The early geological history of the Moon inferred from ancient lunar meteorite Miller Range 13317

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Abstract—Miller Range (MIL) 13317 is a heterogeneous basalt-bearing lunar regolith breccia that provides insights into the early magmatic history of the Moon. MIL 13317 is formed from a mixture of material with clasts having an affinity to Apollo ferroan anorthosites and basaltic volcanic rocks. Noble gas data indicate that MIL 13317 was consolidated into a breccia between 2610 ± 780 Ma and 1570 ± 470 Ma where it experienced a complex near-surface irradiation history for ~835 ± 84 Myr, at an average depth of ~30 cm. The fusion crust has an intermediate composition (Al2O3 15.9 wt%; FeO 12.3 wt%) with an added incompatible trace element (Th 5.4 ppm) chemical component. Taking the fusion crust to be indicative of the bulk sample composition, this implies that MIL 13317 originated from a regolith that is associated with a mare-highland boundary that is KREEP-rich (i.e., K, rare earth elements, and P). A comparison of bulk chemical data from MIL 13317 with remote sensing data from the Lunar Prospector orbiter suggests that MIL 13317 likely originated from the northwest region of Oceanus Procellarum, east of Mare Nubium, or at the eastern edge of Mare Frigoris. All these potential source areas are on the near side of the Moon, indicating a close association with the Procellarum KREEP Terrane. Basalt clasts in MIL 13317 are from a very low-Ti to low-Ti (between 0.14 and 0.32 wt%) source region. The similar mineral fractionation trends of the different basalt clasts in the sample suggest they are comagmatic in origin. Zircon-bearing phases and Ca-phosphate grains in basalt clasts and matrix grains yield 207Pb/206Pb ages between 4344 ± 4 and 4333 ± 5 Ma. These ancient 207Pb/206Pb ages indicate that the meteorite has sampled a range of Pre-Nectarian volcanic rocks that are poorly represented in the Apollo, Luna, and lunar meteorite collections. As such, MIL 13317 adds to the growing evidence that basaltic volcanic activity on the Moon started as early as ~4340 Ma, before the main period of lunar mare basalt volcanism at ~3850 Ma.

INTRODUCTION

Determining the distribution and duration of ancient lunar volcanism is vital to our understanding of the early evolution of the interior and thermal history of the Moon, and in the wider context of volcanism on other planetary bodies. The products of these volcanic eruptions infilled large impact basins, most of which are situated on the lunar near side. The earliest timing of basaltic volcanism is currently unknown, although the majority of the Apollo mare basalts erupted between ~3850 and ~3200 Ma (Head 1976; Head and Wilson 1992; Nyquist and Shih 1992; Stöffler and Ryder 2001; Hiesinger et al. 2003). Ancient lunar volcanism (~3900 Ma) has been documented in the Apollo sample collection by KREEP basalt fragments (Ryder and Spudis 1980; Taylor et al. 1983; Terada et al. 2007) and the ~4300 to 3900 Ma Apollo 14 high-Al mare basalt...
group (Taylor et al. 1983; Nyquist and Shih 1992; Snyder and Taylor 2001; Neal and Kramer 2006). Lunar basaltic meteorites extend the range of these magmatic processes in representing both the youngest (e.g., ~2930 Ma, Northwest Africa [NWA] 032: Borg et al. 2009) and oldest (~4350 Ma, Kalahari 009: Terada et al. 2007; Sokol et al. 2008) known samples. Stratigraphic relationships inferred from remote sensing data and crater counting model age calculations have also indicated occurrences of ancient >3850 Ma volcanism over widespread regions including Mare Frigoris, Mare Australe, Mare Tranquillitatis, and Mare Serenitatis (e.g., Schultz and Spudis 1979, 1983; Hiesinger et al. 2000, 2003, 2008; Whitten and Head 2015).

The age of the oldest known volcanic sample at ~4350 Ma (Kalahari 009: Terada et al. 2007), implies an overlap between the beginning of lunar basaltic volcanism with ancient feldspathic crustal rocks. For example, the ferroan anorthosites (FAN), formed 4560–4280 Ma (Carlson and Lugmair 1988; Nyquist et al. 1995, 2010; Borg et al. 1999, 2011; Nemchin et al. 2009a; Elkins-Tanton et al. 2011; Gross and Joy 2016; Pernet-Fisher and Joy 2016) and ancient magmatic intrusions such as the high-Mg suite (HMS) and the high-Alkali suite formed ~4500 to 4100 Ma, and 4370–4030 Ma (extending to 3800 Ma), respectively (Carlson and Lugmair 1981; Meyer et al. 1989, 1996; Nyquist and Shih 1992; Nyquist et al. 2010; Borg et al. 2011; Carlson et al. 2014).

Lunar meteorites represent invaluable additional samples that provide information about different rock types, originating from regions that are remote from the near side equatorial Apollo and Luna landing sites, including samples from the lunar far side (Warren and Kallemeyn 1991; Korotev et al. 2003; Korotev 2005; Joy and Arai 2013; Calzada-Diaz et al. 2015). In this paper, we report a detailed petrographic, geochemical, and noble gas investigation of lunar meteorite Miller Range (MIL) 13317, coupled with in situ U-Pb analyses of zircon-rich (Zr-rich) and calcium-phosphate (Ca-phosphate) phases. We demonstrate that this meteorite contains various clasts originating from ancient basalts, shedding further light on the earliest phase of the Moon’s magmatic history.

**Miller Range 13317**

Miller Range 13317 is a 32.2 g lunar meteorite that was found in Antarctica in 2013 by the Antarctic Search for Meteorites program (ANSMET). The meteorite was initially classified as an anorthositic breccia (NASA 2015). However, preliminary examinations (Curran et al. 2016; Shaulis et al. 2016; Zeigler and Korotev 2016) indicated that the sample is a mixture of both anorthositic and mafic components, and likely belongs in the category of a “mingled” regolith breccia (e.g., Korotev et al. 2009). MIL 13317 can be further subclassified as moderately mafic (FeO ~9.0 to 16.0 wt%) and Th-rich (5.4 ppm), similar to only seven other lunar stones/meteorite groupings including Lynch 002, the NWA 4472/4485 paired stones, Calcalong Creek, Sayh al Uhaymir (SaU) 169, Dhofar (Dho) 1442, and NWA 6687 (Korotev 2018). Initial studies of Zr-bearing and Ca-phosphate minerals indicated crystallization ages of basalt clasts of 4332 ± 2 Ma (Snape et al. 2018), 4351.8 ± 8.7 Ma, and 4270 ± 24 Ma (Shaulis et al. 2016), which are similar in age to the youngest lunar ferroan anorthosites and ancient magmatic rock suites.

**ANALYTICAL TECHNIQUES**

Analyses of MIL 13317 were performed on a polished thin section (MIL 13317,7), a small polished block (MIL 13317,11a), and bulk chips (from MIL 13317,10 and MIL 13317,11b), all provided by the NASA Meteorite Working Group. The MIL 13317,10 subsplit is from a regolith portion of the sample and MIL 13317,11 is from a fragmental portion. The methodology we employed is outlined below and in more detail in Data S1 and Table S1 in the supporting information.

**Sample Petrography and Chemistry**

Two sections (MIL 13317,7 and MIL 13317,11a) were carbon-coated and analyzed at the University of Manchester. Petrographic characterization was performed using a FEI XL30 field emission gun environmental scanning electron microscope with electron dispersive spectrometer and EDAX Genesis software. Montaged backscattered electron (BSE) images (Fig. 1a) and element maps (Fig. 1b) were processed using the ImageJ software package (Rueden et al. 2017). Optical microscope-cathode luminescence-imaging (CL-imaging) was acquired using a CITL 8200 mk3 “cold” CL system coupled to a transmitted-light microscope; typical conditions were ~10 to 15 kV and a current of ~300 mA.

Mineral phases were analyzed using a Cameca SX 100 electron microprobe (EMPA) equipped with a BSE detector and five wavelength-dispersive spectrometers. Major (Si, Al, Fe, Ca, Mg, Ti) and minor (Na, K, P, Mn, Cr, S, Ni) elements were analyzed with a 15 kV accelerating voltage, a beam current of 20 nA, and a focused beam diameter of 1 μm. Counting times between 10 and 50 s were used for each element.
Typical errors were ~0.1 wt% for major elements and ~0.05 wt% for minor elements for both calibration and sample analyses. The instrument was calibrated against well-characterized mineral standards. Acceptable analytical totals were taken to be between 98 and 101 wt% and mineral compositions were checked for stoichiometry. A defocused beam of 10 \( \mu \text{m} \) was used to assess the bulk composition of fine-grained clasts and homogenous glass areas within the fusion crust.

To further assess the bulk rock composition of MIL 13317, we used glass beads that were produced by fusing MIL 13317,10 (Bulk 1) and MIL 13317,11b (Bulk 2) by laser heating for noble gas analysis (see the Noble Gas Analysis section). Glass beads were mounted on glass slides using superglue adhesive, polished, and carbon-coated. A 10 \( \mu \text{m} \) defocused beam was used to determine major elements by EMPA (Table 1). It is noted that as a result of the high temperatures experienced during melting, there is a potential for the loss of some volatile elements using this approach, although we assume these losses to be negligible as we detected Na in the glass beads made after laser heating (Table 1).

