**OLD VERSUS YOUNG SHERGOTTITES FROM A RE-OS ISOTOPE PERSPECTIVE.** A. D. Brandon[1] ,
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Introduction: A current hotly debated issue is that of what are the igneous crystallization ages of shergottites? The Lu-Hf, Sm-Nd, and Rb-Sr isotope systems give internal isochrons with concordant ages for the shergottites that range from circa 474 Ma to 166 Ma for respective samples ([1,2] and references therin). In contrast, some have argued that the secondary isochrons for individual shergottites using Pb isotopes that give ages circa 4.1 Ga, represent the true igneous crystallization ages and that all other internal isochrons that give younger ages are the result of post igneous crystallization resetting [3,4,5]. Determining which of these scenarios is correct has dramatic implications to understanding shergottite petrogenesis and what it reveals about the evolution of Mars.

To address this issue, the most recent findings from a comprehensive study on the Re-Os isotope systematics of shergottites are presented [6]. The hypothetical constraints derived from diffusion of Re and Os within shergottites are considered to determine if the systematics observed and the young concordant lithophile isotope ages may result from a post igneous crystallization process and in particular, from shock [3,4,5].

Re-Os-Nd Isotope Systematics of Shergottites: Brandon et al. [6] performed a Re-Os isotope study of 23 shergottites. Four fractions from EETA 79001 [6,7] give an isochron age of 164 ± 12 Ma ($\pm2\sigma$, Figure 1) that is concordant to internal mineral isochron ages for ^{87}Rb-^{87}Sr of 173 ± 10, 177 ± 12, and 174 ± 3 Ma, for ^{147}Sm-^{143}Nd of 169 ± 23 Ma, for ^{238}U- ^{206}Pb of 150 ± 15 Ma, and for ^{232}Th-^{208}Pb of 170 ± 36 Ma [8,9,10].
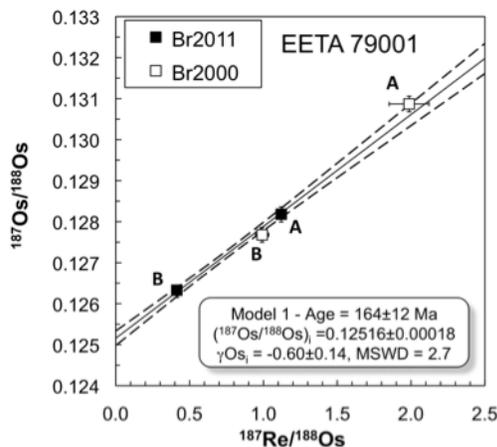
Figure 1. Re-Os isotope systematics of EETA 79001, modified from [6]. A and B designate fractions analyzed from litholigies A and B, respectively. Data from [6] – Br2011, from [7] – Br2000.

Post-crystallization alteration was examined and the data were filtered for such effects [6]. Following this, using the Re/Os ratios for each sample fraction interpreted to be pristine, and the accepted Sm-Nd internal isochron ages for each sample, the initial γOs_i (% deviation from average chondrites at age) was calculated. Figure 2 shows that when γOs_i is combined with εNd_i, a concave upward correlation is apparent from samples with long-term incompatible trace element (ITE) depleted sources, to EETA 79001 and ultramafic samples with intermediate ITE sources, to samples with enriched ITE sources. Similar correla-
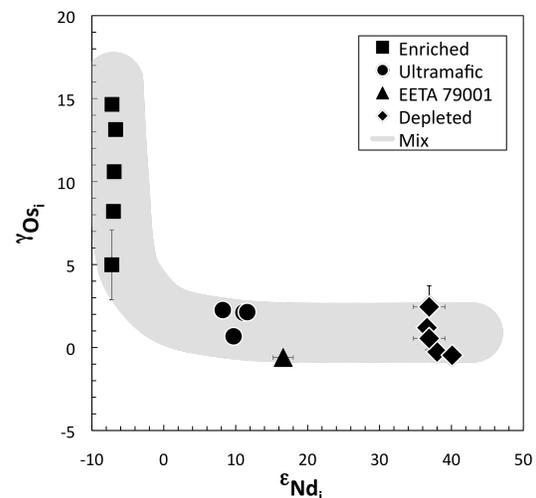

Figure 2. The εNd_i versus $\gamma^{187}Os_i$ for the shergottite data that have been screened for alteration following Brandon et al. [6]. Shaded line represent mixing between end-member sources. Error bars for the symbols are $\pm2\sigma$.

tions exist between the Os isotope data and those for initial Hf and Sr isotopes. Brandon et al. [6] considered these relationships in the context of mixing between ITE-depleted source melts and martian crust, and for mixing between ITE-depleted and ITE-enriched sources or melts. They concluded that crustal contamination cannot likely explain these relationships and that instead, they are best explained by derivation from martian mantle sources that were formed in an early magma ocean and have since remained largely isolated until partial melting to produce shergottite parent magmas. This is consistent with the Hf-Nd isotope relationships for such a model proposed by Debaille et al. [11]. If all of the Nd and Os isotope data are instead recalculated to 4.1 Ga ages, then the well-correlated systematics shown in Figure 2 disappear.

