FERROVOLCANISM: IRON VOLCANISM ON METALLIC ASTEROIDS. J. N. H. Abrahams[1] (abrahams@ucsc.edu), F. Nimmo[1], [1]Department of Earth & Planetary Science, University of California Santa Cruz, Santa Cruz, CA 95064, USA

Summary: We predict iron volcanism to have occurred on metallic asteroids as they cooled. We discuss the implications of this process for both the evolution and the modern appearance of these bodies.

Introduction: Metallic asteroids, the exposed cores of disrupted planetesimals, are expected to have been exposed while still molten. Many would have cooled from the outside in, crystallizing a surface crust which would then grow inward [1]. Because the growing crust is expected to be more dense than the underlying melt, this melt will tend to migrate toward the surface whenever it is able, resulting in iron volcanism. Compressional stresses produced in the crust while it cools will be relieved by thrust faulting, which will also provide potential conduits for melt to reach the surface. Below, we provide a quantitative, order-of-magnitude description of these processes.

Thermal Evolution: Prior to mantle removal, planetesimal cores are insulated and cool very slowly. A mantle thermal diffusivity of 10^{-6} m^2/s predicts that cores as small as 60 km, below an equally thick mantle, are expected to take of order 100 Myr to solidify. As a result, mantle stripping in the early solar system will typically produce fully liquid metallic bodies.

Following mantle removal, the bodies will cool to space and solidify. From the meteorite record we know that some of these bodies solidify outward (like Earth's core) [2], and some cool concentrically inward [1]; here we are concerned with this latter group. Metallic bodies cooling inward will form 20 km crusts in roughly 1 Myr, and solidify completely in several Myr, depending on the size of the body.

Volcanic Cycle: We describe a volcanic cycle which has three main stages: compressional stress accumulation as the body solidifies, faulting and diking when this stress causes the crust to fail locally, and relaxation back to compression after eruptions.

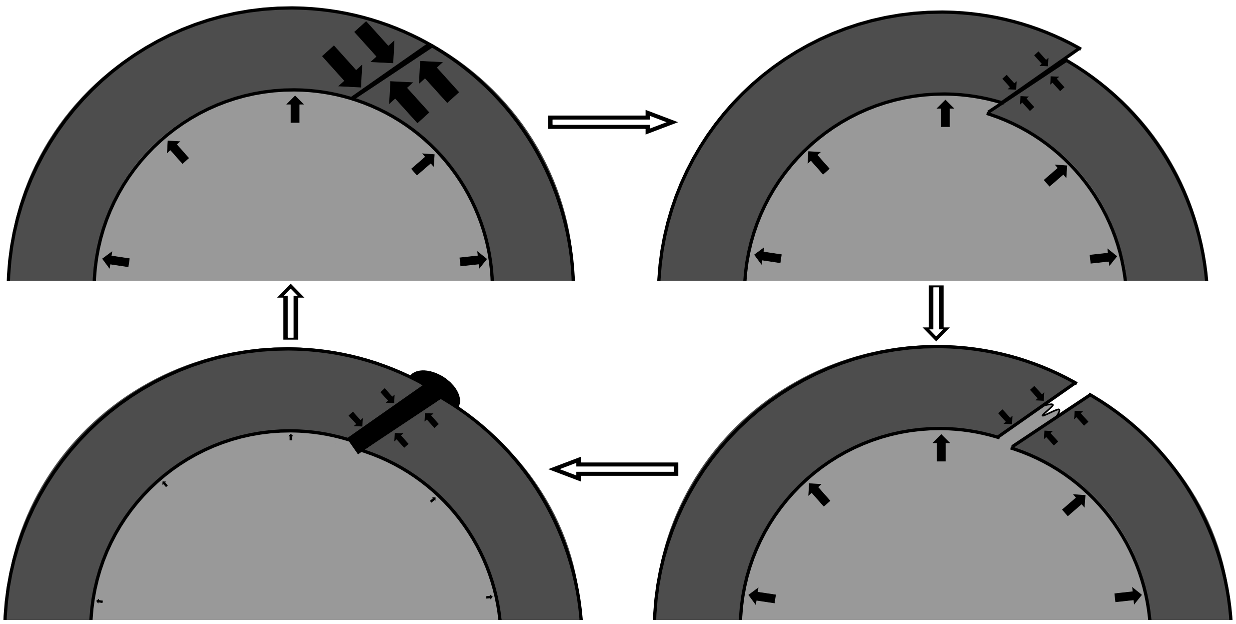
Stage One, Stress Accumulation: New solid will occupy a smaller volume than the melt from which it formed, and the resulting radial contraction generates compression. For a nominal 10 km crust with a linear thermal gradient, that gradient is $\frac{dT}{dz} \approx 10^{-1}$ K/m. This corresponds to a radial contraction (due to crystallization) of $\dot{h}\frac{\Delta\rho}{\rho} = \frac{k\frac{dT}{dz}}{L\rho}\frac{\Delta\rho}{\rho} \approx 10^{-10}$ m/s, where L is latent heat and k is thermal conductivity. This volume loss corresponds to a nominal strain rate of $\dot{\epsilon} = \dot{h}\frac{\Delta\rho}{\rho}/R \approx 10^{-15}$ s^{-1}. The strain rate will evolve substantially with crustal thickness, but is generally within roughly an order of magnitude of this value. For an eruption to occur, melt needs to be able to force open cracks, and, as on Mercury, this compressive environment must be overcome.