**Mineral Trace Element Chemistry**

Mineral trace element analyses were conducted on an ASI RESOlation 193 nm excimer laser ablation system coupled to an Analytik Jena PlasmaQuant MS Elite ICP-MS at the University of Portsmouth. Data were acquired for isotopes of \( ^{27}\text{Al}, ^{29}\text{Si}, ^{43}\text{Ca}, ^{31}\text{P}, ^{39}\text{K}, ^{45}\text{Sc}, ^{49}\text{Ti}, ^{51}\text{V}, ^{52}\text{Cr}, ^{55}\text{Mn}, ^{59}\text{Co}, ^{66}\text{Ni}, ^{87}\text{Rb}, ^{87}\text{Sr}, ^{89}\text{Y}, ^{90}\text{Zr}, ^{93}\text{Nb}, ^{95}\text{Mo}, ^{133}\text{Cs}, ^{137}\text{Ba}, ^{139}\text{La}, ^{140}\text{Ce}, ^{141}\text{Pr}, ^{146}\text{Nd}, ^{147}\text{Sm}, ^{153}\text{Eu}, ^{157}\text{Gd}, ^{163}\text{ Dy}, ^{165}\text{Ho}, ^{166}\text{Er}, ^{172}\text{Yb}, ^{175}\text{Lu}, ^{178}\text{Hf}, ^{181}\text{Ta}, ^{208}\text{Pb}, ^{232}\text{Th}, \) and \( ^{238}\text{U} \). Samples were ablated using a spot size of 60 \( \mu \text{m} \), at a repetition rate of 8 Hz with the laser energy at the target (fluence) regulated at \( \sim 5.5 \text{ J cm}^{-2} \). Laser ablation times of 60 s were used with 10 s dwell times between analyses. SRM NIST 612 was used as the calibration standard. Calcium and Si were used as the internal normalizing standards for plagioclase and other phases, respectively. Repeat NIST 610 and NIST 614 analyses were measured over the analytical campaign in order to assess the data quality (supporting information Table S2). Typical \%RSD for most elements in NIST 610 is better than 5\% (except for Ni—19\%, Ti—8\%, Cr—6\%, Sr—10\%). Typical \%RSD for elements in NIST 614 is better than 10\% (except for Sc—38\%, Cr—27\%, Mn—31\%, Co—23\%, Zr—24\%, Mo—14\%). Detection limits varied by <10 ppb. The abundances obtained for NIST 610 and 614 are within range of certified values reported within the GeoReM database (georem.mpch-mainz.gwdg.de). Data reduction was conducted using Iolite software (v3) using the “Trace Elements IS” data reduction scheme, where the time-resolved display and segment-picking for data integrations allow signal spikes from inclusions to be identified and avoided.

**Noble Gas Analysis**

Two rock chips, from subsplits MIL 13317,10 and MIL 13317,11b, were investigated for their bulk rock noble gas content. Neon, Ar, and Xe were analyzed on a
Table 1. Fusion crust chemistry data for MIL 13317,7, bulk rock chemistry data of MIL 13317,10 and MIL 13317,11b (after laser heating during noble gas analysis), and bulk clast composition data for fine-grained impact melt breccia (imb) clasts in MIL 13317,7. Errors are given as 2σ SD values, where the value represents the average of a sum of analysis (number of points analyzed: fusion crust on MIL 13317,7 = 28, MIL 13317,10 = 27, and MIL 13317,11b = 25).

| Number of analysis | SiO2 | TiO2 | Al2O3 | Cr2O3 | FeO | MnO | MgO | CaO | Na2O | K2O | P2O5 | NiO | SO3 | Total ppm | Ba | La |
|--------------------|------|------|-------|-------|-----|-----|-----|-----|------|-----|------|-----|-----|------------|----|----|
| 13317,7 Fusion     | 28   | 49.13| 1.04  | 15.98 | 0.13| 12.31| b.d | 5.87| 11.60 | 0.31| 0.27 | b.d | b.d | 96.65      | 272.32 | 25.04 |
|                    | ±0.76| ±0.11| ±0.90 | ±0.01 | ±0.98| ±0.27| ±0.32| ±0.02| ±0.04|     |     |     |     |   |            | ±48.51 | ±5.13 |
| 13317,10 Bulk 1    | 27   | 48.95| 0.71  | 19.01 | 0.20| 9.66 | b.d | 6.37| 12.05 | 0.26| 0.13 | 0.10| 0.02| 0.05  | 97.51 |
|                    | ±3.44| ±0.50| ±4.50 | ±0.12| ±2.68| ±3.75| ±3.75| ±0.14| ±0.06| ±0.14| ±0.05| ±0.08|    |   |            |    |    |
| 13317,11b Bulk 2   | 25   | 49.90| 1.22  | 9.83  | 0.34| 16.85| 0.25| 5.35| 10.70 | 0.52| 0.34 | 0.28| b.d | b.d | 95.57 |
|                    | ±2.63| ±0.76| ±2.24 | ±0.81| 3.59 | ±0.09| 4.95 | ±3.07| ±0.42| ±0.18| ±0.19|     |    |   |            |    |    |
| Clast Type         |      |      |       |       |     |     |     |     |      |     |     |     |     |   |   |            |    |    |
| Clast 9a imb       | 30   | 48.48| 0.43  | 23.65 | 0.14| 6.60 | 0.10| 4.14| 14.21 | 0.28| 0.24 | b.d | b.d | 0.15 | 98.42 |
|                    | ±4.58| ±0.77| ±7.20 | ±0.13| ±6.16| ±0.09| ±4.29| ±3.23| ±0.09| ±0.37| ±0.83|    |    |   |            | ±48.51 | ±5.13 |
| Clast 9b imb       | 10   | 46.31| 0.10  | 27.84 | 0.10| 4.27 | 0.06| 3.20| 16.51 | 0.29| 0.08 | b.d | b.d | 0.18 | 98.93 |
|                    | ±5.65| ±0.34| ±22.78| ±0.37| ±15.75| ±0.23| ±11.49| ±8.70| ±0.25| ±0.12| ±1.12|    |    |   |            |    |    |
| Clast 19 imb       | 23   | 49.78| 0.51  | 18.44 | 0.20| 9.69 | 0.14| 5.93| 12.96 | 0.27| 0.14 | b.d | b.d | 0.10 | 98.17 |
|                    | ±6.56| ±0.68| ±19.50| ±0.24| ±13.62| ±0.20| ±9.74 | ±8.71| ±0.24| ±0.21| ±0.25|    |    |   |            |    |    |
| Clast 20 imb       | 15   | 46.45| 0.36  | 25.43 | 0.12| 5.58 | 0.08| 3.91| 15.37 | 0.27| 0.09 | b.d | b.d | 0.12 | 97.80 |
|                    | ±1.73| ±0.25| ±4.87 | ±0.08| ±3.51 | ±0.05| ±2.54 | ±2.15| ±0.06| ±0.04| ±0.24|    |    |   |            | ±48.51 | ±5.13 |

Detection limits for EMPA:

|                  |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |
|------------------|------|------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
|                  | 0.02 | 0.04 | 0.01  | 0.01  | 0.04  | 0.02  | 0.008 | 0.03  | 0.05  | 0.01  | 0.02  | 0.018 | 0.018 | 0.04  |       |       |

aNumber of analysis refers to the number of spot analysis during EMPA for a specific clast.

bData from Zeigler and Korotev (2016).

cIf the average oxide (wt%) composition fell below the detection limits shown here they were determined to be below detection (b.d.).
Thermo Scientific™ HELIX-MC noble gas mass spectrometer at the University of Manchester. A Photonmachines™ fusion diode laser (3 mm beam diameter) was used to step-heat samples from 0.6 to 21 W (in six to nine steps depending on mass of the sample) corresponding to the complete melting of the sample. Prior to analysis, chips were loaded into an aluminum holder in the laser port and baked for 24 h at 200 °C to remove any absorbed atmospheric gases. During analysis, any active gases released from the sample were removed by two SAES getters (NP10 and GP50) during the gas preparation stage (see Section S1.2 in Data S1 for full experimental technique). Blanks and air calibrations were performed before and after every sample analysis. Typical blanks were 6.1 × 10^{-12} \text{cm}^3 \text{STP} 20Ne, 8.6 × 10^{-11} \text{cm}^3 \text{STP} 40Ar, and 1.2 × 10^{-14} \text{cm}^3 \text{STP} 132Xe. All isotopes were background and blank corrected. Neon-22 was corrected for interferences from CO$_2$ and blank corrected. Neon-22 was corrected for contamination with terrestrial common Pb corrected.

