WHAT CAUSES THE RELATIONSHIP BETWEEN LARGE IMPACT BASIN RIMS AND VOLCANISM ON MARS?

Walter S. Kiefer[1] and Matthew B. Weller[2], [1]Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston TX 77058, kiefer@lpi.usra.edu, [2]Institute for Geophysics, Jackson School of Geosciences, Univ. of Texas, Austin TX 78758, mbweller@ig.utexas.edu.

Observations: Several large impact basins on Mars, including Hellas, Isidis, and Utopia, have concentrations of large volcanos on their rims. Hellas (2300 km diameter, [1]) is largely surrounded by slightly younger volcanism in its rim zone, including Hadriacus Mons, Tyrrhenus Mons, Amphitrites Patera, Peneus Patera, and Noachis Terra [2-5]. Isidis (1900 km diameter, [1]) has Syrtis Major on its western rim zone [6, 7]. Utopia (2300 – 3300 km diameter, [1, 8]) has Elysium Mons and Hecates Tholus on its southeast rim zone. Moreover, gravity models of Utopia indicate that it is filled with $5{\cdot}10^7$ km^3 of post-impact fill, whose likely density of > 2850 kg m^{-3} favors a dominant role for volcanic filling [8]. The spatial relationship between these three large impact basins and their basin rim (and in the case of Utopia, possible basin floor) volcanism suggest the possibility of a causal relationship between basin formation and volcanism. We explore possible mechanisms for this causal relationship in this study. It is worth noting, however, that the Argyre basin (1850 km diameter, [1]) does not have volcanism in its rim zone [9].

Table 1 summarizes key constraints that might be useful in testing possible mechanisms for producing impact basin rim volcanism. Volcanic volumes are based solely on the surface edifice for each structure. Subsurface volcanic roots can be an important, even dominant part of the total volcanic volume and have been estimated in some cases by gravity modeling [7, 8, 10]. For Amphitrites and Peneus, the structures have such low relief that is difficult to reliably estimate the volume; in those cases, Table 1 indicates the surface area of the volcanic edifice.

Hellas, Isidis, Utopia, and Argyre are all thought to have formed between 4.0 and 3.8 Ga [11]. The times of peak volcanic activity, as estimated from superposed crater densities [12, 13], are typically only slightly younger than the impact basin ages. This temporal relationship between impact and volcanism is consistent with the possibility of a causal relationship between the two processes. In a few cases, the volcanism apparently ended shortly after it began (Amphitrite, Peneus, and Noachis). On the other hand, for some volcanos, at least modest levels of volcanic activity continued for several billion years after the basin impacts (Syrtis, Hadriacus, Tyrrhenus, and Hecates).

Feature	Volume (km^3)	Peak Activity (Ga)	Last Activity (Ga)
Syrtis Major	200,000	3.7	2.0-0.2
Hadriacus Mons	16,000	3.8	1.6-1.1
Tyrrhenus Mons	21,000	3.8	3.3-1.1
Amphitrites Patera	400,000 km^2	3.8	3.6
Peneus Patera	400,000 km^2	3.8	3.8
Noachis Terra	56,000	3.9	3.8
Elysium Mons	200,000	3.8	3.1
Hecates Tholus	67,000	> 3.5	0.3

Table 1: Summary of key observational constraints for impact basin rim volcanism on Mars. Volumes are from [6, 14]. Duration is from [12, 13]. Noachis Terra is from [5].

Effects of Impact Heating: We explore several possible mechanisms for creating a causal relationship between the formation of large impact basins and subsequent basin rim volcanism. One possible mechanism involves impact heating. For near-vertical impacts, the impact heats a region ~500 km in radius and ~400 km deep for an Isidis-sized impact [15]. The heated region is both broader and deeper for larger impacts; if the Utopia impact is at the large end of the range of basin diameters listed above, this could affect both the volume and duration of the induced volcanism. Portions of the mantle are heated above the solidus, and this impact melt escapes from the mantle to the surface by Darcy flow in a few thousand years. We emphasize that we do *not* consider this impact melt to be volcanism.

