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A long-lived lunar dynamo powered by core crystallization

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ABSTRACT

The Moon does not possess an internally generated magnetic field at the present day, but extensive evidence shows that such a field existed between at least 4.2 and 3.56 Ga ago. The existence of a metallic lunar core is now firmly established, and we investigate the influence of inner core growth on generating a lunar core dynamo. We couple the results of a 3-D spherical thermochemical convection model of the lunar mantle to a 1-D thermodynamic model of its core. The energy and entropy budget of the core are computed to determine the inner core growth rate and its efficiency to power a dynamo. Sulfur is considered to be the main alloying element and we investigate how different sulfur abundances and initial core temperatures affect the model outcomes. For reasonable initial conditions, a solid inner core between 100 and 200 km is always produced. During its growth, a surface magnetic field of about 0.3 μT is generated and is predicted to last several billion years. Though most simulations predict the existence of a core dynamo at the present day, one way to stop magnetic field generation when the inner core is growing is by a transition between a bottom–up and top–down core crystallization scheme when the sulfur content becomes high enough in the outer core. According to this hypothesis, a model with about 6 to 8 wt.% sulfur in the core would produce a 120–160 km inner core and explain the timing of the lunar dynamo as constrained by paleomagnetic data.

1. Introduction

The Moon does not possess an active magnetic field today, but magnetic anomalies originating from its crust are observed at its surface (Purucker and Nicholas, 2010; Tsunakawa et al., 2010; Hood et al., 2013) and some samples returned from the surface possess a natural remanent magnetization (Dyal et al., 1970). Paleomagnetic studies dating from the Apollo era suggest a global field that lasted between 3.8 and 3.6 Ga ago (e.g., Cisowski et al., 1983), and furthermore, that the field strengths were as high as 100 μT (Fuller and Cisowski, 1987 and references therein). For comparison, the present day field of the Earth is on the order of 50 μT.

Lunar rocks are poor magnetic recorders and their thermal histories are often uncertain, therefore many of the Apollo era estimates should be used with caution (Lawrence et al., 2008; Tikoo et al., 2012a). Nevertheless, the current view suggests that a magnetic field of several tens of μT was present at the surface of the Moon between 4.2 and 3.56 Ga ago (Garrick-Bethell et al., 2009; Shea et al., 2012; Tikoo et al., 2012a; Suavet et al., 2013). The lack of data before 4.2 Ga ago implies that the dynamo could be older, and there is also no definitive proof that the dynamo shut down 3.56 billion years ago, only that the surface magnetic field beyond that time was weaker (Tikoo et al., 2012b). Although there is no observed dynamo today, recent studies appear to indicate that a sample younger than 3.3 Ga, and maybe as young as ∼1.3 Ga (Fagan et al., 2013), acquired its primary magnetization from a dynamo field (Tikoo et al., 2014).

The most plausible mechanism for generating long lasting planetary magnetic fields is a core dynamo (Stevenson et al., 1983). Global planetary magnetic fields in terrestrial planets are generated by convection in their liquid outer cores. When more heat is extracted from the core than can be conducted along the adiabat, motion is triggered by thermal instabilities. The strength of the magnetic field is governed by the vigor of convection and the thickness of the convecting shell. However, the Moon is a small body and previous mantle thermal evolution studies found that heat extraction from the core is not large enough to produce a magnetic field for more than a few hundred million years, which is about ten times too short when compared to the paleomagnetic results (Konrad and Spohn, 1997; Laneuville et al., 2013). Models

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with an initially stratified mantle, where KREEP- and ilmenite-rich magma ocean cumulates surrounded the lunar core, were able to produce a thermally induced core dynamo between 4.1 and 3.5 Ga ago (Stegman et al., 2003; Zhang et al., 2013). However, though this timing is marginally consistent with current observations, the lack of a KREEP signature in the titanium-rich mare basalts is potentially inconsistent with the underlying assumptions of this model. A recent study by Evans et al. (in press) proposed that the existence of a wet and initially stratified mantle could prolong the magnetic field era, but did not explicitly account for the history of mare volcanism.

It has been proposed that a lunar dynamo could also have been driven by differential rotation between mantle and core, induced either by precession of the mantle spin axis (Dwyer et al., 2011) or by changes in the rotation rate of the solid mantle following large impacts (Le Bars et al., 2011). In both cases, differential rotation induces large-scale flow in the core, which could have powered a lunar dynamo. The precession induced magnetic field is predicted to last from about 4.2–2.7 Ga ago with intensities of about 1 μT, but may have troubles explaining paleomagnetic data outside of this range as new studies are published. The impact scenario predicts the existence of a magnetic field lasting about 10 thousand years. This could explain the existence of magnetic anomalies associated with the interiors of large impact basins, but cannot explain a dynamo field younger than the Orientale impact at about 3.7 Ga. We note here that it is unclear whether the efficiency and magnetic properties of such dynamos are similar to standard thermo/chemical ones.