Neon-isotope systematics reflect a mixture of cosmogenic (c), trapped solar wind (tr), and trapped fractionated solar wind (fSW) with the following endmember isotope ratios: $(^{20}\text{Ne}/^{22}\text{Ne})_{c} = 0.8$, $(^{20}\text{Ne}/^{22}\text{Ne})_{tr} = 13.8$, $(^{20}\text{Ne}/^{22}\text{Ne})_{fSW} = 11.2$, $(^{21}\text{Ne}/^{22}\text{Ne})_{c} = 0.76$, $(^{21}\text{Ne}/^{22}\text{Ne})_{tr} = 0.033$, and $(^{21}\text{Ne}/^{22}\text{Ne})_{fSW} = 0.0295$ (Wieler 2002a, 2002b; Lorenzetti et al. 2005). For argon, $(^{38}\text{Ar}_{c}$ and $(^{36}\text{Ar}_{tr}$ were derived using the endmember cosmogenic $(^{36}\text{Ar}/^{38}\text{Ar})_{c} = 0.65$ and trapped $(^{36}\text{Ar}/^{38}\text{Ar})_{tr} = 5.35$ compositions (Wieler 2002a, 2002b). Xenon isotopic measurements were first treated as the sums of contributions from cosmic ray spallation, fission, and trapped solar wind components (see Section S1.2 in Data S1). The following assumptions were made: $(^{132}\text{Xe}/^{126}\text{Xe})_{REE} = 0.06$, $(^{128}\text{Xe}/^{126}\text{Xe})_{REE} = 1.22$, $(^{132}\text{Xe}/^{126}\text{Xe})_{REE} = 1.053$, $(^{28}\text{Xe}/^{126}\text{Xe})_{REE} = 1.621$; for the trapped components, we assumed: $(^{132}\text{Xe}/^{136}\text{Xe})_{REE} = 3.33$, $(^{134}\text{Xe}/^{136}\text{Xe})_{REE} = 1.22$; and for fission components, we assumed: $(^{132}\text{Xe}/^{136}\text{Xe})_{REE} = 0.577$, $(^{134}\text{Xe}/^{136}\text{Xe})_{REE} = 0.828$, and $(^{132}\text{Xe}/^{136}\text{Xe})_{REE} = 0.885$, $(^{134}\text{Xe}/^{136}\text{Xe})_{REE} = 0.939$ (Pepin et al. 1995; Swindle 2002; Crowther and Gilmour 2013; Meshik et al. 2014). The method of Hohenberg et al. (1978) was used to calculate production rates to determine $t_{21} (^{21}\text{Ne})$, $t_{38} (^{38}\text{Ar})$, and $t_{126} (^{126}\text{Xe})$ exposure ages ($t = \text{cosmogenic isotope abundance/production rate}$). The antiquity of the sample based on $t_{21} (^{21}\text{Ne})$, $t_{38} (^{38}\text{Ar})$, and $t_{126} (^{126}\text{Xe})$ exposure ages ($t = \text{cosmogenic isotope abundance/production rate}$) was calculated using the method of Joy et al. (2011a).

**Secondary Ion Mass Spectrometry**

The Ca-phosphate (apatite and merrillite), zircon, and baddeleyite U-Pb systems were analyzed using a Cameca IMS 1280 ion microprobe at the NordSIMS facility at the Swedish Museum of Natural History, Stockholm, using a methodology similar to that outlined in previous studies (Whitehouse et al. 1997, 2005; Nemchin et al. 2009b; Snape et al. 2016). Full details of the secondary ion mass spectrometry (SIMS) methodology are included in Section S1.3 in Data S1.

These data were processed using in-house SIMS data reduction spreadsheets and the Excel add-in Isoplot (version 4.15; Ludwig 2012). Three different sets of ages were calculated (1) Pb-Pb isochron ages, using the Pb isotope compositions uncorrected for contamination with terrestrial common Pb, (2) terrestrial common Pb corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ages (using the model of Stacey and Kramers [1975] and their values for present-day terrestrial Pb isotopic ratios), and (3) U-Pb concordia ages. A U-Pb concordia age was not calculated for the baddeleyite and merrillite grains due to the lack of analyses on appropriate matrix-matched external standards and the well-known matrix-orientation problems associated with U-Pb calibration of baddeleyite (Wingate and Compston 2000). All age uncertainties are reported at the 95% confidence limit in the following discussion. Following SIMS analysis, SEM images were acquired of the targeted phases to assess the exact locations of the SIMS pits, to ascertain if any overlapped with mineral grain boundaries (see supporting information Fig. S1). Four such analyses were identified (supporting information Table S3) and excluded from the combined age calculations (e.g., weighted averages of $^{207}\text{Pb}/^{206}\text{Pb}$ ages or Pb-Pb isochrons), and are not considered further.

**RESULTS**

The petrology and geochemistry of MIL 13317 are consistent with a lunar origin. The Fe/Mn ratios of pyroxene and olivine in MIL 13317 plot on previously determined lunar meteorite trend lines (Fig. 2) (Papike 1998; Karner et al. 2003, 2006; Joy et al. 2014). MIL 13317 also contains trapped “solar” noble gases with similar compositions to other lunar meteorites (see the Regolith History section).

**Petrography and Mineral Chemistry of MIL 13317**

Miller Range 13317 is a well-consolidated regolith breccia, comprising a diverse array of polymictic lithic clasts (black, dark-gray, clear, and cream) fused together in a heterogeneous matrix of the same material, with shocked melt veins and mineral fragments (supporting information Figs. S2 and S3). Lithic clasts are up to ~7 mm in size and include regolith breccias, basalts, feldspathic fragments, symplectites, norites, and granulites.
(Figs. 1 and 3). Mineral fragments are up to 1.5 mm in diameter and typically consist of pyroxene, plagioclase, and olivine. Impact melt breccias are abundant in the sample, including clast-rich and crystalline types. Melt veins are variable in size and crosscut both the matrix and clasts in random orientations (Fig. 3a). Rare glass spherules (<100 μm) were found in the sample matrix (see Fig. 3i inset), but no agglutinates were identified. Various clast types are described below (see Fig. 1c for clast location and number). Mineral and bulk clast compositions are given in supporting information Tables S4–S8.

**Basaltic Clasts**

There are numerous medium- and fine-grained basaltic clasts in MIL 13317 (Fig. 1). The largest clasts (Clasts 1 and 4, Figs. 3a and 3b) are ~2.0 × 1.5 mm in size, and are composed of an intergrowth of pyroxene, plagioclase, silica, and areas of mesostasis (e.g., K-glass, phosphates, zircons) associated with late-stage crystallized residual melt material of basaltic magmas. Other basaltic clasts range from fine-grained (minerals <0.2 mm) clasts with “pods” of late-stage mesostasis, to coarse-grained (minerals <1 mm) clasts and fragments with both intergranular and subophitic textures. These textures suggest a range of cooling histories for basaltic clasts in MIL 13317.

Pyroxene in basaltic clasts of MIL 13317 are generally anhedral grains <1 mm in size and show a wide range in Fe/Mg ratios. Complex chemical zoning from Mg-rich cores to Fe-rich rims is present in all pyroxene grains, with augite being the modally most abundant species with lesser amounts of pigeonite (Fig. 4). Chemical compositions of pyroxene are Wo8-37 En0.4-54 Fs28-88. Some more evolved basaltic clasts (e.g., Clast 15) are Fe-rich, with compositions of Wo17-27 En6-10 Fs66-75. Pyroxene grains have both LREE depleted and flat LREE profiles ([La/Sm]CI = 0.17–0.96), display flat HREE profiles ([Dy/Yb]CI = 0.91–1.06) (Fig. 5a), and have strong negative Eu-anomalies.
Pyroxene U and Th abundances range from 0.02 to 1.47 ppm and 0.02 to 8.80 ppm, respectively. Elongated silica grains (SiO$_2$ = 90.0–99.8 wt%) and plagioclase laths are up to 1 mm in size. Plagioclase show a range of An compositions (An$_{78-94}$) across the suite of mineral fragments (Fig. 6a). Plagioclase REE abundances display fractionated LREE relative to the HREE ([La/Yb]$_{CI}$ = 11.9–12.2) and positive Eu-anomalies (Eu/Eu* = 19.8–22.7). Abundances of Sr and Ba in plagioclase range from 222 to 223 ppm.
Fig. 4. Pyroxene mineral composition data for clasts and matrix in MIL 13317.7 and MIL 13317.11a. a) Pyroxene quadrilateral of these data. b) Fe# (atomic Fe/Fe + Mg × 100) versus Ti# (atomic Ti/Ti + Cr × 100). Fields show the extent of pyroxene composition in Apollo VLT, low-Ti and high-Ti mare basalts (from Bence and Papike 1972; Vaniman and Papike 1977; Dymek et al. 1975). c) Fe# (atomic Fe/Fe + Mg × 100) versus Al/Ti (atoms per formula unit). (Color figure can be viewed at tonylinelibrary.com.)
and 72 to 73 ppm, respectively. Mineral trace element analyses of plagioclase and pyroxene grains from basalt Clasts 1 and 4 display REE systematics similar to those reported in minerals from Apollo low-Ti mare basalts (Table S2). Plagioclase CL images (Fig. S3) display a deep red-purple color, indicative of maskelynite (a high-shock plagioclase pseudomorph) (Kayama et al. 2009; Pernet-Fisher et al. 2017). The bulk trace element compositions for fine-grained basalt Clast 18 were estimated by averaging together six 60 µm laser ablation spots, which included a representative range of minerals in the clast. The average compositions display LREE enrichments ([La/Sm]CI = 1.46), fractionated HREE profiles ([Dy/Yb]CI = 1.18), and negative Eu-anomalies (Eu/Eu* = 0.21) (Fig. 5e).

