Earth and Planetary Science Letters 404 (2014) 89–97

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Decline of the lunar core dynamo

Sonia M. Tikoo [a,*], Benjamin P. Weiss [a], William S. Cassata [b], David L. Shuster [c,d], Jérôme Gattacceca [a,e], Eduardo A. Lima [a], Clément Suavet [a], Francis Nimmo [f], Michael D. Fuller [g]

[a] Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology (MIT), 77 Massachusetts Avenue, Cambridge, MA 02139, USA
[b] Chemical Sciences Division, Lawrence Livermore National Laboratory, 7000 East Avenue, Livermore, CA 94550, USA
[c] Department of Earth and Planetary Science, University of California, Berkeley, CA 94720, USA
[d] Berkeley Geochronology Center, 2455 Ridge Road, Berkeley, CA 94709, USA
[e] Centre Européen de Recherche et d'Enseignement, UM34, CNRS/Université Aix-Marseille, France
[f] Department of Earth and Planetary Sciences, University of California, Santa Cruz, CA 95064, USA
[g] University of Hawaii at Manoa, Honolulu, HI 96822, USA

ARTICLE INFO

Article history:
Received 19 February 2014
Received in revised form 24 June 2014
Accepted 10 July 2014
Available online 13 August 2014
Editor: J. Brodholt

Keywords:
Moon
core dynamo
paleomagnetism
planetary magnetism

ABSTRACT

Recent analyses of Apollo samples have demonstrated that a core dynamo existed on the Moon between at least 4.25 and 3.56 billion years ago (Ga) with surface field intensities reaching ∼70 μT. However, it is unknown when the Moon's magnetic field declined. Determining the temporal evolution of the dynamo is important because it constrains secular changes in power at the lunar core–mantle boundary and, by implication, the Moon's thermal and orbital evolution and the field generation mechanism. Here we present paleomagnetic data from several younger mare basalts which demonstrate that the surface magnetic field had declined precipitously to <∼4 μT by 3.19 Ga. It is currently unclear whether such a rapid decrease in field strength reflects either the cessation of the dynamo during this period or its persistence beyond 3.19 Ga in a drastically weakened state.

1. Introduction

A variety of geophysical and geochemical data have established that the Moon is a differentiated body with a small (∼350 km diameter) liquid outer core (Weber et al., 2011; Wieczorek et al., 2006). Remanent magnetization in lunar rocks and the crust indicates that there were substantial ancient magnetic fields on the surface of the Moon (Fuller and Cisowski, 1987). Although impact-generated plasmas are a potential source of magnetic fields (Hood and Artemieva, 2008), recent paleomagnetic studies of Apollo samples and the association of central magnetic anomalies with Nectarian impact basins indicate that a lunar core dynamo existed between at least ∼4.25 and 3.56 billion years ago (Ga) with surface field intensities of ∼30–110 μT (Cournède et al., 2012; Garrick-Bethell et al., 2009; Garrick-Bethell and Weiss, 2013; Hood, 2011; Shea et al., 2012; Suavet et al., 2013). The lack of detailed paleomagnetic studies of lunar rocks younger than 3.56 Ga has meant that it is currently unclear when the dynamo weakened and ultimately ceased (Fuller and Cisowski, 1987; Tikoo et al., 2012).

The intensity and longevity of thermally convective dynamos depend on the superadiabatic heat flux at the core–mantle boundary (Christensen et al., 2009). However, estimates for this adiabatic threshold are poorly constrained. Lunar thermal evolution models suggest that a thermally convective lunar dynamo can persist until sometime between ∼3.7 and ∼2.5 Ga for adiabatic thresholds ranging from 10 to 3 $\mathrm{mW\,m^{-2}}$ (Evans et al., 2014; Konrad and Spohn, 1997; Laneuville et al., 2013; Stegman et al., 2003). This has motivated alternative proposals that the core dynamo was mechanically powered by differential rotation of the lunar mantle, driven by either large impacts (Le Bars et al., 2011) or precession (Dwyer et al., 2011), or thermochemically driven by core crystallization (Laneuville et al., 2014; Soderlund et al., 2013; Zhang et al., 2013). However, because impact-driven changes in rotation are unlikely to have generated a core dynamo after the final large basin-forming impact at 3.72 Ga (Suavet et al., 2013), the persistence of the dynamo until at least 3.56 Ga supports precession or core crystallization as the main field source at this time. By comparison, mantle precession (Dwyer et al., 2011) and core crystallization (Laneuville et al., 2014; Soderlund et al., 2013; Zhang et al., 2013) dynamos may be capable of persisting until as

