MASS-DEPENDENT MOLYBDENUM ISOTOPE FRACTIONATION – A NEW TRACER FOR CORE FORMATION. C. Burkhardt[1], R. C. Hin[1], T. Kleine[2], B. Bourdon[3]. [1]Institute of Geochemistry and Petrology ETH Zurich, Clausiusstrasse 25, CH-8092 Zurich, burkhardt@erdw.ethz.ch. [2]Institut für Planetologie, Westfälische Wilhelms-Universität Münster, Wilhelm Klemm Strasse 10, D-48149 Münster. [3]Laboratoire de Géologie de Lyon, ENS Lyon, CNRS, 46 Allée d'Italie, F-69364 Lyon.

Introduction: Constraining the conditions of core formation in asteroids and terrestrial planets is important for understanding planetary accretion and differentiation. In a companion experimental study [1] we identified significant stable Mo isotope fractionations between liquid metal and liquid silicates, rendering Mo isotopes a promising tracer of metal-silicate separation in planetesimals and terrestrial planets. However, the ideal tracer for core formation must not only exhibit significant isotope fractionation between metal and silicates but should also not be fractionated greatly by other processes. Here we explore the extent and origin of stable Mo isotope fractionations in meteoritic and planetary materials, define the stable Mo isotope composition of bulk planetary bodies and demonstrate that Mo isotopes indeed constitute a powerful new tool to constrain the conditions of core formation.

Samples and methods: Several chondrites, iron meteorites, eucrites, angrites, martian meteorites, and lunar samples were selected for this study. Four terrestrial rock standards were also analyzed. All samples were spiked with a ^{100}Mo-^{97}Mo tracer prior to digestion [see 1,2]. Powder aliquots of chondrites that have not been analyzed yet for nucleosynthetic Mo isotope anomalies [2] were digested without tracer and their nucleosynthetic Mo isotope anomalies determined.

Mo was separated from the sample as described in [2]. Mo isotope measurements were performed by MC-ICP-MS at ETH Zurich (*Nu Instruments Plasma 1700*) and at the University of Münster (*Thermo Neptune Plus*). The double spike inversion utilized methods described in [3,4]. Results are reported in $\delta^{98/95}$Mo as the permil deviation of ^{98}Mo/^{95}Mo from the value of the SRM 3134 standard. Data for meteorites were corrected for nucleosynthetic effects using Mo isotope data obtained for unspiked samples.

Results: All chondrites (except CM, CK) and iron meteorites (except group IAB) have indistinguishable $\delta^{98/95}$Mo averaging at –0.16±0.02 (95 % conf.) (Fig. 1). In contrast, most achondrites, lunar and terrestrial samples exhibit heavier Mo isotopic compositions compared to the chondrite-iron meteorite average (Fig. 2). The terrestrial samples average at $\delta^{98/95}$Mo = –0.10 ±0.10 ‰ (95% conf.), while the lunar samples define a mean $\delta^{98/95}$Mo of –0.05±0.03 (95% conf.). The shergottites, angrites and eucrites exhibit more variable $\delta^{98/95}$Mo extending to very heavy isotope compositions.

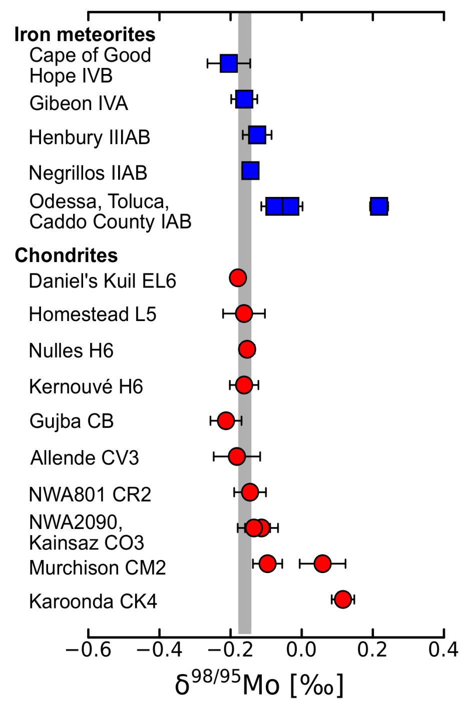

Fig. 1. Mo isotopic data for iron meteorites and chondrites.

Discussion: The new Mo isotope data presented here provide clear evidence for mass-dependent Mo isotope fractionation among meteorites and lunar and terrestrial samples. Most of the investigated silicate samples exhibit heavy Mo isotope compositions compared to chondrites and iron meteorites. Qualitatively this is consistent with experimental results demonstrating that at equilibrium silicates are isotopically heavier than coexisting metals [1]. The Mo isotope fractionation between the silicate samples and chondrites/iron meteorites thus seems to be at least in part a consequence of core formation. However, the variable Mo isotope compositions observed in eucrites, angrites, and shergottites as well as the heavy Mo isotope composition of the IAB irons, CM and CK chondrites indicate that the Mo isotope signatures cannot solely be interpreted as a signature of core formation, but that other processes also resulted in Mo isotope fractionation.