Large regions of symplectites and mesostasis (up to 0.7 × 0.4 mm in Clasts 1 and 4) are present, typical of Apollo mare basalts (e.g., Pernet-Fisher et al. 2014;
The mesostasis areas within these basaltic clasts consist of predominantly anhedral Fe-rich pyroxene (Wo11-29 En0.6-10 Fs61-88), fayalite (Fa96-97), Si- and K-rich glass (3–5 wt% K₂O and on average 74 wt% SiO₂), ilmenite, and silica. Rare mineral phases (<100 μm) include irregular grains of troilite, apatite, merrillite, tranquillityite, baddeleyite, and zircon.

Highland and Impact Melt Breccias Clasts

Miller Range 13317 contains a range of nonbasaltic rocks and minerals showing an affinity to lunar highland rock types. Several clasts have mafic minerals (pyroxene and olivine) with high-Mg numbers (Mg#>70) and calcic plagioclase (Fig. 7a). A 0.6 × 1.2 mm sized noritic clast (Clast 5, Fig. 3e) with an igneous texture contains highly anorthositic plagioclase (An95-97) and subhedral Mg-rich orthopyroxene (Mg#71-80, Wo5-9 En65-78 Fs19-27). The plagioclase grains in Clast 5 display large positive Eu-anomalies (Eu/Eu* = 20.08 and 52.98) and LREE enrichments ([La/Sm]Cl₁ = 1.85 and 5.67) (Fig. 5c). Metagabbroic granulites (Clasts 12 and 14, Fig. 3g) and matrix fragments include feldspathic varieties (Clast 12; An>94, Fig. 7a) and pyroxene and/or olivine with Mg# between 63 and 87. Trace element systematics for plagioclase from granulitic Clast 12 (Eu/Eu* = 15.83; [La/Sm]Cl₁ = 1.77) are similar to that of plagioclase from igneous Clast 5 (Figs. 5c and 5d).
Fig. 7. a) Anorthite component (mol%) in plagioclase versus Mg# (=Mg/[Mg + Fe] × 100) in mafic minerals in MIL 13317. Clast compositional relationships where the data points show the average An content in plagioclase (An#) and the Mg# content in pyroxene per clast. The error bars represent the total range of mineral chemistries determined per clast. Rock suite fields are from Shearer et al. (2006). b) Comparison of FeO (wt%) versus Al₂O₃ (wt%) showing the fusion crust composition for MIL 13317. MIL 13317 is compared to other Miller Range and the YAMM group of basaltic meteorites (data from Korotev and Zeigler 2014). MIL 13317 is also compared to another Th-rich meteorite Calclalong Creek (Hill and Boynton 2003) and Th-poor meteorite Kalahari 009. (Color figure can be viewed at wileyonlinelibrary.com.)
There are a variety of texturally diverse impact melt breccia clasts in MIL 13317 (Figs. 3c–g). The largest is a 2.0 x 7.0 mm feldspathic fragment (Clast 9, Fig. 7b) with a noritic anorthosite composition composed of calcic plagioclase (An$_{94-98}$), low-Ca pyroxene (Mg$\#$ 55-70, Wo$_{4.4}$–En$_{69.6}$–Fs$_{28.4}$), and minor amounts of troilite. The FAN-like plagioclase (Fig. 7a) and trace element systematics for Clast 9 were estimated by averaging data from six 60 $\mu$m laser ablation spots. Clast 9 displays fractionated LREE/HREE profiles ([La/Yb]$_{CI}$ = 2.2) and negative Eu-anomalies (Eu/Eu* = 0.61) (Fig. 5e). Concentrations of Sm are lower than Apollo mafic impact melt breccia groups (Fig. 5f) (Korotev et al. 2009). They are, however, similar to some lunar meteorites which are more mafic and less feldspathic (i.e., anorthositic norite and troctolite) than feldspathic lunar meteorites in general (Korotev et al. 2009; Korotev 2018). Hypocrystalline impact melt clasts are also abundant, and comprise interstitial plagioclase crystals surrounded by mafic glassy melt.

Other impact melt breccia clasts in MIL 13317 include crystalline types with igneous-like textures (bas/IMB in Figs. 3c and 3d). These types (Clasts 2, 7, and 8) consist of zoned pyroxene (Wo$_{4.4}$–En$_{19.28}$–Fs$_{59.68}$), plagioclase (An$_{64.90}$), silica, globular Fe-metal, and troilite. Plagioclase grains are richer in K (Or$_{9.9-28}$) than those seen in the basalt and highland lithologies in MIL 13317 (Fig. 6a). Pyroxene displays fine exsolution lamellae (<1 $\mu$m) with pigeonite hosting augite. Pyroxene compositions lie on a mixing line between fersilite (average Wo$_{10}$ En$_{27}$ Fs$_{66}$) and Ca-rich augite (average Wo$_{36}$ En$_{50}$ Fs$_{43}$). Accessory phases include Ca-phosphates (both apatite and merrillite, <50 $\mu$m) and zircon grains (<20 $\mu$m).

Rare glass spherules (50–100 $\mu$m) are sporadically distributed through MIL 13317. One glass spherule (Fig 3i) was determined to have an impact origin (Mg/Al <1.5 and Mg/Ca <1), rather than volcanically sourced (Mg/Al >1.5 and Mg/Ca >1), as indicated by its Mg/Al = 0.34 and Mg/Ca = 0.40 ratios (Delano and Livi 1981; Joy et al. 2011a).

**Mineral Fragments**

The matrix of MIL 13317 consists of single mineral fragments up to 1.5 mm in size. The majority of mineral fragments are plagioclase and pyroxene with a wide range of compositions (An$_{78.99}$ and Wo$_{4.44}$ En$_{19.70}$ Fs$_{51.88}$, respectively, Figs. 4 and 6). Trace element compositions of the matrix pyroxene are similar to pyroxene within the basaltic clasts (Fig. 5b). The REE patterns display negative Eu-anomalies (Eu/Eu* = 0.01–0.05) and variable LREE depletions of [La/Sm]$_{CI}$ = 0.09–0.17 (Fig. 5b). Silica and ilmenite are also present with minor amounts of Mg-rich olivine up to 1.5 mm in size. Accessory minerals in the matrix are mainly Ca-phosphate (merrillite and apatite) and Zr-bearing phases <100 $\mu$m in size. Three Zr-bearing phases were identified: zircon, baddeleyite, and the rare lunar mineral tranquillityite (see Fig. S1). These accessory phases are typically associated with symplectic areas with pyroxene, minor olivine, K-rich glass, and K-feldspar.

**Evolved Lithologies**

Highly evolved lithologies of granitic (granophyric) K-feldspar and silica intergrowths, up to 0.7 mm are present in MIL 13317 (Fig. 1—cyan colored phases). The K-feldspar occurs as blocky phases with lath-like silica. These assemblages are commonly associated with regions of mesostasis and areas of K-rich glass, and are most similar to the Apollo high-alkali suite rock types. It is possible that some of these are fragments of mesostasis from the basaltic clasts.

**Fusion Crust and Bulk Rock Analysis**

A dark-brown vesicular fusion crust is present on the MIL 13317,7 section, which has an average composition shown in Table 1 (Fig. S2). The fusion crust displays LREE enrichment ([La/Sm]$_{CI}$ = 1.40), fractionated HREE profile ([Dy/Yb]$_{CI}$ = 1.15), and a negative Eu-anomaly (Fig. 5e). The Th-composition of the fusion crust is 5.41 ppm (Table S2).

Two bulk compositions we determined from chips MIL 13317,10 (Bulk 1) and MIL 13317,11b (Bulk 2). Bulk 1 composition (MIL 13317,10, Table 1) is slightly more ferroan than the fusion crust and most similar to the Apollo 17 regolith breccias in terms of Mg$\#$ (29–47) and K$_2$O (0.06–0.15 wt%) contents. Bulk 2 composition (MIL 13317,11b) has a higher K$_2$O content and is richer in Fe (Table 1; Fig 7b) compared to the fusion crust and bulk 1. The fusion crust composition is intermediate between bulk 1 and 2 (Table 1; Fig. 7b).

**Noble Gas Inventory**

Five separate chips of MIL 13317 (three from MIL 13317,10 and two from MIL 13317,11b) were analyzed for Ne and Ar isotopes and one chip (from MIL 13317,10) for Xe isotopes. A summary of the noble gas data is given in Table 2 and Tables S9–S11 in the supporting information.

