DECLINE OF THE ANCIENT LUNAR CORE DYNAMO. S. M. Tikoo[1], B. P. Weiss[1], T. L. Grove[1], M. D. Fuller[2] [1]Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, MIT 54-825, 77 Massachusetts Ave., Cambridge, MA 02139, smtikoo@mit.edu. [2]Hawaii Institute of Geophysics and Planetology, University of Hawaii at Manoa, Honolulu, HI 96822.

Introduction: A variety of data indicate that the Moon is differentiated, with a ~240 km radius solid inner core surrounded by a ~350 km radius liquid outer core [1-4]. During the Apollo era, paleomagnetic studies of lunar samples suggested that an ancient core dynamo may have existed from ~3.9-3.2 billion years ago (Ga). The core field at 3.2 Ga was thought to be present but 1-2 orders of magnitude weaker than the at 3.8 Ga [5]. Recent analyses of samples 76535, 10020, and 10017 confirm that a lunar core dynamo generated a long-lived, stable magnetic field from at least 4.2 to 3.6 billion years (Ga) ago [6-8]. However, when the dynamo ultimately decayed remains unclear because retrieving high-fidelity paleointensities (i.e., the strength of the magnetizing field) for weak fields is difficult for the vast majority of lunar samples [9]. Thus far, no lunar samples from 3.3-3.2 Ga in age have been reported to unambiguously record a thermoremanence (TRM) acquired while cooling in the presence of a core dynamo-generated field. In fact, a recent study of a cataclastic anorthosite proposed an upper limit of 5 μT for the lunar field at 3.34 Ga [10].

In this study, we studied the natural remanent magnetization (NRM) of some of the highest magnetic fidelity young mare basalts to investigate the state of the lunar dynamo at 3.2-3.3 Ga. Our preliminary results indicate that the lunar dynamo likely produced at most weak (< 7 μT) to null surface fields at those dates.

Samples: When choosing lunar samples for the purpose of investigating the existence of a dynamo, several criteria should ideally be met. Samples should not be significantly shocked (< 5 GPa) and should have cooled from the Curie point of their magnetic carriers (780°C) to ambient surface temperatures slowly relative to the duration of an impact-generated field. These conditions will exclude transient impact fields as the magnetizing source for any stable magnetization. Impact fields are estimated to last a maximum of ~1 hour for the largest, basin-forming impacts [11]; at ~3.3 Ga, any impact fields likely only lasted at most seconds since most impacts at this time were much smaller. The samples used in this study meet these criteria.

12022 is a medium-grained ilmenite basalt with an $^{40}Ar/^{39}Ar$ plateau age of 3.18 ± 0.04 Ga [12]. Our petrographic study demonstrates that it displays no evidence of shock (peak pressures < 5 GPa) and that it expeienced late-stage cooling over ~2 days (following ref. [13]).

15556 is a fine-grained, highly vesicular olivine normative basalt. It has an $^{40}Ar/^{39}Ar$ age of ~3.4 Ga [14]. Our petrographic study indicates the sample is unshocked (peak pressures < 5 GPa) and experienced late-stage cooling over months.

15597 is a ~3.3 Ga [14] vitrophyric pigeonite basalt. Ref. [13] found that its late-stage cooling timescale was ~2 hours. Again, the sample displays no petrographic evidence of shock above 5 GPa.

NRM behavior: From each parent rock, we prepared numerous mutually oriented subsamples which were subjected to identical three axis alternating field (AF) demagnetization measurements up to 85 or 290 mT (following refs. [6,7]). Measures were taken during data processing to reduce noise from undesirable acquisition of spurious gyroremanent magnetization (GRM) and anhysteretic remanence (ARM).

12022. Most subsamples from 12022 exhibit a low coercivity (LC) magnetization component blocked up to ~8 mT and one or two medium coercivity (MC) components ranging from the end of the LC component up to ~25-85 mT (Fig. 1a). The LC and MC components were roughly unidirectional from subsamples collected from adjacent locations within the parent rock, but diverged from those collected from more distal areas of the parent rock (Fig. 1b).

15556. Subsamples of 15556 generally have two poorly defined components: an LC component blocked up to ~5 mT, and an MC component ranging from the end of the LC component to ~6-23 mT. All subsamples displayed noisy demagnetization behavior with scatter in moment magnitude and direction. The LC and MC magnetization directions were largely non-unidirectional across subsamples.

15597. AF demagnetization of 15597 revealed two poorly defined components: an LC component blocked up to ~5 mT, and a noisy MC component ranging between ~5 mT and ~20 mT. The LC and MC magnetization directions were inconsistently oriented between subsamples.

Fidelity of magnetic records*:* Since high coercivity magnetizations cannot easily be identified in 12022, 15556, and 15597, surface fields from any existing lunar dynamo at 3.2-3.3 Ga seem to have been too weak to retrieve stable magnetizations from these samples. To explore this issue further, following ref. [9], we determined that stable TRM records acquired in paleofields ≤ 7 μT are incapable of being successfully retrieved by AF methods for 12022, 15556, and 15597.

