THERMAL CHALLENGES FOR AN ANCIENT DYNAMO ON VESTA.

J. H. Roberts, A. S. Rivkin, and N. L. Chabot, Johns Hopkins University Applied Physics Laboratory, 11100 Johns Hopkins Rd., Laurel, MD 20723, (James.Roberts@jhuapl.edu)

Introduction: Vesta is the second-largest asteroid in the solar system, and the smallest known silicate body to be fully differentiated [1-2]. It is thought that the Howardite-Eucrite-Diogenite (HED) meteorites [3,4] represent pieces of Vesta's crust [5], ejected during an impact (possibly the collision that formed the Rheasilvia Basin at the south pole [6]). Remanent magnetism observed in the eucrite Allan Hills A81001 is consistent with the parent body having cooled in a 2 μT magnetic field [7]. This magnetization has been interpreted as evidence for a magnetic field at the surface of Vesta at the time the rock cooled [7]. Models of thermal evolution in protoplanets predict that the maximum lifetime of a dynamo an object the size of Vesta is ~100 My [8], far earlier than the 3.69 Gy age of the meteorite. This age difference suggests that the origin of the magnetization of the meteorite is remanent magnetism in the surrounding crust, that was itself magnetized by a global dynamo-driven magnetic field shortly after accretion.

A self-consistent model of the transfer of heat between the core and mantle is desired in order to better understand core cooling, which is a key control on dynamo activity. The vastly different material properties and timescales have made modeling of the full interior intractable. Here, we present preliminary results of coupled models of mantle dynamics and parameterized thermal evolution of the core on Vesta. This allows us to more accurately characterize the conditions under which dynamo activity may be more favorable. Models of cooling magma oceans [8] on planetesimals suggest that the core heat flux is sufficient to sustain dynamo activity during this early period, but it is not clear whether the crystallized rocks will be cool enough to record a remanent magnetic signature at this stage. Here we focus on thermal evolution of Vesta at later times

Thermal Modeling: We begin our models after accretion and differentiation of Vesta are complete, and consider a 152.5 km thick silicate mantle over a 110-km radius iron core. Thermophysical parameters for both layers are taken from [8], assuming the bulk composition to be that of CV chondrites. We model thermal evolution in the mantle using the finite element code Citcom in 2D axisymmetric geometry [9]. Because lateral mixing in the core occurs rapidly with respect to mantle dynamics, we only consider radial temperature variations in the core and model the thermal evolution using a 1D radial parameterization [10-12]. At a given timestep, we fix the mantle temperature and solve the 1D enthalpy equation in the core and lower thermal boundary layer (TBL) of the mantle over a time corresponding to a mantle timestep. We then update the temperature at the core-mantle boundary (CMB) and TBL, and let the mantle convection progress for another timestep. We continue this iteration until the core temperature becomes almost adiabatic and the entire core is convecting. Mantle convection then proceeds while the cooling core retains an adiabatic temperature.

In the absence of significant internal heating, the mantle of Vesta is unable to convect. Assuming an instantaneous accretion of chondritic material such that no melting has occurred, the Rayleigh number does not exceed the critical value for any realistic rheology. However, the high degree of differentiation on Vesta suggests that it accreted less than 1.5 My after CAI formation [13], and therefore was substantially heated by ^{26}Al [7,8]. We incorporate mantle heating by ^{26}Al (as well as long-lived radioactive isotopes) that decays with time. We assume no radiogenic heating in the core. The initial CMB temperature is 1200 °C (the CV chondrite solidus [8]). We start with an adiabatic temperature distribution in the core, and a conductive profile in the mantle. We note that this setup may not be consistent with models of differentiation that result in magma oceans [8], but these initial conditions do not greatly affect the outcome.

Thermal Evolution: For a model starting only 1 My after CAI formation, the ^{26}Al heating is enormous. The temperature rapidly exceeds the solidus at all depths below about 5 km. Thus the lack of a starting magma ocean is not particularly significant; it develops immediately. The core is heated somewhat due to the extremely hot mantle above, resulting in a temperature inversion as in [8]. The core is hottest at the top, and becomes stably stratified. Thermal convection in the core is halted along with any dynamo activity. The first rocks to cool will be completely unmagnetized. Figure 1 shows the evolution of the radial temperature structure in the core and mantle. The mantle cools into space and into the core. As this temperature drops, the core temperature rises, although it is still hottest at the CMB and stable to convection. After ~90 My the mantle has finally cooled below the core temperature. At this point heat flux out of the core becomes positive again (see figure 2). The core temperature just below the CMB becomes adiabatic shortly thereafter and the outer core begins to cool by convection. This convecting region expands downwards as the inner core cools, and the entire core is convecting again less than 10 My later. The core heat flux continues to increase positively, peaking at about 200 My, after which it slowly decreases towards zero. The core continues to cool and eventually freezes by 400 – 600 My depending on the sulfur content of the core.