Neon data for MIL 13317,10 are dominated by trapped $^{20}$Ne/$^{22}$Ne and $^{21}$Ne/$^{22}$Ne ratios (Fig. 8a), whereas MIL 13317,11b shows a trend of increasing $^{21}$Ne/$^{22}$Ne and decreasing $^{20}$Ne/$^{22}$Ne ratios with increasing laser heating steps (see Table S9). Trapped $^{20}$Ne/$^{22}$Ne ratios ($^{20}$Ne$^{22}$Ne)$_{tr}$ were determined using a least squares fitting to the data on a three-isotope mixing diagram (Fig. 8a). Overall, Ne in MIL 13317
has a $^{20}\text{Ne}/^{22}\text{Ne}$ ratio between 10.8 and 12.6 (for both subsplits, Table 2). The concentrations of $^{20}\text{Ne}_{\text{tr}}$ in MIL 13317,10 are higher than in MIL 13317,11b (Table 2; Fig. 8a). The $^{36}\text{Ar}_{\text{tr}}$ concentrations were also ~10 times higher in MIL 13317,10 compared to MIL 13317,11b (Table 2). The $^{40}\text{Ar}/^{36}\text{Ar}_{\text{tr}}$ values of 2.03–4.81 indicate an antiquity age of between 2610 ± 783 Ma and 1570 ± 471 Ma (using the method of Eugster et al. 2001; Joy et al. 2011b) (Table 2).

The concentration of cosmogenic $^{38}\text{Ar}_{\text{c}}$ is an order of magnitude higher in MIL 13317,10 compared to MIL 13317,11b (Table 2). There is also a factor of approximately 2 difference in $^{38}\text{Ar}_{\text{c}}$ within the three subsplits of MIL 13317,10. In contrast, $^{21}\text{Ne}_{\text{c}}$ concentrations for all analyzed chips are similar and mostly within error of each other (Table 2).

Three-isotope mixing diagrams of $^{124}\text{Xe}/^{130}\text{Xe}$ versus $^{126}\text{Xe}/^{130}\text{Xe}$ (see Fig. 8b) illustrate the components present in MIL 13317,10. The sample is dominated by a solar wind component (98%), but contains minor amounts of spallation products from Ba: the bulk spallation Xe signature corresponds to (30 ± 11%) (1σ) of $^{126}\text{Xe}$ derived from Ba. Heavy Xe isotopes formed by fission of U or Pu are not evident in any of the samples (supporting information Fig. S4). Petrographic observations revealed that MIL 13317 contains REE-bearing phases like zircon, apatite, merrillite, and K-rich glass, which account for the presence of some spallogenic xenon components. The depth-sensitive shielding indicator $^{131}\text{Xe}/^{126}\text{Xe}_{\text{c}}$ is 5.6 ± 0.4 (1σ), assuming an average lunar regolith density of 1.5 g cm$^{-3}$ (Carrier et al. 1991), this corresponds to an average burial depth on the Moon of about 30 cm. This depth was used to calculate production rates of $P_{21} = 0.1789 \times 10^{-8}$ cm$^3$/STP/Ma, $P_{38} = 0.1369 \times 10^{-8}$ cm$^3$/STP/Ma, and $P_{126} = 0.2250 \times 10^{-8}$ cm$^3$/STP/Ma (based on the method of Hohenberg et al. 1978). These production rates yield exposure ages for neon ($t_{21}$) of between 71 ± 28 and 320 ± 117 Myr, argon ($t_{38}$) of between 457 ± 1.4 and 562 ± 159 Myr, and xenon ($t_{126}$) of 835 ± 84 Myr (Table 2).

### U-Pb and Pb-Pb Chronology

Analyses of a zircon and five baddeleyite grains in MIL 13317,7 yield terrestrial common Pb corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 4326 ± 12 Ma and 4349 ± 16 Ma (Table 3). The zircon analyses were carried out on a single ~70 μm grain. Calculating either a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age (4337 ± 8 Ma; MSWD = 2.9; $P = 0.034$; Fig. 9), or a Pb-Pb isochron age (4331 ± 49 Ma; MSWD = 4.1; $P = 0.016$; Fig. 10c) from these data results in high MSWD values and low probabilities of fit, potentially due to partial resetting of the U-Pb system within different parts of the grain. Nonetheless, a U-Pb concordia age of 4334 ± 17 Ma is determined for the three zircon analyses with less than 10% discordance (concordance and equivalence MSWD = 1.4; $P = 0.24$; Fig. 11a). Despite being located in separate areas within the breccia matrix, the five baddeleyite grains analyzed were all associated with similar symplectite assemblages (Fig. S1) and all yield indistinguishable $^{207}\text{Pb}/^{206}\text{Pb}$ ages of between 4341 ± 10 Ma and 4345 ± 8 Ma. Considering the
baddeleyite as a single population gives weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of $4344 \pm 4\text{ Ma}$ (MSWD = 0.11; $P = 0.98$; Fig. 9) and a Pb-Pb isochron age of $4344 \pm 5\text{ Ma}$ (MSWD = 0.12; $P = 0.95$; Fig. 10d).

Two of the 12 matrix apatite analyses were obtained from a single ~40 $\mu$m apatite grain, within which it was possible to place two SIMS spots. These two SIMS spots yielded the lowest $^{207}\text{Pb}/^{206}\text{Pb}$ ratios in the sample, with a
weighted average $^{207}\text{Pb} / ^{206}\text{Pb}$ age of $3889 \pm 16$ Ma (MSWD = 0.06; $P = 0.8$) and a U-Pb concordia age of $3890 \pm 33$ Ma (concordance and equivalence MSWD = 1.3; $P = 0.29$; Fig. 11b). The remaining 10 matrix apatite analyses have similar $^{207}\text{Pb} / ^{206}\text{Pb}$ ratios to the zircon and baddeleyite grains, equating to $^{207}\text{Pb} / ^{206}\text{Pb}$ ages between $4333 \pm 14$ Ma and $4349 \pm 14$ Ma (Table 3). It was also noted that one of these analyses was an apatite grain within a large symplectite assemblage (Clast 16, Fig. 3h), similar to the smaller symplectite assemblages containing the baddeleyite grains. If these 10 matrix apatite analyses are considered to form a single population, a weighted average $^{207}\text{Pb} / ^{206}\text{Pb}$ age of $4342 \pm 4$ Ma (MSWD = 0.59; $P = 0.81$; Fig. 9) is obtained. A U-Pb concordia age of $4342 \pm 13$ Ma (concordance and equivalence MSWD = 1.2; $P = 0.28$; Fig. 11c) is obtained for the six analyses that are less than 10% discordant; however, the $^{204}\text{Pb} / ^{206}\text{Pb}$ and $^{207}\text{Pb} / ^{206}\text{Pb}$ ratios of these matrix apatite analyses do not provide sufficient spread to constitute a 10-point isochron. The similar ages of the matrix zircon, baddeleyite, and apatite mean that it is possible to combine all of these analyses into a single weighted average $^{207}\text{Pb} / ^{206}\text{Pb}$ age of $4342 \pm 3$ Ma

Table 3. Summary of U-Pb isotope data for phases (zircon, baddeleyite, apatite, merrillites) in matrix minerals and basalt clasts in MIL 13317,7 determined by SIMS analysis. Uncertainties for the isochron ages are stated at the 95% confidence level.

| Clast/Matrix | Phase     | $^{207}\text{Pb} / ^{206}\text{Pb}$ | $\pm 1\sigma$ % | Age (Ma) | $\pm 1\sigma$ (abs.) |
|--------------|-----------|-----------------------------------|-----------------|----------|----------------------|
| Clast 4 basalt | Apatite   | 0.5389                             | 0.53             | 4351     | 8                    |
| Clast 4 basalt | Apatite   | 0.5341                             | 0.46             | 4338     | 7                    |
| Clast 4 basalt | Apatite   | 0.5322                             | 0.19             | 4333     | 3                    |
| Clast 4 basalt | Apatite   | 0.5332                             | 0.19             | 4336     | 3                    |
| Clast 4 basalt | Apatite   | 0.5329                             | 0.20             | 4335     | 3                    |
| Clast 4 basalt | Apatite   | 0.5307                             | 0.16             | 4329     | 2                    |
| Clast 10 basalt | Apatite   | 0.5357                             | 0.36             | 4343     | 5                    |
| Symplectite assemblage | Baddeleyite | 0.5358 | 0.41 | 4343 | 6 |
| Symplectite assemblage | Baddeleyite | 0.5365 | 0.24 | 4345 | 4 |
| Symplectite assemblage | Baddeleyite | 0.5351 | 0.34 | 4341 | 5 |
| Symplectite assemblage | Baddeleyite | 0.5364 | 0.47 | 4344 | 7 |
| Symplectite assemblage (Clast 16) | Apatite | 0.5343 | 0.43 | 4339 | 6 |
| Melt glass/symplectite assemblage | Baddeleyite | 0.5361 | 0.19 | 4344 | 3 |
| Matrix | Zircon   | 0.5296                             | 0.42             | 4326     | 6                    |
| Matrix | Zircon   | 0.5365                             | 0.37             | 4345     | 5                    |
| Matrix | Zircon   | 0.5382                             | 0.53             | 4349     | 8                    |
| Matrix | Zircon   | 0.5318                             | 0.39             | 4332     | 6                    |
| Matrix | Apatite  | 0.5358                             | 0.39             | 4343     | 6                    |
| Matrix | Apatite  | 0.5380                             | 0.48             | 4349     | 7                    |
| Matrix | Apatite  | 0.5366                             | 0.50             | 4345     | 7                    |
| Matrix | Apatite  | 0.5323                             | 0.47             | 4333     | 7                    |
| Matrix | Apatite  | 0.5382                             | 0.45             | 4349     | 7                    |
| Matrix | Apatite  | 0.5328                             | 0.60             | 4335     | 9                    |
| Matrix | Apatite  | 0.5351                             | 0.48             | 4341     | 7                    |
| Matrix | Apatite  | 0.5352                             | 0.52             | 4341     | 8                    |
| Matrix | Apatite  | 0.5373                             | 0.63             | 4347     | 9                    |
| Matrix | Merrillite | 0.5253                             | 0.36             | 4314     | 5                    |
| Matrix | Merrillite | 0.5268                             | 0.86             | 4318     | 13                   |
| Matrix | Merrillite | 0.5363                             | 0.55             | 4344     | 8                    |
| Matrix | Merrillite | 0.5577                             | 1.31             | 4401     | 19                   |
| Matrix | Merrillite | 0.5259                             | 0.77             | 4316     | 11                   |
| Matrix | Merrillite | 0.5121                             | 0.39             | 4277     | 6                    |
| Matrix | Merrillite | 0.5255                             | 0.60             | 4314     | 9                    |
| Melt clast | Merrillite | 0.5255                             | 0.74             | 4281     | 11                   |
| Melt clast | Merrillite | 0.5202                             | 0.50             | 4299     | 7                    |
Apatite analyses were also obtained from two of the basaltic clasts (six analyses from Clast 4 and one from Clast 10), which give similar $^{207}\text{Pb}/^{206}\text{Pb}$ ages (between $4329 \pm 4$ Ma and $4351 \pm 16$ Ma) to the majority of the matrix apatite grains. A Pb-Pb isochron age of $4334 \pm 4$ Ma ($MSWD = 1.02; P = 0.39; \text{Fig. 10e}$) was calculated for Clast 4, although calculating a weighted average of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages ($4333 \pm 5$ Ma) results in a relatively high MSWD and low probability of fit ($MSWD = 2.2; P = 0.055; \text{Fig. 9}$). The relatively good statistical fit of the Pb-Pb isochron for apatite in Clast 4 compared with the scatter in $^{207}\text{Pb}/^{206}\text{Pb}$ ages may reflect an inappropriate common Pb correction, assumed to have a modern terrestrial Pb composition, and the grains actually contain a small proportion of more radiogenic (when compared with terrestrial Pb) lunar initial Pb (Fig. 10e).