Chondrites and magmatic irons: Most of the investigated chondrites display uniform $\delta^{98/95}$Mo, which is indistinguishable from the value obtained for magmatic iron meteorites (Fig. 1). Since Mo is strongly siderophile most of the Mo will reside in the core and so the Mo isotopic composition of the metallic core is

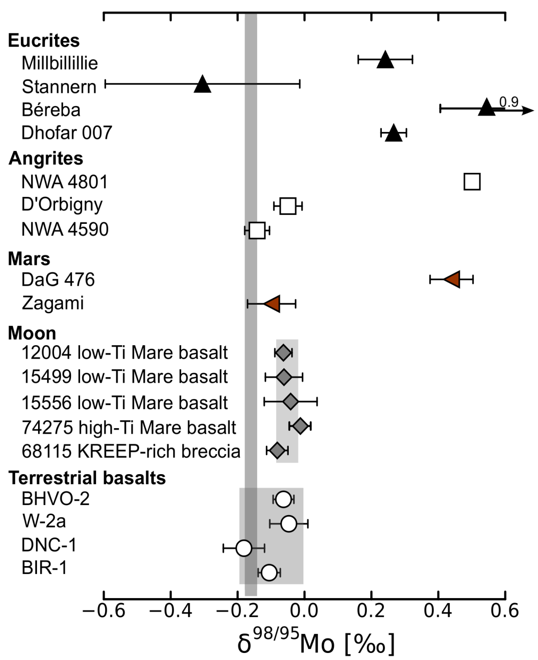

Fig. 2. Mo isotopic compositions of achondrites, lunar and terrestrial samples. Most samples have a heavier Mo isotope composition compared to the chondrite-iron meteorite average, which defines the bulk composition of their parent bodies (grey bar).

indistinguishable from that of the bulk body. The $\delta^{98/95}$Mo of the irons thus reflect those of their parent bodies. The indistinguishable $\delta^{98/95}$Mo of most chondrites and all magmatic irons investigated so far indicates that most planetary bodies in the inner solar system are characterized by a common $\delta^{98/95}$Mo of –0.16 ±0.02. Knowledge of this value is essential for utilizing Mo isotopes to evaluate the magnitude of isotope fractionation induced during core formation (s. below).

CM and CK chondrites: The CM and CK chondrites deviate from the homogeneous $\delta^{98/95}$Mo of the other chondrites. This most likely is due to the redox sensitivity of Mo, which is highly refractory under reducing conditions, but becomes increasingly volatile in oxidizing environments [5]. The CM and CK chondrites formed under highly oxidizing conditions, and so isotopically light Mo might have been preferentially lost as volatile oxides, resulting in the heavy Mo isotope composition observed for these chondrites.

IAB irons: The non-magmatic IAB irons exhibit a heavier and more variable Mo isotopic composition compared to the magmatic irons (Fig. 1). This is consistent with the different formation process of this group of irons and may reflect the loss of isotopically light Mo during impact processes, which were important during the formation of the IAB irons [6].

Moon: All investigated lunar samples exhibit indistinguishable $\delta^{98/95}$Mo, indicating that Mo isotope fractionation during igneous processes on the Moon was minor to absent. The mean $\delta^{98/95}$Mo of the lunar samples, therefore, represents the Mo isotopic composition of the bulk silicate Moon. Using the experimental calibration of [1], the difference between the lunar mantle and the bulk planetary body (as defined by chondrites and magmatic irons) of –0.11±0.04‰ would translate into a metal-silicate equilibration temperature of 1830±370 °C, in good agreement with estimates based on the partitioning of siderophile elements [7].

Earth: Mass-dependent Mo isotope variations at the ~0.1 ‰ level among the investigated terrestrial samples make it difficult to define the $\delta^{98/95}$Mo of the bulk silicate Earth. This will require more work and in particular a detailed understanding of Mo isotope fractionations during igneous processes within the Earth's mantle. Taken at face value the mean of the four terrestrial samples investigated here would correspond to a metal-silicate equilibration temperature of ~2500 °C.

Achondrites: Compared to the lunar and terrestrial samples, the investigated shergottites, angrites and eucrites exhibit more variable $\delta^{98/95}$Mo (Fig. 2). The very high $\delta^{98/95}$Mo of NWA 4801 and DaG 476 probably reflect terrestrial weathering of these desert finds. The $\delta^{98/95}$Mo values of the remaining angrites and shergottites are similar to the values obtained for the lunar and terrestrial samples, suggesting similar temperatures of metal-silicate equilibration in these bodies. However, the $\delta^{98/95}$Mo of both angrites and shergottites is not resolved from chondrites and so the metal-silicate equilibration temperatures cannot reliably be determined. The Mo isotope data for eucrites are difficult to interpret and the origin of the large $\delta^{98/95}$Mo variation remains unclear. For most of the eucrites only very small quantities of Mo were available for analysis. Thus some of the observed scatter may be analytical.

Conclusions: We have shown that most planetary bodies in the inner solar system are characterized by a common $\delta^{98/95}$Mo. Samples derived from the silicate portion of differentiated bodies tend to have slightly heavier Mo isotope compositions, which at least in part must be due to Mo isotope fractionation during core formation. Based on our Mo isotope data a temperature of core formation of ~1800 °C can be determined for the Moon. Although more work is needed to define the $\delta^{98/95}$Mo of the bulk silicate portion of other planetary bodies, the results presented here demonstrate that Mo isotopes are a powerful new tool to constrain core formation in planetary bodies.

References: [1] Hin R. C. et al. (2013) *EPSL, submitted*. [2] Burkhardt C. et al. (2011) *EPSL, 312,* 390-400. [3] Siebert, C. et al. (2001) *G^3, 2,* 2000GC-000124. [4] Rudge J. et al. (2009) C*hem. Geol. 265,* 420-431. [5] Fegley B. and Palme H. (1985) *EPSL 72*, 311-326. [6] Choi B. et al. (1995) *GCA, 59,* 593-612. [7] Righter K. (2002) *Icarus 158*, 1-13.