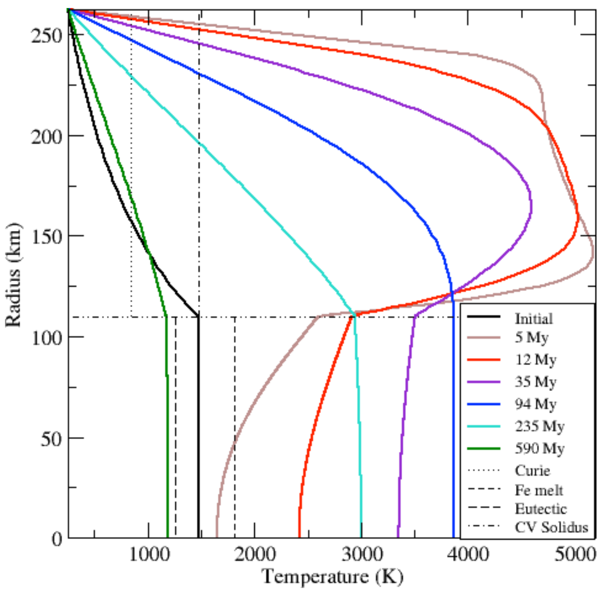

Figure 1: Temperature in the core and mantle of a differentiated Vesta beginning 1 My after CAI formation. Vertical lines marking the Fe melting point and Fe-FeS eutectic temperature, CV Chondrite solidus, and Curie temperature of magnetite are shown for reference. The horizontal line marks the CMB.

Discussion and Future Work: It is currently not clear from our preliminary results that the peak CMB heat flux is fast enough to sustain vigorous core convection, particularly at later times once the silicate portion cools below the Curie temperature. This situation is unfavorable to thermally-driven core dynamo. Once the core begins to freeze, compositional convection may promote core circulation. However, the gravity on a Vesta-sized body is so low that the core may have crystallized inwards, inhibiting compositional convection, as suggested for some iron meteorite parent bodies [14]. Basin-forming impacts (e.g., Rheasilvia) have the potential to deliver substantial amounts of additional heat to the interior [15], but a projectile large enough to re-melt the core would be more likely to disrupt Vesta altogether [16]. Furthermore, since the Rheasilvia impact is a probable cause of the ejection of the HED meteorites, fragments that have already left the parent body could not be affected by any impact-induced dynamo.

The preliminary results presented here do not account for the rheological change in the mantle upon melting. Thus the times given for the mantle to cool to the solidus are significantly overestimated. Moreover, a substantial amount of heat is consumed by the melting process and will moderate the maximum temperatures obtained. Future work will focus more strongly on the time-evolving thickness of crustal material that has cooled below the Curie temperature and could potentially record an ancient dynamo signature, as well as further evaluation of the heat flux conditions required to sustain dynamo activity.

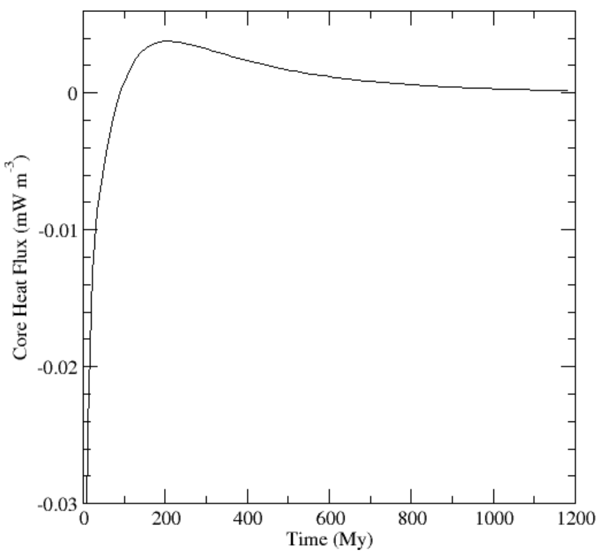

Figure 2: Heat flux across the core-mantle boundary.

References: [1] Weiss B. P. et al. (2008) *Science, 322*, 713-716. [2] Russell C. T. et al. (2012) *Science, 336*, 684-686. [3] Righter K. and Drake M. D. (1997), *MPS, 32*, 929. [4] Ruzicka, A. et al. (1997) *MPS, 32*, 825. [5] de Sanctis M. C. et al. (2012) *Science, 336*, 697-700. [6] Schenk P. et al. (2012) *Science, 336*, 694-697. [7] Fu R. R. et al. (2012) *Science, 338*, 238-241. [8] Elkins-Tanton L. T. et al. (2011), *EPSL, 305*, 1-10. [9] Roberts J. H. and Zhong S. (2004), *JGR 109*, E06013. [10] Arkani-Hamed, J. (2012), *PEPI, 196-197*, 83-96. [11] Roberts J. H. and Arkani-Hamed (2012), *AGU Fall Meet. 2012*, DI31B-2198. [12] Arkani-Hamed J. and Roberts J. H. (2013), *LPS XLIV*. [13] Ghosh A. and McSween H. Y. (1998), *Icarus, 134*, 187-206. [14] Goldstein J. J. et al. (2009) *Chemie der Erde, 69*, 293-325. [15] Reese C. C. (2002) *JGR, 107*, 5082. [16] Roberts J. H. et al. (2011) *LPS XXXII*, 2242.