However, once the impact melt escapes, the residual mantle will be at its solidus and can rise viscously due to its thermal buoyancy. Because this material is already at it solidus, any upwelling will lead to adiabatic decompression melting. In our view, the magmatism that results from this adiabatic decompression melting can appropriately be considered volcanism.

We have modeled this using an impact heating model [15] as the initial condition for a finite element mantle flow calculation. Because of the approximate cylindrical symmetry of the impact basins, we model the results in spherical axisymmetric geometry. Magma production is calculated as in recent mantle plume volcanism modeling [16].

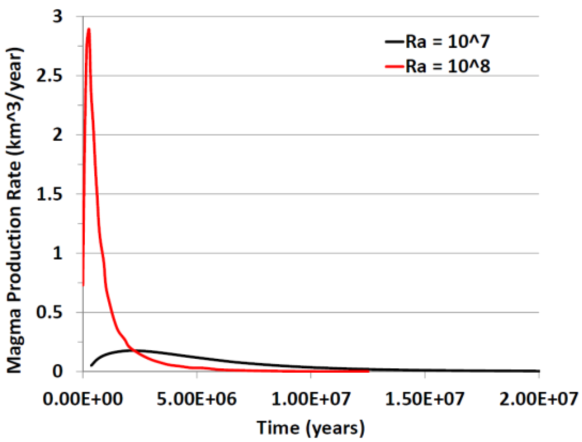

Figure 1: The time history of adiabatic decompression magma production for an Isidis-sized impact.

Figure 1 shows results for an Isidis-sized impact. The vigor of the convective flow is constrained by the Rayleigh number (Ra), with larger Ra indicating more vigorous flow. A thermal of Ra=10^8 is likely appropriate for Mars at 4 Ga, and there is a strong but brief pulse of magma production lasting just 5 million years (red line). Less vigorous convection (black line) extends the magma production only slightly. The induced volcanism in these models is too brief to explain the range of volcanism ages in Table 1. Moreover, the resulting magma production is concentrated in the basin center; this does not explain the basin rim volcanism although it might contribute to the dense basin fill inferred at Utopia [8]. Oblique impacts can push the impact heating away from the basin center. However, hydrocode modeling shows that oblique impacts have lower peak shock pressures (hence less impact heating) as well as a shallower distribution of impact heating [17]. Both factors act to reduce the magnitude and duration of post-impact adiabatic decompression melting.

Effects of Basin Structure: Large impact basins strongly modify the structure of the crust near the impact site. In an annulus outside the basin rim, the crust is thickened and high porosity ejecta is deposited. Because the radioactive elements U, Th, and K are concentrated in the crust, the thickened crust is a region of enhanced heating relative to normal thickness crust. High porosity ejecta will have lower thermal conductivity than intact rock of the same composition. Thus, both of these effects may lead to local heating of the sub-crustal mantle at or just outside the basin rim [18, 19]. GRAIL gravity observations of the lunar Orientale basin reveals thickening of the crust by up to 10 km in the basin rim zone [20]. The ejecta deposit around Orientale is up to 2.9 km thick [21], and the porosity inferred from GRAIL gravity exceeds 18% at both Orientale and Moscoviense [22]. Similar magnitudes of crustal thickening and ejecta deposition are presumably present at large martian impact basins.

These crustal modifications may help to explain both the spatial relationship between impact basins and basin rim volcanism as well as the temporal relationship. The enhanced radioactive heating in the thickened crust will heat the uppermost mantle with time and thus can explain a phase lag between impact basin formation and the onset of basin rim volcanism. In addition, the decay of radioactive heating with time may provide a natural explanation for the termination of volcanism at these structures. We are currently developing finite element simulations to quantify these effects. Basin loading stresses may also play a role in controlling the location of these volcanos [23].

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