* Corresponding author at: Department of Earth and Planetary Science, University of California, Berkeley, CA 94720, USA. Tel.: +1 573 382 6948; fax: +1 510 644 9201.
E-mail address: smtikoo@berkeley.edu (S.M. Tikoo).

http://dx.doi.org/10.1016/j.epsl.2014.07.010
0012-821X/

late as a few hundred million years ago under certain conditions (Supplementary material).

Constraints on when the dynamo declined in intensity would constrain the power sources and, by implication, the field-generating mechanism(s) for the dynamo. In particular, given that at least thermally convective and perhaps also precession dynamo field intensities are thought to scale with the available (i.e., superadiabatic, not total) power (Christensen et al., 2009), evidence for a decline in the field intensity after 3.56 Ga could constrain the thermal and orbital evolution of the Moon.

Two key impediments to lunar paleomagnetic studies are that the majority of lunar samples have poor magnetic recording properties (Tikoo et al., 2012) and complex thermal and shock deformational histories (Supplementary material). Therefore, many earlier paleomagnetic studies are unreliable both in inferred paleointensities and age of magnetization. Consequently, since the end of the Apollo era, there have been two competing hypotheses about the state of the late lunar magnetic field. Fuller (1998) suggested that there was no evidence for a dynamo after 3.72 Ga, whereas Runcorn (1996) proposed that the dynamo persisted until at least ∼3.2 Ga (the age of the youngest returned mare basalts). Distinguishing between these possibilities requires a combination of paleomagnetic and petrographic studies and thermochronometry to constrain the extent of post-formational shock and thermal processes that could have modified any primary magnetization in samples. Two recent studies of young (<3.56 Ga) samples did not come to firm conclusions about the origin of their natural remanent magnetization (NRM). Cournède et al. (2012) studied ∼3.3 Ga mare basalt 12002, but did not confidently isolate a primary remanence (Supplementary material). Lawrence et al. (2008) suggested that the NRM of cataclastic anorthosite 60015 may not have been acquired from a dynamo field (i.e., their data imply any ambient field was <5 μT), but both this paleointensity constraint and the age to which it applies are uncertain because the sample may have been shocked or thermally demagnetized well after its $^{40}Ar/^{39}Ar$ plateau age of 3.46 Ga (Supplementary material). With the goal of resolving the state of the lunar dynamo at 3.2–3.3 Ga, we conducted a new paleomagnetic study of Apollo 12 and 15 mare basalts which accounts for secondary shock and thermal effects. Our goal is to constrain the paleointensity of the late lunar field.

2. Samples

We chose to focus on mare basalt samples 12022 and 15597 because Apollo-era analyses and our own measurements found that these rocks have unusually high fidelity magnetic recording properties relative to other young (<3.6 Ga) mare basalts (Supplementary material). Sample 12022 also offers a fortuitous opportunity for lunar paleomagnetic studies because it was sawn at Johnson Space Center (JSC) into multiple mutually oriented blocks in April and May 1970, just five months after return from the Moon (Supplementary material). These blocks were subsequently stored in unknown, almost certainly differing orientations without magnetic shielding. This early saw-cutting and subsequent long-term storage enables a test of whether NRM in 12022 was acquired as viscous remanent magnetization (VRM) contamination from long-term exposure to the terrestrial field (Supplementary material) or is pre-terrestrial in origin.