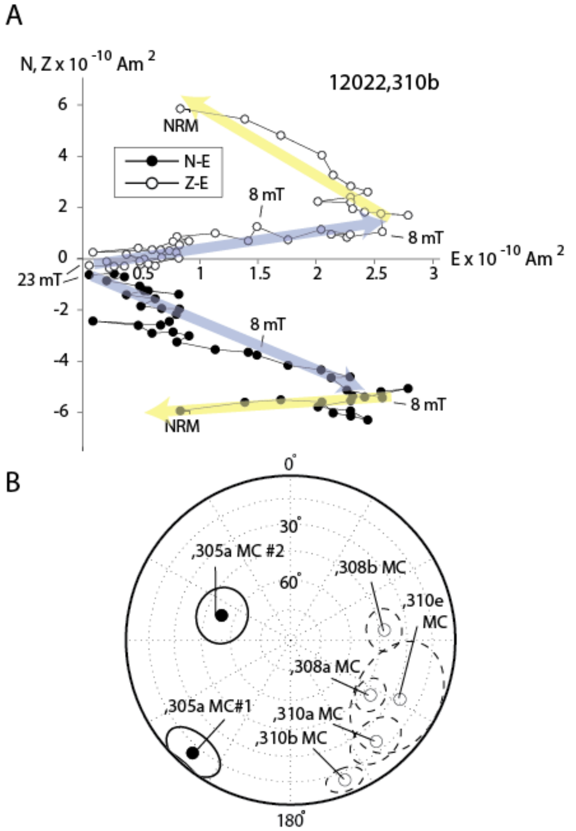

Fig. 1. (A) Two-dimensional projection of the NRM vectors of 12022,310b during AF demagnetization. Solid (open) symbols represent end points of magnetization projected onto the horizontal N-E (vertical Z-E) planes. (B) Equal-area plot showing directions of MC components identified in sample 12022.

Nearly all subsamples of 12022, 15556, and 15597 lack a stable magnetic component at AF fields above 25 mT. To investigate whether magnetization existed at higher coercivities, we used the ARM method (see refs. [6,7]) to compute paleointensities for the magnetic components in our samples. At low coercivities, we retreived substantial paleointensities (15 to 175 μT for the three samples). Above 25 mT, the slope of the NRM lost vs. ARM gained curve flattens, indicating a 7 ± 2 μT paleofield (Fig. 2). This value is an upper bound because it is within error of the sample's TRM fidelity limit (see ref. [9]). Other subsamples from 12022 yielded paleointensities ranging from -7 to 10 μT at high coercivities. We also observe this behavior in paleointensity plots for all other subsamples of 12022 as well as most subsamples of 15556 and 15597.

Conclusions: Our results suggest that, by 3.3 Ga, the lunar dynamo ceased or at least generated surface fields too weak (< 7 μT) to be retrieved by AF-based demagnetization and paleointensity methods on our samples. This contrasts strongly with the high paleointensities (60-70 μT) inferred for the Moon just 300-400 million years earlier [7,8]. Although Apollo-era paleointensity compilations have identified lunar paleofields ranging fom 0.1-10 μT at times ranging from 3.3 Ga to <200 Ma, we have recently argued that many or all of these values may also only be upper limits [9]. On the other hand, our upper limit of 7 μT is consistent with models which show that dynamo surface fields at 3.3-3.2 Ga may have ranged from 0.3-15 μT [15,16]. If a lunar dynamo existed at this time, thermal paleointensity techniques may be able to retrieve weak paleofields if sample alteration during the experiment can be mitigated.

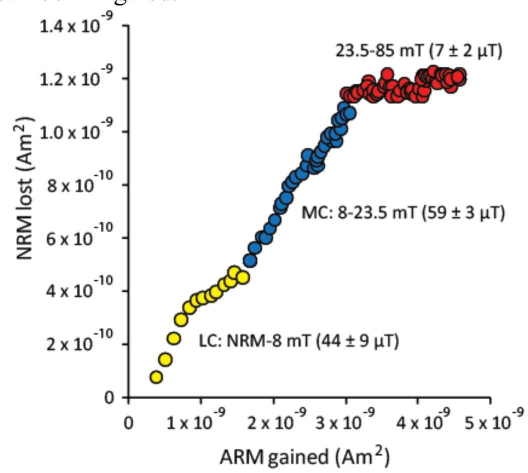

Fig. 2. ARM paleointensity plot for 12022,310b depicting NRM lost during AF demagnetization as a function of ARM gained during stepwise acquisition (AC field increasing to 85 mT, DC bias field = 0.2 mT) at an equivalent AC field. Labels note component type, AF levels, and paleointensities.

References: [1] Konopliv A. et al. (1998) *Science, 281*,1476-1480. [2] Hood L. et al. (1999) *Geophys. Res. Lett., 26,* 2327-2330. [3] Weber R. C. et al. (2011) *Science, 331,* 309-312. [4] Garcia R. F. et al. (2010) *EOS Trans. AGU Fall Meeting,* Abstract #GP43B. [5] Runcorn S. K. (1996) *Geochim. Cosmochim. Acta, 60, 1205-1208.* [6] Garrick-Bethell I. et al. (2009) *Science, 323,* 356-359. [7] Shea E. K. et al. (2011) *Science*, in press. [8] Suavet C. et al. (2012) *LPS XLIII*, submitted. [9] Tikoo, S. M. et al. (2012), submitted. [10] Lawrence K. et al. (2008) *PEPI, 168,*71-87. [11] Hood, L. L. and Artemieva N. A. (1987) *Icarus, 193,* 485-502. [12] Alexander E. C. et al. (1972) *LSC III,* 1787-1795. [13] Grove T. L. and Walker D. (1977) *LPS VIII,* 1501-1520. [14] Kirsten T. et al. (1972) *LSC III,* 1865-1889. [15] Dwyer C. et al. (2011) *Nature, 479,* 212-214. [16] Le Bars M. et al. (2011) *Nature, 479,* 215-218.