Stage Two, Dike Formation: In the shallow, brittle crust, dikes will propagate as long as the metallic magma ocean pressure is enough to overcome compression and force open existing cracks. The excess pressure in the ocean is given by $P_{ex} = gh\Delta\rho \approx \frac{1}{40}\rho gh$, where $\Delta\rho$ is the density contrast between solid and liquid and ρ is the solid density. The maximum stress which faults can support is given by $\sigma_{max} = \frac{2f_s\rho gh}{(1+f_s^2)^{1/2}-f_s} \approx \frac{1}{4}\rho gh = 10P_{ex}$. The strain rate of 10^{-15} s^{-1} will cause faulting roughly every kyr, so if faults can relieve 90% of their accumulated stress, the ocean can regularly force apart the brittle crust in the local area where stress was relieved.

In the deeper, ductile crust, slow stress accumulation will not be able to produce fractures. With a maxwell timescale of order one second near its melting point, the deep crust needs to be rapidly strained in order to form dikes. The most likely source of this stress is cratering, and we find that impacts can generate sufficient peak pressures at depths of 10 to 100 projectile radii, enough for the largest impactors to fracture through the entire crust. Once a fracture has formed, it will propagate if the forces trying to extend it exceed the forces resisting that extension. In particular, there is a critical length past which dikes will extend themselves [3], given by $K_c = (\pi l_{crit})^{1/2}\left[T + 2g\Delta\rho l_{crit}/\pi\right]$, where K_c is fracture toughness and T is the background stress (positive for tensile stress). With 100 kPa of compressive stress (5% of what faults can support), this critical length is 11 km, essentially the whole crust in a 100 km body. However, delamination, solid dripping off from the base of the crust as diapirs, can produce substantial local extensional stresses. Delamination can decrease this critical length to $\sim$1 km, while impacts will cause even smaller critical lengths; delamination events are expected to be more frequent than sufficiently large impacts.

Melt forced upward through a crack will lose heat to crack walls and eventually refreeze. For eruptions to occur, the timescale of freezing needs to be longer than the timescale of melt ascent. We find that dikes need to be at least 1 m wide to avoid refreezing, a similar value to Earth, indicating that dikes should be able to propagate to the surface.

Stage Three: Reestablishing Compression: Eruptions are self-limiting processes, because compressional stress reaccumulates as the liquid volume reduces and the crust subsides. An eruption will stop when the stress in the crust is equal to the excess pressure in the ocean. If faulting were able to relieve all the stress in the crust, the volume of erupted material would form a global layer of



Cartoon depicting the volcanic cycle we describe. Beginning in the top-left is the typical state of the crust, where compressional stress prevents the melt at depth from reaching the surface. Volcanism then begins with a faulting event and the resulting decreased local compressional stress in the crust is no longer larger than hydrostatic pressure. Next, the liquid interior forces open the fault and melt migrates through the newly formed dike. Finally, melt reaches the surface and the now-depressurized ocean is no longer able to support the crust, causing the crack to close again. Subsequent crustal growth increases contraction in the crust slowly until faulting can occur again.

thickness $\delta = \frac{gh\Delta\rho R}{E} \approx 0.3$ m for body radius R and elastic modulus E. The amount of solid which needs to plate onto the base of the crust to cause a faulting event is given by $\Delta h = \frac{\frac{\pi}{3}\rho^3 R^2 hG}{\Delta\rho E} \approx 150$ m. The fact that much more material, 150 m, needs to freeze at depth than the amount erupted, 0.3 m, indicates that eruptions do not play a major role in the body's thermal evolution, and erupted material is a small fraction of its total volume. Note that this assumed stress accumulation and relief occurs on a single fault, which is certainly not the case, but is a useful way to evaluate the importance of volcanism.

Potential Observations: A key prediction that follows from metallic volcanism is that bodies hosting it will have two end-member types of solids which experienced very different histories. Material crystallizing onto the bottom of the crust will cool slowly and will exhibit an inverse correlation between cooling rate and light element content. Erupted material will crystallize very quickly, and unless it is reheated will show $\sim$instantaneous cooling. This quenching will prevent elemental fractionation, so erupted material will record the (non-volatile) element abundances of the liquid interior at the time it was erupted. Incompatible element concentrations will be much larger than deeply formed solids. If volatile exsolution is involved, the quenched solids would likely contain vesicles. However, because erupted material is a small volume fraction of the body, quenched meteorites should be rare.

In addition, bodies which hosted metal volcanoes may preserve geomorphic evidence of the process. The Psyche mission will have the unique opportunity to look for in situ evidence of volcanoes. Predicting what iron volcanoes would look like, both when they are fresh and after Gyrs of modification, will be important going forward. Iron volcanism, a process fundamentally coupled to compressional tectonics, is likely to bear similarities to volcanism on Mercury, where volcanic features are associated with craters [4], and a lot can be learned from testing this expectation.

It is likely that metallic asteroids will posses dynamos while they crystallize, and surface volcanic flows will capture instantaneous snapshots of this field. Magnetometers may be able to confirm the presence of ferrovolcanism and study both volcanic processes and what they recorded about dynamo processes.

Conclusions: We describe the existence of a novel geophysical process, ferrovolcanism. Metallic asteroids which cool from the outside-in have the potential to erupt some of their interior melt onto their surfaces. Their compressional stress environments will substantially limit this volcanism, likely leading to erupted material being a small fraction of their total volume.

References: [1] Yang J., et. al. (2008) *GCA*, 72. [2] Yang, J., et. al. (2010) *GCA*, 74. [3] Crawford, G. D., & Stevenson, D. J. (1988). *Icarus*, 73(1). [4] Klimczak, C., et. al. (2018). *Icarus*, 315.