The merrillite grains show a significantly wider range of Pb isotopic compositions than the apatite, zircon, or baddeleyite grains, resulting in $^{207}\text{Pb}/^{206}\text{Pb}$ ages from $4277 \pm 12$ Ma to $4401 \pm 38$ Ma. This is shown on a plot of $^{207}\text{Pb}/^{206}\text{Pb}$ versus $^{204}\text{Pb}/^{206}\text{Pb}$ (Figs. 10a and 10b), with the merrillite grains forming a separate trend below and to the right of the other phases.

**DISCUSSION**

Miller Range 13317 represents a mixture of lunar material which is of basaltic, anorthositic, and impact origin. The ages of the mineral phases (>3.9 Ga) within the sample provide unique insights into the nature of ancient lunar volcanism. In the following discussion, we synthesize the results of our study to understand MIL 13317’s geological history within the context of other lunar meteorites and Apollo samples.

**Constraining the Source Region of MIL 13317**

Due to the regolithic nature of MIL 13317, remote sensing chemical data can be used to help locate potential source regions on the Moon. Previous work has used this approach to constrain the source regions of specific chemical types of lunar rocks (Kramer et al. 2015) and lunar meteorites (e.g., Gnos et al. 2004; Arai et al. 2010a, 2010b; Joy et al. 2010b, 2011c; Calzada-Diaz et al. 2015; Zeng et al. 2018). Data from the Lunar Prospector (published in Prettyman et al. 2006),
Fig. 10. $^{207}\text{Pb}/^{206}\text{Pb}$ versus $^{204}\text{Pb}/^{206}\text{Pb}$ compositions of all the phases analyzed by SIMS in MIL 13317 (a, b). These data were also used to determine Pb-Pb isochron ages for the zircon (c), baddeleyite (d), and Clast 4 apatite (e) grains. It is possible to construct a single combined isochron (f) through all of the matrix zircon, baddeleyite, and apatite data, assuming that all of these grains originated from the same igneous precursor. PBC = common terrestrial Pb, calculated following the model assumptions of Stacey and Kramers (1975). Uncertainties on the Pb-Pb isochron ages are stated at the 95% confidence level. (Color figure can be viewed at wileyonlinelibrary.com.)
which features FeO and Th abundances in bins of 2° equal area per pixel have been used to identify areas of the lunar regolith showing compositional similarities to MIL 13317 (Fig. 12). A range of FeO (9.6–12.3 wt%, Table 1) and Th (4.6–5.4 ppm) was used to account for compositional heterogeneities between sample subsplits. Regions with similar FeO contents to MIL 13317 (red pixels, Fig. 12) are located mainly in the Procellarum KREEP Terrane (PKT) and a few outcrops in the South Pole–Aitken basin region. However, the high-Th content (blue pixels, Fig. 12) constrains MIL 13317 to the near side of the Moon with matches for the Th content only outcropping in the PKT region. The overlap of both the FeO and Th contents of MIL 13317 (green pixels, Fig. 12) occurs only in a few locations in the near side PKT region. These locations, where the FeO and Th content overlaps, are the most likely source regions of MIL 13317.

Regions on the Moon that are most similar in composition to MIL 13317 (i.e., the green pixels in Fig. 12), include the northwest region of Oceanus Procellarum, the eastern edge of Mare Frigoris, and to the east and northeast of Mare Nubium. Crater counting data sets from basalt surface units of these two regions (Hiesinger et al. 2003, 2010) indicate that the lava flows are no more than ~3700 Ma. This suggests that if MIL 13317 originates from one of these regions, then the basalt component in MIL 13317 represents ancient cryptomare basalts that were buried by subsequent lava flows and/or impact ejecta within the PKT (e.g., Head and Wilson 1992).

### Relationship to Other Lunar Meteorites

#### Differences to Other Miller Range Lunar Meteorites

MIL 13317 is the seventh lunar meteorite found in the Miller Range region of Antarctica to date. Other Miller Range brecciated meteorites include MIL 090036 and the grouped MIL 090034/090070/090075 (MIL 09 group) stones, which are feldspathic regolith breccias grouped on the basis of their chemical composition and petrography (Korotev et al. 2011; Zeigler et al. 2012), and exposure ages (Nishiizumi and Caffee 2013). The lack of basaltic material in the group of MIL 09 samples (Liu et al. 2011; Korotev and Zeigler 2014; Calzada-Diaz et al. 2015; Martin et al. 2017), and low bulk rock Th and KREEP-rich elements suggest that they are unlikely to be source paired with MIL 13317 (supporting information Fig. S5a). MIL 090036 is chemically distinct from MIL 13317, as it is more feldspathic (27.1 wt% Al₂O₃, 5.0 wt% FeO) —Korotev and Zeigler 2014: compared to MIL 13317 from this work with 15.9 wt % Al₂O₃, 12.3 wt% FeO and has no basaltic material (Fig. S5a) (Liu et al.
2011; Korotev and Zeigler 2014; Calzada-Diaz et al. 2015; Martin et al. 2017). Similar to MIL 13317, the MIL 07006 stone is a regolith breccia that contains basaltic clasts (Korotev et al. 2009; Liu et al. 2009; Joy et al. 2010a; Robinson et al. 2012). However, MIL 07006 basalts are much more Mg-rich, Ti-depleted, and KREEP-poor (e.g., Th = 0.36 ppm, Fig. S5b) compared to MIL 13317, ruling out a pairing relationship.

Miller Range 13317 is petrographically similar to the mingled anorthositic and basaltic regolith breccia Meteorite Hills (MET) 01210, which is grouped with MIL 05035 (as well as Yamato-793169 and Asuka-881757, collectively known as the YAMM meteorites: Day et al. 2006; Joy et al. 2008; Arai et al. 2010a, 2010b). MIL 13317 bulk rock Ti compositions are low (1.0 wt% TiO₂, Table 1) like MET (1.53 wt% TiO₂; Korotev and Zeigler 2014), which is indicative of a very low-Ti (VLT) to low-Ti source region. However, MIL 13317 has lower FeO (12.3 wt%) and higher K₂O (0.27 wt%) content than MET (16.4 wt% FeO, 0.05 wt% K₂O; Korotev and Zeigler 2014). Furthermore, MET 01210 contains agglutinates (Arai et al. 2010a, 2010b), which were not found in MIL 13317. In summary, we exclude a pairing relationship with any of the other Miller Range lunar meteorites found to date.

**Similarity to Th-Rich Brecciated Meteorites**

Northwest Africa (NWA) 4472/4485 are KREEP-rich lunar regolith breccias (Joy et al. 2011a), which have incorporated basaltic rock fragments and sampled Pre-Nectarian age lithics and mineral fragments (Arai et al. 2010a, 2010b; Joy et al. 2011a). However, pyroxene in basaltic clasts of NWA 4472 are generally more Mg-rich and Fe-poor (~25>Mg#<65; Joy et al. 2011a) than the basaltic clast in MIL 13317 (~0>Mg#<60). Therefore, we discount the possibility of a source-pairing relationship between these two meteorites.

