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A basal magma ocean dynamo to explain the early lunar magnetic field

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ABSTRACT

The source of the ancient lunar magnetic field is an unsolved problem in the Moon's evolution. Theoretical work invoking a core dynamo has been unable to explain the magnitude of the observed field, falling instead one to two orders of magnitude below it. Since surface magnetic field strength is highly sensitive to the depth and size of the dynamo region, we instead hypothesize that the early lunar dynamo was driven by convection in a basal magma ocean formed from the final stages of an early lunar magma ocean; this material is expected to be dense, radioactive, and metalliferous. Here we use numerical convection models to predict the longevity and heat flow of such a basal magma ocean and use scaling laws to estimate the resulting magnetic field strength. We show that, if sufficiently electrically conducting, a magma ocean could have produced an early dynamo with surface fields consistent with the paleomagnetic observations.

1. Introduction

Apollo samples of lunar crust indicate a surface field of $\sim$77 μT that was present between 4.2 and 3.56 Ga and likely continued at a lower magnitude for billions of years (Tikoo et al., 2017; Garrick-Bethell et al., 2017; Weiss and Tikoo, 2014; Suavet et al., 2013; Shea et al., 2012). Considering the large uncertainty in the paleomagnetic field inferred from measurement of these samples, a model that predicts at least $\sim$35 μT during the early high-field epoch would be consistent with observation (Weiss and Tikoo, 2014). However, numerical modeling of an Earth-like dynamo driven by thermochemical core convection instead predicts magnitudes of $\lesssim$1 μT (Stegman et al., 2003; Evans et al., 2014, 2018; Laneuville et al., 2014; Scheinberg et al., 2015). Dwyer et al. (2011) considered tidal forcing as an additional energy source for the dynamo, while Le Bars et al. (2011) considered impact-driven differential rotation. These alternative mechanisms are more promising, but the estimated magnetic fields are still several times lower than measurements ($\lesssim$9 μT and 4 μT, respectively).

Obtaining the observed lunar field magnitude with a core dynamo is particularly challenging because of the proportionally small lunar core. Dipole fields decay as $1/r^3$, where r is the distance from the dipole center. On Earth, this means that the dipole field at the surface is smaller than the dipole field at the core–mantle boundary by a factor of $\sim(6370\text{ km})^3/(3400\text{ km})^3 = 6.6$. In contrast, given the lunar radius of 1737 km and a core radius in the range 200–380 km (Williams et al., 2014), the dipole field of a core dynamo is 100–700 times smaller at the surface than at the core–mantle boundary. However, if the dynamo originated in a region extending to within 800 km of the surface, the field would be reduced only by a factor of ten from the dynamo to the surface.

1.1. Basal magma ocean formation

The Moon is believed to have become largely molten during its formation due to heat of accretion, radiogenic heating, and tidal forces (Warren, 1985). This magma ocean may have comprised as much as the upper 1000 km of the Moon, and have remained molten for as long as 200 m.y. due to continued tidal interactions with Earth (Meyer et al., 2010). Fractional crystallization is also highly likely to have occurred. In this scenario, crystals forming in the magma ocean would have settled to the bottom and sequestered themselves from chemical equilibrium with the remaining liquid, causing an evolving liquid composition in the magma ocean. The final, highly evolved liquids, the 'ur-KREEP', would have contained the majority of all incompatible elements, not just the potassium, rare Earth elements, and phosphorus for which they are named. These last liquids would also contain the majority of the uranium and thorium, which, together

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with potassium, are responsible for most of the radiogenic heating on Earth.

Fractional solidification of the global lunar magma ocean would have resulted in a gravitationally unstable mantle profile (Spera, 1992). Solidification models show that the density contrast between the KREEP layer and the remaining mantle range from $\Delta\rho = 300$–$500\ \mathrm{kg\,m^{-3}}$, depending on the amount of interstitial fluid trapped within the solidified cumulate (Elkins-Tanton et al., 2011). These models further suggest a large amount of KREEP material, $2{\cdot}10^9\ \mathrm{km}^3$ or ~9% of the mantle by volume. Though initially located just underneath the plagioclase crust, at least some fraction of this material would have overturned due to gravitational instability and formed a dense, radiogenic mantle layer located just above the core. This fraction is poorly constrained; in Appendix A, we provide a basic model indicating anywhere from 0 to 1 is plausible. If settled in its entirety on the core–mantle boundary (CMB), the material would constitute a ~450 km-thick layer.

Previous studies adopted significantly smaller values of $\Delta\rho$ (45–90 $\mathrm{kg\,m^{-3}}$) (Stegman et al., 2003; Zhang et al., 2013; Scheinberg et al., 2015). In those studies, thermal expansion of the KREEP layer was sufficient to overcome the compositional density difference, causing the material to re-mix into the overlying mantle. In contrast, using $\Delta\rho = 300$–$500\ \mathrm{kg\,m^{-3}}$, the KREEP layer would remain sequestered at the CMB. Highly radiogenic and still near its solidus temperature, the layer would fully melt. The result would be a metalliferous basal magma ocean (BMO) that persists until convective heat loss across the overlying solid mantle causes ocean solidification. Indeed, recent modeling by Zhang et al. (2017) concluded that a relatively stable, partially molten, ilmenite-bearing cumulates layer may surround the lunar core to present day.

