SHORT-LIVED LUNAR DYNAMOS POWERED BY ACCRETION OF COLD IMPACTOR CORE MATERIAL.

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Introduction: Measurements of the lunar crust and returned lunar samples suggest that the Moon likely had an early magnetic field with high surface field intensities of 40 to 120 μT between 4.3 and 3.5 Ga followed by a period of weak field intensities on the order of a few μT that may have persisted until 1 Ga [1, 2]. Previous studies have investigated several mechanisms for producing a lunar magnetic field to match the inferred lunar paleorecord including models of core convection [3-9], mechanical stirring driven by differential rotation of the mantle and core due to large impacts [10], and mantle precession [11-13] as well as magnetization by impact plasmas [14]. To date, none of these mechanisms can individually reproduce both the intensity and lifetime of the inferred lunar magnetic field [2].

Furthermore, models of thermochemical core convection – the process most commonly recognized to sustain dynamos within planetary cores of rocky bodies – are substantially underpowered and have been unable to produce field intensities much greater than ~1 μT [4]. Here we propose a new mechanism for producing short-lived, high intensity magnetic fields in which cold, dense impactor material sinks to the lunar core-mantle boundary (CMB) and drives more vigorous convection of the lunar core, see Figure 1. We find that this mechanism could result in short-lived periods of high magnetic field strengths which could reach some of the unusually high field intensities found in the early lunar paleorecord.

Methodology: With the nominal range of viscosities expected for the lunar mantle ($10^{18} – 10^{22}$ Pa s) [15, 16], impactor core material will descend through the mantle in the Stokes regime and will not disperse further during descent [17]. For the impactor core material to remain in a layer above the CMB, a lighter constituent alloy such as sulfur is required to be present in sufficient concentrations in the impactor core material (i.e., 18 wt.% for the best fit lunar core density of 6193 kg/m^3 from [18]). We use of the simplified dynamo scaling relation of [4], which requires the simplifying assumption that the cold impactor core material surrounds the lunar core uniformly. We extrapolate from the size of the impactor material at the CMB to the initial impactor size by assuming that the impactor material at the CMB represents the entire core of a differentiated impactor with a ratio of 30 to 70 wt.% for the core and mantle, respectively [19].

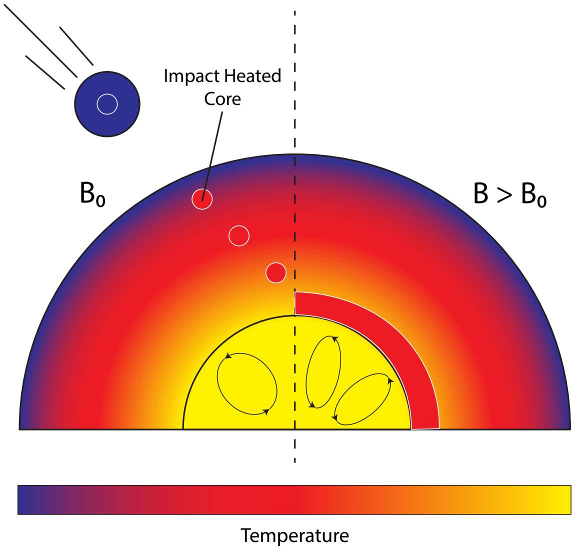

Figure 1: Schematic illustrating pre-impact (left) and post-impact (right) accretion event on the Moon. Left: The impact-heated core material from the impactor descends through the lunar mantle. Right: impactor core material is distributed laterally across the CMB. Idealized convection cells inside the lunar core are shown. As the cold impactor material remains above the lunar core (right), more vigorous thermochemical convection within the core is driven by the large temperature difference and higher heat flow at the CMB resulting in a larger magnetic field intensity.

For impactor core material to promote more vigorous convection within the hot, early lunar core with a temperature of 1800 – 2300 K, the impactor material must reach the CMB at a lower temperature than the displaced material which we assume to be a steady-state thermal boundary layer. This is influenced by the sinking speed of the impactor material and the conductive timescale as well as the level of heating upon impact. We assume that even for the extreme case in which the impactor core material is in thermal equilibrium with the mantle as it sinks, it will still be cooler than the thermal boundary layer material it displaces and therefore cooler than the core.

The amount of additional energy transported across the CMB due to the cooler surrounding impactor material is determined by the thermal evolution of the overlying layer of impactor material. To quantify the maximum effect of the impactor material on the lunar core dynamo, the impactor material at the CMB is assumed to be exclusively heated by the core and the lunar core's temperature is assumed to be constant. This allows us to solve the 1-D thermal diffusion equation in the spherical domain to determine analytically the heat flux across the CMB as a function of the thickness of the layer of impactor material and the initial temperature difference between the impactor material and the lunar core. We then employ the dynamo scaling law of [20] as modified by [4] to approximate the field strength in time for a particular impactor size and temperature difference between the lunar core and the impactor material. This is shown for impactor sizes corresponding to the Korolev, Orientale, Imbrium, and SPA basin sizes and temperature differences of 100 and 500 K in Figure 2. The impactors are 43, 108, 160, and 315 km in diameter, respectively based on basin sizes from [21] and the scaling law of [22] assuming a 45 degree impact at 10.6 km/s.