Concentrations of Sc and Sm in the MIL 13317 fusion crust are similar to those in Th-rich meteorites like Calcalong Creek and Lynch 002 (Fig. 5f). Calcalong Creek (Hill et al. 1991; Hill and Boynton 2003) and Lynch 002 (Smith et al. 2012; Korotev 2013; Robinson et al. 2016) are the only lunar meteorites found in Australia (located over 800 km apart from each other). Although not paired with each other, they have features in common with MIL 13317. All three meteorites have a mixed clast assemblage of anorthosite, mare basalts, and KREEP-rich material, and have comparable Th (~4–5 ppm) and FeO (~9–10 wt%) contents (Korotev 2018). Calcalong Creek was potentially launched from around the edge of the Imbrium region (Calzada-Diaz et al. 2015), we suggested a similar region of the Moon for MIL 13317 (Fig. 12). Like MIL 13317, pyroxene in basaltic clasts of Lynch 002 show basaltic chemical fractionation trends typical of VLT to low-Ti mare basalts of the Apollo suite, and the basaltic clasts are rich in silica. These petrographic and compositional similarities may indicate a source-pairing relationship between Lynch 002 and MIL 13317, although this needs to be tested further by radiogenic and cosmogenic nuclide analysis.
(e.g., crystallization age, ejection age, terrestrial age, and trapped and cosmogenic components) of Lynch 002.

Regolith History

Miller Range 13317 is a heterogeneous mix of phases that have undergone different burial and re-exposure histories. The few spherules present affirm a regolith origin for the sample, but their scarcity suggests that the parent regolith from which the sample consolidated was relatively immature (McKay et al. 1991).

The two subsplits analyzed for noble gases have sampled different components within MIL 13317: one being dominated by a regolith breccia component (MIL 13317,10) and the other (MIL 13317,11b) dominated by a more fragmental breccia, with smaller amounts of regolith material (see Fig. S6). Analysis of the two subsplits of MIL 13317 yields exposure ages that range from 71 ± 28 to 2860 ± 74 Myr (supporting information Table S11). The anomalously high argon cosmic ray exposure durations (2860 ± 74 and 5652 ± 159 Myr: Table S11) from two of the subsplits, are likely to be a result of the sample being saturated with implanted parentless $^{40}$Ar$_{tr}$ and solar wind implanted $^{36}$Ar$_{tr}$. As a result, we cannot adequately correct for the presence of these components in these subsplits, which has an effect on the calculated $^{38}$Ar$_{cos}$. Although the Ne exposure ages are generally consistent between subsplits, they are lower than those obtained from cosmogenic Ar and Xe isotopes. This may be explained by partial loss of Ne, potentially during an impact event. Thus, a cosmic ray exposure age of 835 ± 84 Myr is adopted for MIL 13317 based on the $t_{12\alpha}$ exposure age, as heavy noble gases are less likely to have been affected by shock-related diffusional gas loss or subject to large corrections for solar wind-derived trapped components.

The lithification of MIL 13317 from soil into breccia components is estimated to be between 2610 ± 780 and 1570 ± 470 Ma (determined from the $^{40}$Ar/$^{36}$Ar$_{tr}$ ratio antiquity indicator). MIL 13317 was consolidated into its current form and exposed to the cosmic environment for an overall period of 835 ± 84 Myr, where it has experienced a complex history at an average depth of ~30 cm.

Implications for the Age Resetting History of MIL 13317

The ages determined from the zircon, baddeleyite, and the majority of the phosphate grains in MIL 13317 provide evidence of ancient lunar volcanism occurring at or before ~4340 Ma (see also Snape et al. 2018). Although it is possible that these ages could represent the timing of impact induced resetting of the U-Pb system, this would still require formation of the igneous protolith(s) from which the grains originated, prior to this time. Equally, while it is possible that different grains within the meteorite breccia matrix originate from different protoliths, the association of the matrix apatite and baddeleyite grains with similar symplectite assemblages, combined with the similarity in their $^{207}$Pb/$^{206}$Pb ages, suggests that the simplest explanation for the similar ages identified in the majority of the grains is that they originated from the same igneous protolith. Based on this interpretation, the weighted average $^{207}$Pb/$^{206}$Pb age of 4342 ± 3 Ma (MSWD = 0.99; $P = 0.47$; Fig. 9), or the combined Pb-Pb isochron age of 4342 ± 4 Ma (MSWD = 1.14; $P = 0.31$; Fig. 10f) for the matrix zircon, baddeleyite, and apatite analyses provides the best estimate for the crystallization age of the igneous protolith. These ages are similar to those reported by Shaulis et al. (2016) for matrix phosphate and zircon-bearing phases (4351 ± 9 Ma) in a separate thin section (MIL 13317,5) of the same meteorite. The apatite grains within basalt Clast 4 appear to be slightly younger (an average $^{207}$Pb/$^{206}$Pb age of 4333 ± 5 Ma and a Pb-Pb isochron age of 4334 ± 4 Ma) than the matrix grains, but are consistent with Pb-Pb isochron ages (4332 ± 2 Ma) determined for several other basaltic clasts in the MIL 13317 meteorite in a separate study (Snape et al. 2018). These younger ages may indicate that the clasts represent fragments of a separate mafic protolith; however, this is difficult to confirm given the tight clustering of ages, and the uncertainties associated with correcting for terrestrial Pb contamination.

The range of Pb isotope compositions in the merrillite grains may indicate that the U-Pb system in this phase is more easily reset than in apatite. However, this is contrary to the behavior observed in Ca-phosphate minerals in Apollo breccias (e.g., Merle et al. 2014; Snape et al. 2016; Thiessen et al. 2017, 2018), where breccia formation ages based on consistent $^{207}$Pb/$^{206}$Pb ages in both apatite and merrillite phases, indicate that the U-Pb system is reset in both phases to a similar extent. Furthermore, the youngest $^{207}$Pb/$^{206}$Pb ages of the merrillite analyses (~4280 Ma) in MIL 13317 are consistent with younger ages reported by Shaulis et al. (2016) of 4276 ± 21 Ma for two phosphate grains and 4270 ± 24 Ma for 22 baddeleyite grains, which they interpret as representing the crystallization age of a separate basalt lithology.

Although the mechanism by which the younger apatite grain was introduced to the breccia is unclear, it is notable that the U-Pb age of 3890 ± 33 Ma is close to that determined for the formation of the Imbrium impact basin in several studies (Gnos et al. 2004; Liu et al. 2011) including analyses of phosphates in breccias from Apollo 12, 14, and 17 (Snape et al. 2016; Thiessen et al. 2017, ...
2018). Without precisely knowing the provenance of the meteorite or this particular apatite grain, it is unclear if this age is (1) indicative of a link with the formation of the Imbrium basin, (2) a reflection of the general last stage of basin formation, or (3) a coincidental intermediate partial resetting age between the original crystallization of the apatite and a later impact event. As discussed earlier, considering the susceptibility of the apatite U-Pb system to be reset by impact events, and the observed tendency for phosphate phases analyzed in Apollo breccias to be completely reset (e.g., Nemchin et al. 2009b; Snape et al. 2016; Thiessen et al. 2017, 2018), the third interpretation seems slightly less likely. Without a clearer understanding of the Moon’s impact history, and a resolution to the long-standing debate surrounding the potential ~3900 Ma lunar cataclysm (Bottrke and Norman [2017] and references therein; Morbidelli et al. [2018] and references therein), it is difficult to say which of the first two interpretations are more likely.

Geological Context and Significance of MIL 13317

Petrology and composition of MIL 13317 suggests that the sample is formed from a regolith in which there was a mixture of mare-like basaltic material and highland anorthositic material, with a minor KREEP component (Figs. 5 and 7). Based on comparisons with remote sensing data (Fig. 12), our best estimate is that the MIL 13317 meteorite is derived from the edge of the PKT on the lunar near side. Assuming this is the case, the mineral and clast chemistry, and age of components can provide new information about the geological evolution of the lunar crust in this region of the Moon.

What Types of Highland Rocks and Impact Melts Are Distributed in Different Crustal Regions?

MIL 13317 contains highland igneous rocks and impact melt breccias that are mainly associated with the FAN suite found at many of the Apollo landing sites. For example, plots of [Eu/Sm]CI versus Sm in plagioclase can be used to distinguish parent lithologies (Fig. 5d) (Russell et al. 2014) and indicates that plagioclase in Clast 5 falls within the field of lunar anorthosites as sampled by the Apollo missions and lunar meteorites. Trace element systematics for plagioclase from granulitic Clast 12 overlap with that of plagioclase from Clast 5 (Eu/Eu* = 15.83; [La/Sm]CI = 1.77) and are also similar to the Apollo FAN suite (Figs. 5c and 5d). Granulitic Clast 14 is more similar to magnesium granulite found in feldspathic lunar meteorites (Treiman et al. 2010; Gross et al. 2014; Kent et al. 2017) that fall into the compositional gap between Apollo FAN and HMS (Fig. 7a). Furthermore, Clast 9 in particular has sampled material which is a mix of KREEP, feldspathic, and basaltic material with trace element characteristics that are distinct from the Apollo impact melts. These compositions and petrology are generally represented by the mineral grains in the matrix which have an affinity to ferroan anorthosite parent rocks.

