SUPPRESSING THE MARTIAN DYNAMO WITH ONGOING HYDROGENATION OF THE CORE BY HYDRATED MANTLE MINERALS. J. G. O'Rourke and S.-H. Shim, School of Earth and Space Exploration, Arizona State University, Tempe, AZ (jgorourk@asu.edu).

Introduction: Mars lacks an internally generated magnetic field today. However, crustal remanent magnetism discovered by Mars Global Surveyor evidences dynamo activity prior to ~4.1 billion years ago [e.g., 1]. Vigorous fluid motions must constantly occur in the core to drive a magnetic dynamo. Any field would dissipate within ~10^4 years after the fluid stagnates [e.g., 2]. Thermal and/or compositional convection, in particular, requires a minimum amount of heat flow across the core/mantle boundary. Popular explanations for the cessation of internal magnetism on Mars rely on a decrease in the rate of core cooling as mantle convection becomes sluggish [3], possibly hastened by radiogenic heating in cumulates above the core/mantle boundary [4]. Alternate hypotheses include thermal stratification of the core caused by shock heating after a giant impact [e.g., 5]. Identifying the mechanism(s) of dynamo activity is key to the internal evolution of Mars in general.

Here we present a new scenario for the cessation of dynamo activity in Mars. Hydrogen diffusing out of hydrated silicates into the core creates a sink of gravitational energy that overwhelms other power sources for the dynamo. The mantle becomes dehydrated over time as wadsleyite and ringwoodite equilibrate with the core—losing almost all their hydrogen—but the core/mantle heat flux never rises above the critical value to drive a dynamo. Geophysical missions like InSight could detect several signatures of a hydrogen-rich core.

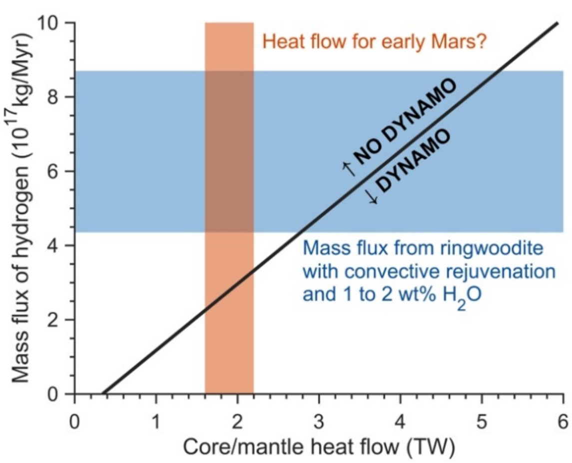

Figure 1 | Preliminary calculations. Rates of hydrogenation that will kill any dynamo (black line) compared to the mass flux derived for ringwoodite (blue region) and estimates for the actual heat flux around the age of magnetized crust (orange region) [3]. This predicts that equilibrium hydrogen partitioning from hydrated silicates would quickly suppress convection in the core.

Delivering hydrogen to the core: We focus on the mass of hydrogen that is available for transport into the core over geologic time. Hydrogen that is already mixed in the core immediately after accretion does not affect the dynamo except by reducing the liquidus temperatures—thus delaying the formation of an inner core.

Total mass budget: Minerals like wadsleyite and ringwoodite—found in the transition zone between depths of 410 and 660 km on Earth—can host up to ~2.5 wt% of water in their crystal structures. Studies of ringwoodite inclusions in ultra-deep diamonds imply that the water content could be ~1 wt% in the region [6]. Earth's core is blanketed with relatively dry materials like bridgmanite. However, in Mars, which is smaller, the equivalent of Earth's transition zone starts at a depth of ~1100 km and extends to the core/mantle boundary. We estimate a total budget of hydrogen as ~10^{20} kg based on the water storage capacity of wadsleyite, ringwoodite, and majorite—assuming that water is predominantly stored in the olivine polymorphs.

Rate of mass transfer: The partition coefficient for hydrogen between ringwoodite and iron—at pressures appropriate to Mars—was experimentally determined as ~26, which means that ~97 mol% of hydrogen in ringwoodite should partition into metallic iron at equilibrium [7]. However, chemical diffusion limits the actual rate of mass transfer into the core.

We can derive a characteristic scaling for hydrogenation by equating the timescales for chemical diffusion and convective overturn. The diffusivity of hydrogen in ringwoodite at 1800 K is D_m ~ 6 x 10^{-11} m^2/s [8], and the convective velocity is plausibly v_m ~ 10^{-10} m/s. A natural timescale is thus $t_m = D_m/(v_m)^2$ ~ 6 x 10^9 s. We assume that all hydrogen within l_m ~ $(D_m t_m)^{1/2}$ ~ 60 cm above the core/mantle boundary is transported into the core in t_m, after which this boundary layer is rejuvenated by solid-state convection. The analogous time scale in the core is only t_c ~ 0.1 s since v_c ~ 10^{-3} m/s. Therefore, the initial rate of mass transfer is ~4.3 to 8.6 x 10^{17} kg/Myr for ~1 to 2 wt% water in ringwoodite. The transition zone will gradually dehydrate over time.

Delayed delivery? Depending on the actual size of the core and the initial temperature profile, the basal mantle may start within the stability field of bridgmanite [9-11]. Ringwoodite might form at the core/mantle boundary only after a period of cooling and hydration through diffusion or convection because of the negative

Clapeyron slope of the post-spinel transition. If hydrogenation destroys the dynamo, then the youngest magnetized crust may mark the disappearance of bridgmanite—connecting the interior and surface evolution.