2.1. Petrographic descriptions and ages

12022 is a medium-grained porphyritic ilmenite basalt (Brett et al., 1971; James and Wright, 1972; McGee et al., 1977; Neal et al., 1994; Weill et al., 1971). The sample has an $^{40}Ar/^{39}Ar$ plateau age of 3.194 ± 0.025 Ga (Alexander et al., 1972; Supplementary material). The phenocrysts in 12022 are predominantly 1–2 mm

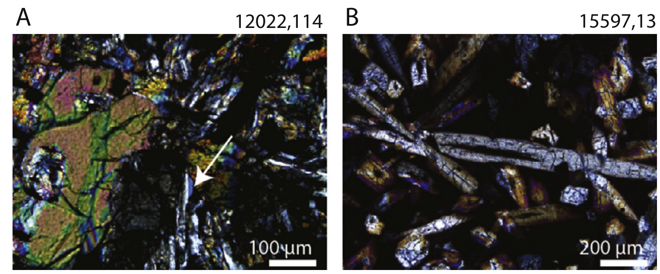

Fig. 1. Photographs of 30 μT thin sections in transmitted light with crossed polars. (A) 12022,114 and (B) 15597,13. White arrow points to a plagioclase crystal used for cooling rate determinations. Pyroxene is present as large phenocrysts displaying high order interference colors in sample 12022; it also appears as large laths within a glassy matrix in sample 15597.

diameter pyroxene crystals and ∼300 μm diameter olivine crystals (Fig. 1A). Its matrix consists of 0.05–1 mm diameter feldspar laths, 30–200 μm long ilmenite laths, 600–800 μm diameter pyroxene grains and trace amounts of aluminosilicate glass.

15597 is a vitrophyric quartz-normative basalt (Ryder, 1985; Weigand and Hollister, 1973). It has whole-rock Rb and Sr isotope ratios consistent with a model age of ∼3.3 Ga (Compston, 1972) and an $^{40}Ar/^{39}Ar$ plateau age of ∼3.1–3.5 Ga (Kirsten et al., 1973). The sample contains elongated (up to 300 μm long) pyroxene phenocrysts in a brown glassy matrix with sparse vesicles ranging from 10 to 500 μm in size (Fig. 1B).

Our petrographic study indicates that both samples lack evidence for shock (peak pressures <5 GPa): plagioclase shows no mechanical twinning, fracturing, or alteration to maskelynite, and there is no undulatory extinction in olivine or pyroxene (Stöffler et al., 2006). Therefore, shock demagnetization or remagnetization of any existing primary thermoremanent magnetization (TRM) is likely modest and confined to low coercivity grains (Supplementary material).

2.2. Ferromagnetic mineralogy

Our electron microscopy analyses found that metal grains in 12022 have compositions of $Fe_{1-x}Ni_x$ with $0.05 < x < 0.19$ and no detectable P (<0.03% by mass). A previous study observed a similar compositional range along with an additional population of nearly pure Fe grains (Reid et al., 1970). Metal grains in 15597 have compositions of $Fe_{1-x}Ni_x$ with $0.02 < x < 0.12$ (with 2 out 11 analyzed grains having $x < 0.03$) and trace P (0.02–0.08% by mass). These compositions, the samples' fast cooling rates (Section 2.3) and the homogeneity of Ni contents and lack of exsolution textures within most metal grains collectively indicate that both kamacite (α-Fe) (grains with $x < 0.05$) and martensite (α_2-Fe) (grains with $x > 0.05$) are the main ferromagnetic minerals in both rocks (Supplementary material). This is supported by our observation that laboratory anhysteretic remanent magnetization (ARM) and isothermal remanent magnetization (IRM) unblock mostly by ∼600 °C [close to the austenite-finish temperatures (i.e., martensite recrystallization temperatures) expected for the observed range of Ni abundances] with a small fraction of IRM persisting to higher temperatures (consistent with kamacite's 780 °C Curie temperature) (Supplementary material). The presence of kamacite is also indicated by previous thermomagnetic analyses showing magnetization persisting to 780 °C (Helsley, 1971). Early-formed kamacite should acquire a TRM after cooling below its Curie temperature, while martensite should predominantly acquire a TRM after passing through the martensite-finish temperature (ranging from ∼600 to 120 °C and 600 to 350 °C for the observed compositions in 12022 and 15597, respectively; Swartzendruber et al., 1991).