Re-Os Diffusion During Shock: Importantly for the age debate, the Re-Os isotopes track those for Nd, Hf, and Sr isotopes with respect to giving young concordant ages and well-correlated initials for each isotope system. This means that if the young shergottite ages for each reflect post-igneous crystallization, then complete resetting for these lithophile systems and highly siderophile elements occurred while leaving Pb isotopes from a 4.1 Ga magmatic crystallization unchanged. Sulfide grains in shergottites, that are the primary hosts for Re and Os, are disseminated throughout these rocks on grain boundaries and included within silicates and oxides. To reset the Re-Os isotope systematics during post crystallzation, the Re and Os must diffuse through such minerals and fully re-homogenize.

Shock likely heated these rocks to 900-1200°C, below the melting point of most of the minerals but above that which produces maskelynite from plagioclase as observed in most shergottites, which then cooled quickly enough to be quenched to produce glass with undulatory extinction as observed ([12] and references therein). The average length scale each Re and Os ion has to equilibrate and re-distribute to completely reset to the shock age is about 5 mm, which is the typical diameter of each fraction used for the Re-Os isotope study. Lacking diffusion data for Re and Os at this time, it can be assumed that they can diffuse through pyroxene as fast as rare earth elements (trace) and Mg (major) at these temperatures. As compiled in Cherniak and Liang [13] using all reliable data from the literature and their new data, the minimum diffusion coefficient D for these cations in these phases at this range of temperatures is 10^{-20} m^2/s. More typically the D's are 10^{-21} to 10^{-22}. For a 5 mm (0.005 m) length scale using the parabolic rate law where distance x = $(D_{app}t)^{1/2}$ and solving for time (t), for a D_{app} of 10^{-20} the time to diffuse 5 mm is ~80 million years (Figure 3). For a D of 10^{-21}, the time to diffuse 5 mm is ~800 million years. This means using the higher range of D's for diffusing through pyroxene, it takes 10's to 1000's of millions of years to re-homogenize a rock if these temperatures are sustained. Lower D_{app} values that may be more applicable results in even longer equilibration times. This not only applies to the Re-Os isotope system but also would also apply to the Rb-Sr and Sm-Nd isotope systems that give concordant ages in shergottites. So high and sustained temperatures are required to get concordant ages for all of these systems through subsolidus re-equilibration. This is inconsistent with the igneous zonation and undulatory extinction in maskelynite as observed [12,14]. It is thus extremely unlikely that shock produced the young concordant ages in shergottites as they don't have the ob-
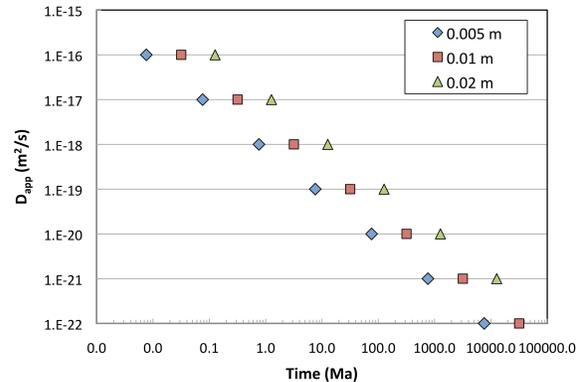
Figure 3. Re-quilibration times versus apparent diffusion rate in millions of years (Ma) for length scales of 0.005, 0.01, and 0.02 meters (m).

served textures of sustained high temperature metamorphism. Experimental diffusion data for Re and Os is necessary to examine this issue further.

Of note, the Re-Os isochron for EETA 79001 (Figure 1) is on two fractions that were sampled ≥ 2 cm from each other in the parent rock (A and B lithologies). It is very unlikely that these fractions equilibrated in a shock diffusion process. The concordancy of this 'whole rock' Re-Os isochron with Rb-Sr and Sm-Nf internal mineral isochrons is inconsistent with diffusion on a smaller length scale as it would have resulted in age decoupling between these.

Hafnium and Cr diffusion in clinopyroxene are also too slow to reset the Lu-Hf isotopes in shergottites during shock [15]. In addition, the Sm-Nd and Rb-Sr isotope systems are less disturbed and the U-Pb and ^{206}Pb-^{207}Pb isotope systems are the most disturbed by shock and annealing [16].

Conclusions: The igneous textures and concordancy observed in shergottites for Re-Os, Rb-Sr, Lu-Hf, and Sm-Nd is inconsistent with shock resetting. In order to fully reset these systems to young ages, sustained periods of high temperatures are required which should produce metamorphic textures.

References: [1] Nyquist et al. 2001, *Rev. Chronol. Evolut. Mars* **96**, 105–164. [2] Shafer et al. 2010, *GCA* **74**, 7307–7328. [3] Bouvier et al. 2005, *EPSL* **240**, 221-233. [4] Bouvier et al. 2008, *EPSL* **266**, 105-124. [5] Bouvier et al. 2009, *EPSL* **280**, 285-295. [6] Brandon et al., 2011. [7] Brandon et al. 2000, *GCA* **64**, 4083-4095. [8] Chen & Wasserburg 1986, *GCA* **50**, 955-968. [9] Nyquist et al. 2001, *LPS* **XXXII**, #1407. [10] Wooden et al., 1982, *LPS* **XIII**, 879-880. [11] Debaille et al., *EPSL* **269**, 186-199. [12] Meyer, Mars Meteorite Compend. [13] Cherniak $ Liang 2007, *GCA* **71**, 1324-1340. [14] Peslier et al. 2010, *GCA* **74**, 4543-4576. [15] Bloch & Ganguly 2011, *Mineral. Mag.* **3**, 536. [16] Gaffney et al. 2011, *MAPS* **46**, 35-52.