diameter of 170 km and an apparent diameter of 140 km – very close to the modeled [7] and estimated [5] apparent diameter for Vredefort. In the modeled Martian "Vredefort" the SU is about 15 km – approximately 70% of the terrestrial SU for a crater of the same diameter. This is obviously due to the lower Martian gravity. This result is very close to the simple estimate: $SU \approx 0.1D_a$ discussed above. Consequently we can state that according to the best available numerical models to date, the relation $SU \approx 0.1D_a$ while underestimating the structural uplift for terrestrial craters in the diameter range of 150 to 200 km, it best describes the first model results for Martian craters of a similar size.

Discussion: Predictions for the presence and evolution of hydrothermal systems in large Martian craters depend on estimates of amplitudees of the structural uplift in the crater center. In the case of a large impact, the transient crater collapse uplifts the deepest crustal rocks. If the current assumptions about the depth of the Martian aquifer are realistic, we can distinguish two situations in the interactions of impact craters with the water-saturated rocks: (1) in small craters (D<100 km) the structural uplift is less than 10 km and the crater-forming flow occurs completely in the water-bearing rocks and, (2) in larger craters (D>100 km) the central uplift consists mostly of deep seated rocks, located below the aquifer. Even in the first case (e.g., a crater with D~25 km, modeled in [1-3]), the uplift is made of rocks with low (~5%) initial porosity, and the amount of water present, before the penetration of ground water from the external terrain, is relatively small. In the second case, deep dry rocks make up the entirety of the central uplift. This is a possible explanation for the visible lack of impact-related hydrologic activity in the case of the crater Lyot [9].

We conclude that the starting conditions for possible hydrothermal activity in large craters are mostly depend on the thermal field created by the impact event. Water will have to be delivered to the central uplift later from outside – a process controlled by the permeability of rocks and the amount of liquid water available in the outer parts of the crater.

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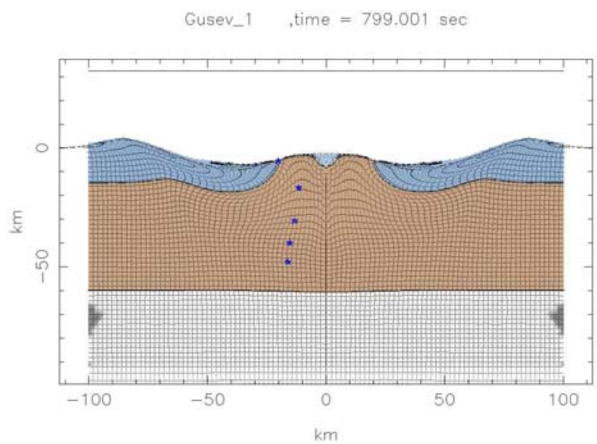

Fig. 1. Cross-section of a model impact crater about 170 km in diameter in a three-layered target. Blue represents the upper (H_2O bearing) crust layer; brown represents the lower crust over the mantle below 60 km, in gray.

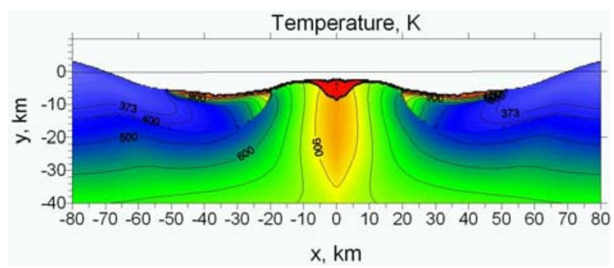

Fig. 2. Final temperature field under a model Martian crater ~170 km in diameter. The temperature within the central uplift (~±20 km from the crater center) is above ~600 K, resulting from the combination of lower (and initially hotter) crustal rocks uplift and shock and deformational heating. We can assume high fracturing and high permeability of these (initially dry) rocks. Beyond a radial distance of ~50 km (inner crater slope) the ground ice should not be melted (approximately bounded by the 373K isotherm) by the impact event. The "wettest" region appears to be a surficial layer in the annular trough (~20 km to ~50 km from the crater center). After crater collapse, this annular zone is covered with impact melt and breccia layer, providing extra heat for melting of ground ice.

[1] http://sci2.esa.int/Mars/MarsExpressConference2005.pdf