Results & Discussion: Overall, we find that even the smallest expected differentiated impactors (40 km in diameter [23]) are able to produce short-lived fields with strengths between 40 and 100 µT for the first two days of their evolution as long as they are at least 100 K cooler than the lunar core. Since the cooling timescale of mare basalt samples on the surface is hours to days [1, 4], these high-intensity fields last long enough to be recorded if a given sample cooled on the surface at the time at which the impactor material reaches the lunar CMB.

Larger impactor sizes and temperature contrasts allow for high intensities to be maintained over longer timescales. For impactors larger than 65 km in diameter, field strengths above 15 µT can be supported for a year as long as the impactor material is at least 100 K cooler than the lunar core. Impactors with diameters of about 80 km and larger with temperatures at least 600 K below the lunar core can support field strengths above 35 µT for a year.

Although field strengths are unlikely to remain in the high intensity regime past one year, they may remain higher than a pre-impact lunar dynamo strength for much longer. Impactors with diameters greater than 100 km can maintain field strengths above 10 µT for tens to hundreds of years. For our full range of impactor sizes and an initial temperature difference of 500 K, the adiabatic heat flux remains less than 1% of the enhanced heat flux for between two months and 1,200 years.

The relatively small impactor sizes required to produce short-lived fields of high intensity suggest that impacts may have a signature in the lunar paleorecord. Any such signatures in the paleorecords of the Moon or other inner solar system bodies could be used to inform the bombardment history of the early solar system.

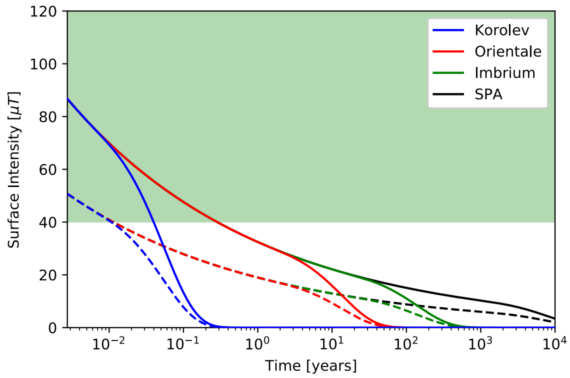

Figure 2: Magnetic field surface intensities (µT) versus time (years) for impactor sizes corresponding to the Korolev, Orientale, Imbrium, and SPA basin sizes (43, 108, 160, and 315 km in diameter, respectively). Initial temperature differences between the impactor material and the lunar core of 100 and 500 K are shown with dashed and solid lines, respectively. The green rectangle covers field intensities of 40 – 120 µT, considered to be the field strengths associated with the high-intensity era of the lunar paleorecord.

References: [1] Weiss, B. P. & S. M. Tikoo (2014) *Science, 346,* 1198-1208. [2] Tikoo, S. M. et al. (2017) *Sci. Adv., 3,* e1700207. [3] Evans, A. J. et al. (2014) *JGR, 119,* 1061-1077. [4] Evans, A. J. et al. (2018) *GRL, 45,* 98-107. [5] Konrad, W. & T. Spohn (1997) *Adv. Space Res., 19,* 1511-1521. [6] Laneuville, M. et al. (2013) *JGR, 118,* 1435-1452. [7] Scheinberg, A. et al. (2015) *Icarus, 254,* 62-71. [8] Stegman, D. R. et al. (2003) *Nature, 421*, 143-146. [9] Zhang, N. et al. (2013) *JGR, 118,* 1789-1804. [10] Le Bars, M. et al. (2011) *Nature, 479,* 215-218. [11] Dwyer, C. A. et al. (2011) *Nature, 479,* 212-214. [12] Stanley, S. et al. (2017) *LPS XLVIII Abstract* #1462. [13] Stys, C. & M. Dumberry (2020) *JGR, 125,* e2020JE006396. [14] Oran, R. et al. (2020) *Sci. Adv., 6*, eabb1475. [15] Hirth, G. & D. Kohlstedt (2003) *Geophys. Monogr. Ser., 138,* 83-105. [16] Zhang, N. et al. (2013) *JGR, 118,* 1789-1804. [17] Samuel, H. (2012) *EPSL, 313-314,* 105-114. [18] Matsuyama, I. et al. (2016) *GRL, 43,* 8365-8375. [19] Marchi, S. et al. (2018) *Nat. Geosci., 11,* 17-81. [20] Christensen, U. R. et al. (2009) *Nature, 457,* 167-169. [21] Neumann, G. A. et al. (2015) *Sci. Adv., 1*, e1500852. [22] Johnson, B. C. et al. (2016) *Icarus, 271,* 350-359. [23] Hevey, P.J. & I.S. Sanders (2006) *Meteor. & Planet. Sci., 41,* 95-106.