1.2. A basal magma ocean dynamo

Ziegler and Stegman (2013) showed that a basal magma ocean on Earth could potentially drive a dynamo if it is unstable to convection and has sufficient electrical conductivity, σ. This requirement is manifested in a magnetic Reynolds number, which must exceed a critical value for magnetic field generation to occur (Roberts, 2007): $Rm_c \gtrsim 10$. This parameter relates the Ohmic diffusion timescale to the convective timescale and is defined as $Rm = \mu_0 \sigma U L$, where U and L are the system's characteristic velocity and length scale, respectively, and μ_0 is the permeability of free space.

The principle difficulty of a magma ocean dynamo hypothesis is that silicate material has a far lower electrical conductivity than a metallic iron-alloy core. However, the BMO is expected to have a relatively large electrical conductivity due to its molten nature and its particularly high titanium and iron contents (van Kan Parker et al., 2012; Khan et al., 2014; Elkins-Tanton et al., 2011). Using olivine melts at lunar mantle conditions as a lower bound (Pommier et al., 2015), we anticipate $\sigma > 80$ S/m. The presence of Fe–Ti oxides in terrestrial rocks is observed to increase the electrical conductivity up to three orders of magnitude (Bartetzko et al., 2005), suggesting an upper bound of $\sigma \sim 10^4$ S/m. More recently, experimental work by Lin et al. (2017a) and Lin et al. (2017b) argues for a substantial water component in the Moon and suggests that the deep magma ocean was extremely iron rich. Sufficient conductivity is thus plausible, though it cannot be definitively inferred from available evidence. Fortunately, the basic BMO dynamo hypothesis can be studied regardless of the uncertainty in lunar interior composition.

In this study, we estimate the longevity and heat flow of such a basal magma ocean, calculate the critical conductivity σ_{crit} above which dynamo action can be maintained, and predict the resulting magnetic field strength.

2. Methods

To simplify the analysis, we assume that three events have already occurred: (1) the magma ocean has solidified; (2) mantle overturn has caused some percentage of the dense layer to fall to the CMB while a fraction f_{KREEP} remained stuck directly beneath the plagioclase lid; and (3) sufficient heating to completely remelt this fallen material has occurred. We initiate the models at 4.2 Ga, when paleomagnetic observations indicate the field began and after which overturn and melting has occurred. (The time of initialization affects only the model's radiogenic heating values. If the model began at 4.3 Ga instead, initial radiogenic heating levels would be 5% higher.) As initial conditions, we assume a fully molten basal magma ocean at 1700 K, a homogeneous temperature of 1500 K within the solid cumulate mantle, and a conductive thermal profile in the upper 250 km. The thickness of this conductive layer is somewhat arbitrary, but is only an initial condition; once a simulation is launched, mantle convection will result in a thermal profile consistent with the input parameters of that simulation.

Radiogenic heating within the BMO will initially cause a temperature increase. When it cools back to the initial temperature, the model run is stopped. Furthermore, the fully molten core is presumed to contain enough sulfur ($\gtrsim$4%) that its liquidus temperature, which is strongly dependent on sulfur content, is never reached during a model run. These two model restrictions enable us to avoid the additional complexity (and associated assumptions) of solidification of the basal magma ocean and core while still determining the ocean's lifetime and potential dynamo action. Melting of the overlying solid cumulate materials and their assimilation into the BMO is similarly neglected since such melt would be buoyant compared to the surrounding solid mantle.

2.1. Solid mantle model

We use the spherical finite-element model CitcomS to simulate convection in the solid portion of the mantle. The model assumes that the mantle, excluding the basal magma ocean (BMO), is an incompressible spherical shell and that variations in density affect the equations of motion only through changes in buoyancy (the Boussinesq approximation). Mantle evolution is computed by iteratively solving the conservation equations for mass, momentum, and energy (see Zhong et al., 2008, for details). The numerical grid consists of 12 regions with $36 \times 36 \times 48$ elements each, providing a vertical and average horizontal resolution of approximately 30 km. Our lowest-viscosity (and therefore least resolved) simulation was repeated with 2.4× the volumetric resolution ($49 \times 49 \times 65$). The resulting fields were within 2%, indicating sufficient resolution.

The lunar surface temperature is held constant at 250 K. The upper 60 km is buoyant and thus behaves as a cold, stagnant lid crust. The temperature at the magma ocean boundary, T_{B}, is coupled to the BMO model described in the next section. The material parameters employed in the model are listed in Table 1. Our chosen geometry and parameter values are consistent to within 1% with present-day mass and moment of inertia measurements (Williams et al., 2014).

Solid mantle viscosity is temperature-dependent and determined by a Newtonian Arrhenius equation:

$$\mu = \mu_0^* \exp\left[\frac{E}{R_g T} - \frac{E}{R_g T_{ref}}\right], \tag{1}$$

where $E = 150$ kJ/mol is the effective activation energy, R_g is the ideal gas constant, and reference temperature T_{ref} is 1300 °C (Christensen, 1984; Karato and Wu, 1993). We vary the reference viscosity over $\mu_0^* = 10^{18}$–10^{20} Pa s.