CONTROLLED-ATMOSPHERE THERMAL DEMAGNETIZATION AND PALEOINTENSITY ANALYSES OF LUNAR ROCKS. C. Suavet[1], B. P. Weiss[1], E. A. Lima[1], J. Gattacceca[2], and T. L. Grove[1], [1]Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139-4307, United States, csuavet@mit.edu, [2]Centre Européen de Recherche et d'Enseignement en Géosciences de l'Environnement, UM 34, CNRS, Aix-Marseille Université, Aix-en-Provence, France.

Introduction: Recent paleomagnetic studies of lunar rocks have established that the Moon had a dynamo magnetic field between at least 4.2 Ga and 3.56 Ga [1-4]. However, a central and as yet unrealized goal of extraterrestrial paleomagnetism has been to accurately measure the intensities of this dynamo field, which would help to constrain the core size, and the power requirements and generation mechanism of the lunar dynamo.

Records of these fields are provided by natural remanent magnetization (NRM) in the form of thermoremanence (TRM) acquired during cooling [5]. TRM is best characterized using stepwise thermal demagnetization and remagnetization methods because they reproduce the natural process of progressive magnetization by thermal unblocking and blocking [6].

A major limitation to this methodology has been that lunar rocks formed in environments that are orders of magnitude more reducing than at the Earth's surface. As a result, heating such rocks in the Earth's atmosphere induces oxidation reactions that alter the magnetic carriers, demagnetizing them via recrystallization rather thermal unblocking and also permanently changing their magnetic properties.

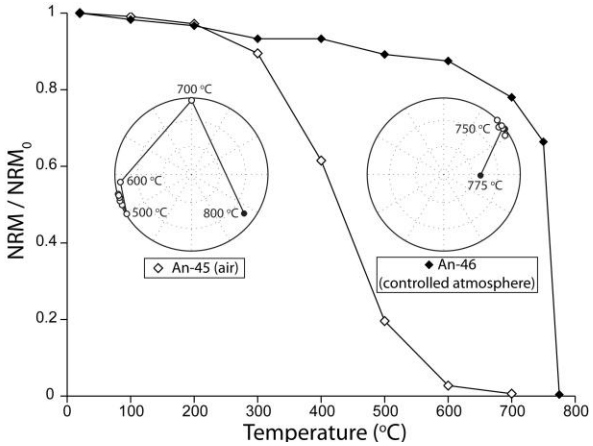

Fig. 1: Thermal demagnetization of lunar basalt analogs in air (open diamonds) and an oxygen fugacity-controlled atmosphere. Shown is the direction and intensity of magnetization normalized to the total TRM throughout the demagnetization. Insets show equal area plots of the magnetization direction, with closed and open symbols representing lower and upper hemispheres, respectively.

Previous attempts to prevent alteration of magnetic carriers – by heating the rocks in vacuum or in mixtures of inert and reducing gases – have failed due to methodological flaws (e.g. [7-10]).

Oxygen fugacity-controlled thermal demagnetization: We designed a controlled-atmosphere thermal demagnetization apparatus that prevents oxidation of metal-bearing rocks. Mixtures of H_2 and CO_2 are used to set the oxygen fugacity 1 log unit (bars) below the iron-wüstite buffer [11-12], appropriate for lunar basalts [13]. The system was calibrated by exploring the stability of fragments of pure iron in different gas mixtures. Tests were then conducted on mare basalt analogs [14]: thermal demagnetizations and a Thellier-Thellier paleointensity experiment showed that alteration is successfully prevented by the gas mixture (Fig. 1).

Thermal demagnetization of mare basalts: Apollo 11 mare basalts 10017, 10020 and 10049 have been shown to contain a record of the lunar dynamo carried by kamacite grains [2,4]. We conducted thermal demagnetization in controlled atmosphere on these samples, expecting to observe demagnetization close to the Curie temperature (780°C). However, the magnetization became directionally unstable at 150-250°C and rapidly decayed in intensity. Acquisition and demagnetization of anhysteretic remanent magnetization (ARM) before and after each heating step shows that the coercivity spectrum is not modified by heating, which confirms that controlled atmosphere prevents alteration. Therefore, the loss of magnetization at low temperature is not caused by alteration of the magnetic carriers. Partial thermoremanent magnetization (pTRM) experiments show a peak in pTRM acquisition at ~300°C. AF demagnetization of the pTRM shows that the magnetization acquired is of very high coercivity (median destructive field (MDF) >85 mT; Fig. 2). This peak was also observed in previous studies (e.g., [10]), which proposed that it is related to an interaction between troilite and kamacite due to their intergrowth structure. Although perfectly crystalline stoiochiometric troilite is antiferromagnetic [15], it has been proposed that a defect magnetization can be acquired.