As an alternative to previously proposed models, we study the influence of inner core growth on dynamo generation. This scenario has been studied in the case of the terrestrial planets (e.g., Stevenson et al., 1983) and asteroids (Nimmo, 2009), but has to date never been proposed for the Moon. In this model, compositional buoyancy due to the release of light elements at the inner core boundary helps to sustain convection in the outer core against dissipation, even when the heat extracted by the mantle is smaller than what could be conducted along the core's adiabat. This hypothesis has not been tested for the Moon before, in part because the bare existence of a lunar core – let alone an inner core – was debated (e.g., Wieczorek et al., 2006). A range of datasets has been used to constrain the lunar core size and state (including seismic analyses), and its radius is believed to lie between 250 and 450 km (Garcia et al., 2011; Weber et al., 2011) with at least some portion being partially molten at the present time (Williams et al., 2001). Recent lunar thermal evolution studies (Laneuville et al., 2013; Zhang et al., 2013) have suggested that core crystallization should indeed occur, and we therefore investigate this process in more detail in this study. We start by presenting our model in Section 2, which includes the coupling of a core energetics model to a 3-D mantle thermal evolution model. The predictions of our model are presented in Section 3, and we discuss some of the implications of this model in Section 4.

2. Lunar core evolution and magnetic field scaling

In order to estimate the strength of the surface magnetic field, the power available to drive the dynamo has to be estimated (Christensen and Aubert, 2006; Aubert et al., 2009). This power is directly linked to the sum of the thermal and chemical buoyancy forces within the core. As core and mantle evolution are coupled, we first need to model the thermal evolution of the mantle. The growth of the inner core is then obtained through the core energy budget, which is coupled to the bottom boundary condition of the mantle. Finally, we use the entropy budget to determine the part of the power available to dynamo action and a scaling law to relate the magnetic field strength and power available from core convection.

Table 1
Parameters used for the mantle thermal evolution simulations (see Laneuville et al., 2013).

Symbol	Description	Value
R_p	Moon radius	1740 km
R_c	Core radius	330 km
D_c	Crustal thickness	40 km
Ω_K	PKT angular radius	40°
D_K	KREEP layer thickness in PKT	10 km
T_{surf}	Surface temperature	250 K
T_0	Reference temperature	1600 K
η_0	Reference viscosity	10^{21} Pa s
η_{max}	Maximum viscosity	10^{28} Pa s
E	Activation energy	3×10^5 J mol^{-1}
L	Mantle latent heat of melting	6×10^5 J kg^{-1}
$c_{p,m}$	Mantle specific heat capacity	1000 J kg^{-1} K^{-1}
k_c	Crust thermal conductivity	1.5 W m^{-1} K^{-1}
k_m	Mantle thermal conductivity	3 W m^{-1} K^{-1}
κ_0	Reference thermal diffusivity	10^{-6} m^2 s^{-1}
ρ_0	Reference density	3400 kg m^{-3}
$\alpha_{0,m}$	Thermal expansivity	2×10^{-5} K^{-1}
g	Surface gravity acceleration	1.62 m s^{-2}

We model the thermal evolution of the Moon using the Gaia 3-D thermochemical convection code with a temperature-dependent viscosity in a spherical shell (Hüttig and Stemmer, 2008). We follow closely Laneuville et al. (2013) and consider both core cooling and time-dependent radioactive decay of heat sources. We solve the conservation equations of mass, momentum and energy for an incompressible fluid under the Boussinesq assumption, with free-slip boundary conditions at both the surface and core mantle boundary. The consumption of latent heat through melting is taken into account assuming a peridotitic mantle. Mantle depletion from melting also adds a buoyancy source, which is then monitored by tracer particles, varying by 60 kg m^{-3} between 0 and 30% depletion (the latter corresponding to harzburgite). However, heat source partitioning through mantle depletion is not considered. The rheology is Newtonian with a reference viscosity of 10^{21} Pa s at 1600 K, corresponding to a dry mantle. Gravity is assumed constant throughout the mantle, which somewhat overestimates the buoyancy sources in the lower mantle. A table with relevant parameters for the mantle thermal evolution simulations can be found in Table 1. For a complete description of the model, as well as discussion about possible limitations to the model, the reader is referred to Laneuville et al. (2013).

A 1-D geometry is used to model the core because we are interested only in its long-term, averaged evolution rather than in the short-term perturbations associated with core convection. We ignore explicitly any potential complications that may arise due to a non-uniform core mantle boundary heat flow (Glatzmaier et al., 1999; Takahashi and Tsunakawa, 2009). Fig. 1 is a schematic of our core model, showing the core temperature profile and the liquidus temperature in the outer core. The liquid outer core is assumed to be well-mixed, and thus to follow an adiabatic temperature profile. This approximation is not valid when core convection is not occurring, such as after the termination of an initial thermal dynamo stage and before the onset of core crystallization. Nevertheless, this should affect only the time at which core crystallization occurs by a few 100 million years. The solid inner core is assumed to be isothermal due to its high thermal conductivity.

The inner core size is computed by comparing the adiabatic temperature profile in the liquid core to the liquidus of the iron alloy. As the inner core grows, the outer core becomes enriched in sulfur and the liquidus temperature decreases. A simple mass balance provides $\chi(r_i) = \chi_0/(1 - f^3)$, where χ_0 and χ are the sulfur mass fraction in the core initially and as a function of in-