Several factors could affect the eventual rate of hydrogenation. First, the partition coefficient could change for metal that already contains some light elements like sulfur. Second, the abundance of hydrated silicates is possibly heterogeneous across the core/mantle boundary. Because the density anomalies that drive flow in the (low viscosity) core are quite small, baroclinic instabilities will swiftly erase lateral density variations [12].

Energetics of the dynamo: We calculate the dissipation available to generate a global magnetic field using a one-dimensional parameterization of the radial structure of the core [13, 14]. Neglecting radiogenic heating, the global energy balance is simply $Q_C = Q_S - Q_H$. Here Q_C is the total heat flow across the core/mantle boundary, Q_S represents secular cooling, and Q_H is the (sink of) gravitational energy associated with hydrogenation. We may write the minimum core/mantle heat flow required for a dynamo as $Q_C > (T_C Q_H + T_S T_C E_K)/(T_S - T_C)$, where T_C is the temperature at the core/mantle boundary and T_S is close to the average temperature in the core. The entropy sink associated with thermal conduction along the adiabatic temperature gradient is E_K. Even without hydrogenation, $Q_C > 0.34$ TW is required to drive a dynamo if the thermal conductivity of the core is 40 W/m/K. We can also equivalently write an upper bound for $Q_H < [(T_S - T_C)/T_C]Q_C - T_S E_K$. Because of the Carnot-like efficiency term that penalizes Q_S (and thus Q_C), $Q_H = 0.2$ TW mandates $Q_C > 5$ TW.

We calculate Q_H as the change in gravitational energy caused by redistributing an infinitesimally thin layer of H-rich fluid from the top of the core. We assume uniform mixing throughout the core for simplicity. To first order, the gravitational energy associated with instead forming a thin layer is independent of the thickness of the layer given a specific mass flux. Therefore, hydrogenation is physically analogous (in a reverse sense) to novel chemical processes—precipitation of MgO [15, 16] and/or SiO_2 [17] and transport of these light species from the core to the mantle—recently proposed to power Earth's dynamo. For example, $Q_H \sim 0.2$ TW is equivalent to a mass flux of ~8 x 10^{17} kg/Myr.

Formation of stable stratification: Concentrating hydrogen in a thin, stable layer may produce a seismic signature relative to a completely homogenous core. This might resemble the low density layer found at the top of Earth's core. Diffusion of oxygen (and possibly silicon) downwards from Earth's core/mantle boundary has been proposed to create ~100s km of stratification over geologic time [e.g., 18], which could explain seismic and geomagnetic observations. Following previous analyses [e.g., 19], we can estimate whether convection is likely to erode a H-rich layer in the core of Mars. If the layer thickness is h, then the maximum rate of erosion is $dh/dt = v_c/\mathrm{Ri}$, where $\mathrm{Ri} = (NR/v_c)^2$ is the Richardson number with N as the buoyancy frequency and $R \sim 1700$ km as the radius of the core. Compared to the situation on Earth, convective erosion is ~40 times more efficient on Mars because the core is half as large and hydrogen has two orders of magnitude larger diffusivity than oxygen (so h is larger and thus N is smaller). Still, the predicted erosion depth for a layer of pure FeH over Fe is <5 km, which may imply that stratification is quite likely. Convection can only efficiently mix fluids that have small differences in the molarity of hydrogen.

Consequences of a hydrogen-rich core: Models of the interior structure of Mars typically assume that sulfur is the only light element present in the core [e.g., 9, 10]. Oxygen and silicon are probably immiscible (especially in the presence of sulfur) under the pressure/temperature conditions relevant to Mars. However, hydrogen has been recognized as a plausible constituent of the core since iron and hydrogen are miscible at pressures >10 GPa [11]. Hydrogen depresses the liquidus temperature by as much as 700 K for pure FeH compared to Fe, which increases the amount of total cooling required to form an inner core. This diminishes the motivation to invoke >13 wt% of sulfur—several times the abundances considered geochemically reasonable for Earth's core—to keep the core partially liquid until present day and explain the observed mass, radius, and moment of inertia. Whereas models that consider only sulfur in the core predict that an inner core and stratification in the outer core would often occur together [20], hydrogenation provides a mechanism to produce stratification but no inner core. Overall, this study demonstrates the importance of core/mantle interactions to dynamical processes that are relevant to surface conditions.

References: [1] Acuña et al. (1999) *Science* 284, 790-3. [2] Stevenson (2003) *EPSL* 208, 1-11. [3] Williams & Nimmo (2004) *Geology* 32, 97. [4] Elkins-Tanton et al. (2005) *EPSL* 236, 1-12. [5] Arkani-Hamed & Olson (2010) *GRL* 37, 1-5. [6] Pearson et al. (2014) *Nature* 507, 221-4. [7] Shibazaki et al. (2009) *EPSL* 287, 463-70. [8] Sun et al. (2015) *GRL* 42, 6582-9. [9] Rivoldini et al. (2011) *Icarus* 213, 451-72. [10] Khan et al. (2017) *JGR* in press. [11] Gudkova & Zharkov (2004) *PEPI* 142, 1-22. [12] Stevenson (1987) *GJI* 88, 311-9. [13] Labrosse (2015) *PEPI* 247, 36-55. [14] O'Rourke et al. (2017) *EPSL* 458, 263-72. [15] O'Rourke & Stevenson (2016) *Nature* 529, 387-9. [16] Badro et al. (2016) *Nature* 536, 326-8. [17] Hirose et al. (2017) *Nature* 99-102. [18] Brodholt & Badro (2017) *GRL* 44, 8303-10. [19] Landeau et al. (2016) *Nature Geo.* 1, 1-5. [20] Helffrich (2017) *PEPS* 4, 24.