To test this hypothesis, we conducted thermal demagnetization and pTRM experiments on samples of troilite from Del Norte County, CA [16], using the

same gas mixtures as for the lunar rocks. We found that the troilite samples have a low-coercivity natural remanent magnetization (NRM) blocked below 50 mT, possibly carried by inclusions of magnetite or pyrrhotite [16], and that they acquire a very high-coercivity pTRM (MDF = 85 mT; Fig. 2) similar to that of mare basalts when cooling in field from temperatures in the range of 300-500°C. Thermal demagnetization of saturation isothermal remanent magnetization (SIRM) shows that the magnetization is blocked below ~500°C. Alternating field (AF) demagnetization before and after heating shows that alteration occurred: the SIRM acquired after heating is ~35 times the initial SIRM, and it is magnetically harder, with MDF increasing from ~25 mT to ~45 mT (Fig. 2). This could be due to breakdown of troilite, with formation of iron (and possibly also ferromagnetic pyrrhotite), which was observed by some authors when heating lunar rocks in controlled atmosphere [17] and in vacuum [18].

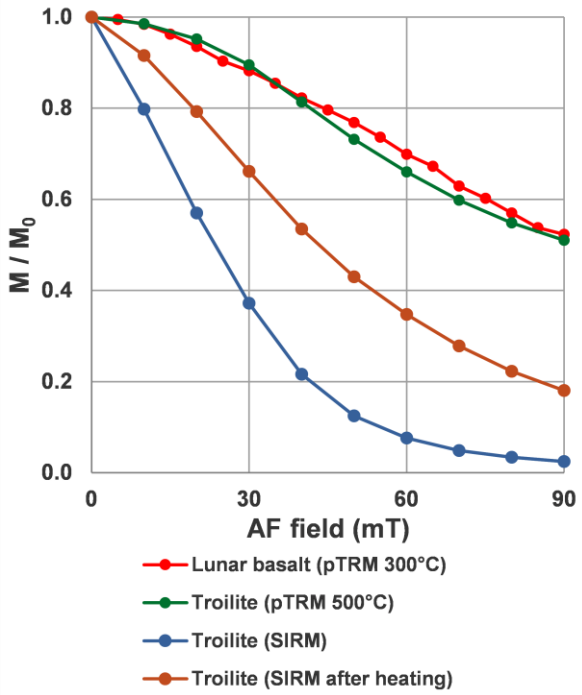

Fig. 2: AF demagnetization of lunar basalt and troilite.

Conclusion: Troilite seems to be unstable in the oxygen fugacity-controlled gas mixtures. It may be possible to prevent alteration of troilite by controlling for both oxygen and sulfur fugacity using mixtures of H_2, CO_2, and SO_2 [19]. Alternatively, mare basalts may not be the most appropriate rocks to obtain reliable lunar paleointensities. Coarse-grained crustal rocks, whose magnetization is carried by kamacite inclusions in silicate crystals and do not contain troilite, may be better candidates as simple oxygen fugacity control may be sufficient to prevent alteration. However, these rocks tend to have older ages than mare basalts, and would only provide paleointensity estimates for the early lunar dynamo.

References: [1] Garrick-Bethell I. et al. (2009) *Science, 323*, 356–359. [2] Shea E. K. et al. (2012) *Science, 335*, 453–457. [3] Cournède C. et al. (2012) *EPSL, 331–332*, 31–42. [4] Suavet C. et al. (2013) *PNAS, 110*, 8453–8458. [5] Weiss B. P. et al. (2010) *Space Sci. Rev, 152*, 341–390. [6] Tauxe L. (2010) *Essentials of Paleomagnetism.* [7] Dunn J. R. and Fuller M. (1972) *The Moon, 4*, 49–62 [8] Sugiura N. et al. (1979) *LPS X*, 2189–2197. [9] Lawrence K. et al. (2008) *PEPI, 168*, 71–87. [10] Pearce G. W. et al. (1976) *LPS VII*, 3271–3297. [11] Nafziger R. et al. (1971) *Research Techniques for High Pressure and High Temperature*, 9–41. [12] Prunier A. R. and Hewitt D. A. (1981) *Geol. Soc. Am. Bull., 92*, 414–416. [13] Sato M. et al. (1973) *LPS IV*, 1061–1079. [14] Grove T. L. and Beaty D. W. (1980) *LPS XI*, 149–177. [15] Pearce C. I. et al. (2006), *Rev. Miner. Geochem., 61*, 127–180. [16] Eakle A. S. (1922) *Am. Mineral., 7*, 77–80. [17] Watson D. E. et al. (1974) *LPS V*, 827–829. [18] Larson E. E. (1978) *LPS IX*, 633. [19] Gaetani G. A. and Grove T. L. (1994) *LPS XXV*, 397.