In general, there is an apparent lack of HMS rock types in MIL 13317 considering the likely source region. The distribution of HMS rocks has been strongly linked to an origin in and surrounding the PKT region due to their possible association with KREEP-rich material (Snyder et al. 1995; Korotev 2000; Wieczorek et al. 2006). Similarly, the FANs have been associated with rocks typical of the feldspathic lunar highlands. However, feldspathic lunar meteorites, which represent a more global sampling of the feldspathic lunar highlands, generally lack Apollo-like FAN, KREEP, and HMS material, and have a feldspathic component that is more magnesian (e.g., the magnesian anorthosites) than the Apollo FAN suite (Korotev et al. 2003, 2012; Korotev 2005; Zeigler et al. 2012; Gross et al. 2014). This suggests that the typical highland rocks and impact melts found during the Apollo missions (e.g., FANs, HMS) may not be globally distributed across the Moon (Gross et al. 2014). The highland components of MIL 13317 are compositionally more similar to the Apollo highland rock suite than typical feldspathic lunar meteorites, supporting this suggestion.

Was There a Single Magmatic Source for Basalt Clasts in MIL 13317?

Basalt clasts in MIL 13317 provide further insight into the magmatic processes occurring in regions remote from the Apollo and Luna landing sites, both in terms of composition and basaltic volcanism through time. The abundances of Sc and Sm in some bulk clasts (e.g., Clast 18, Fig. 5f) show that these clasts have also incorporated KREEP-rich material and are intermediate to the Apollo VLT-mare basalts, KREEP-rich basalts, and Apollo impact melts (see also Zeigler and Korotev 2016). The Fe# to Ti# correlations of pyroxene suggest that all the basalt clasts appear to be derived from lavas intermediate to the Apollo VLT and the low-Ti mare basalt source regions (Fig. 4b; after Arai et al. 1996; Robinson et al. 2012). Pyroxene in basalt clasts show a typical fractionation trend for lunar basalts of increasing Fe# (atomic Fe/Fe + Mg, Fe# of 0.40–0.99) with increasing Ti# (atomic Ti/Ti + Cr, Ti# of 0.25 to 1) (Fig. 4b). The Al/Ti ratio can be compared with the Fe# to indicate element partitioning from the parental melt to further understand this crystallization trend (Nielsen and Drake 1978; Arai et al. 1996). The trends in MIL 13317 basalt clasts show that pyroxene (Fe# 40 to Fe# 46) was first to crystallize out of the parent melt (Al/Ti ~>3) (Fig. 4c). As
the melt evolved, plagioclase began to co-crystallize with pyroxene (Fe# 46 to Fe# 70) and the Al/Ti ratio decreased as Al was removed from the melt. An Al/Ti ratio of ~1.5 indicates that Ti was removed from the melt and ilmenite co-crystallized with pyroxene and plagioclase, as the Al/Ti versus Fe# trend flattens out from Fe# 70 and Al, Ti, and Fe are all extracted from the melt. At this point, the parent magma would have had an Fe-enrichment allowing for a late-stage formation of symplectites. These trends support a common igneous origin for the basalt clast in MIL 13317 and suggest that the clasts are comagmatic, and that differences in texture must be caused by slightly different crystallization histories.

**Implications for Ancient Lunar Volcanism**

Based on ages of the Ca-phosphate phases in basalt Clast 4, and the ages determined for the basalt clasts analyzed by Snape et al. (2018), the basaltic material sampled by MIL 13317 is interpreted as having crystallized at ~4330 Ma (Figs. 9–11; Table 3). Thus, VLT and low-Ti basaltic volcanism on the Moon recorded by lunar meteorites extended across 1.4 billion years, from ~4330 to ~2930 Ma (recorded by young lunar meteorite NWA 032: Borg et al. 2009). The ancient ages of MIL 13317 places it among the oldest suite of lunar basalt samples, including the Apollo 14 high-Al basalts (Taylor et al. 1983; Snyder and Taylor 2001), Apollo 17 KREEP basalts (Nyquist et al. 1975; Shih et al. 1992), and the high-Al, VLT lunar meteorite Kalahari 009 (Terada et al. 2007; Shih et al. 2008; Sokol et al. 2008; Snape et al. 2018) (Fig. 13c). Observations from the Apollo suite (e.g., Nyquist and Shih 1992) indicate a correlation between the age of the basalt and Ti-content, with older mare basalts (e.g., Apollo 11 basalts; ~3900 to 3500 Ma) having higher Ti-content, and the younger basalts (e.g., Apollo 12, ~3400 to 3200 Ma; Luna 24 ~3200 Ma) having more intermediate to lower Ti-content (Fig. 13a). Using the method of Robinson et al. (2012), we calculate the parent bulk-rock Ti-contents of basalt clasts in MIL 13317 to be between 0.14 and 0.32 wt% making it a VLT basalt (Fig. 13a), as well as having a high-Al content (Fig. 13b; Table 1). MIL 13317 and another ancient high-Al, VLT lunar meteorite Kalahari 009 (Figs. 13a and 13b) indicate that the most ancient (~3800 Ma) mare and mare-like basalts in lunar meteorites tend to have some of the lowest Ti-contents and highest Al-contents of all lunar sampled basalts (Figs. 13a and 13b; see also Joy and Arai 2013). Further age dating of basaltic clasts in other lunar meteorites is needed to see if this trend is supported across a wider geographic distribution of lunar material. However, the observation of the most ancient lunar basalts being low in Ti (<3 wt% TiO₂) is also seen in coupled crater counting-chemical remote sensing studies of mare basalt regions (e.g., Kodama and Yamaguchi 2005; Morota et al. 2011; Kramer et al. 2015; Sato et al. 2017). The discovery of new types of high-Al basalts suggests that this variety may be a more common basalt type than suggested by their relatively low abundances in the Apollo and Luna sample collection (Kramer et al. 2015).

Outstanding questions relate to how such early melting is associated with the timing of LMO ilmenite-driven density instabilities and overturn (e.g., Hess and Parmentier 1995; Elkins-Tanton et al. 2002), and the role of urKREEP (Apollo 14: e.g., Neal and Kramer 2006) or lack thereof (MIL 13317 and Kalahari 009: Terada et al. 2007; Snape et al. 2018) driving this earliest phase of magmatic activity. Terada et al. (2007) promoted impact-driven decompression melting as a possible mechanism of such ancient KREEP-poor volcanism. However, partial melting of the earliest olivine and pyroxene LMO cumulates during LMO overturn and adiabatic ascent is also feasible (Joy et al. 2008; Elkins-Tanton and Bercovici 2014).

The relationship between high-Al abundances in the earliest ~4300 Ma volcanism is also intriguing. High-Al abundances in Apollo 14 basalts have been modeled via assimilation fractional crystallization, where the assimilant was a high-Al granitic component either added during ascent (Neal and Kramer 2006; Hallis et al. 2014) or via thermal erosion of the underlying regolith during the flow of lava across the lunar surface (Hui et al. 2011). However, both MIL 13317 and Kalahari basalts are KREEP-poor (Terada et al. 2007; Snape et al. 2018; this study), and the high-Al component instead could have been contributed from assimilation of ITE-poor ferroan anorthositic crust. There are challenges with assimilating such refractory materials into early lunar VLT/low-Ti mafic magmas (Finnila et al. 1994) during magma chamber storage (Neal et al. 1988; Head and Wilson 1992) or via thermal erosion of the underlying regolith. However, partial melting of the earliest phase of magmatic activity. Terada et al. (2007) promoted impact-driven decompression melting as a possible mechanism of such ancient KREEP-poor volcanism. However, partial melting of the earliest olivine and pyroxene LMO cumulates during LMO overturn and adiabatic ascent is also feasible (Joy et al. 2008; Elkins-Tanton and Bercovici 2014).
Fig. 13. Comparison of MIL 13317 mineral phase ages with (a) lunar basalts TiO$_2$ content (adapted from Joy and Arai 2013),
(b) lunar basalt Al$_2$O$_3$ content, and (c) the range in crystallization ages for non-mare ancient highlands rocks and KREEP
basalts (see text for references). (Color figure can be viewed at wileyonlinelibrary.com.)
~4300 Ma (Elkins-Tanton and Bercovici 2014). Such elevated crustal thermal conditions may have facilitated dissolution (Reiners et al. 1995) of anorthitic plagioclase from the warm country feldspathic rock. This would allow an Al-rich assimilant to be available in ancient high-Al basaltic lava flows. Alternatively, the Hui et al. (2011) model of post-eruption assimilation by a lava flow thermally eroding the underlying KREEP-poor feldspathic glass-rich regolith might be an attractive alternative hybrid magma-surface interaction model. The nature of transport of basalt magmas through, and onto, hot and cold lunar crust should be explored to better understand ancient basalt petrogenesis (Finnila et al. 1994) and investigate effects of how crustal assimilation could mask isotopic signatures of mantle sources (Reiners et al. 1995) in such ancient basalts.

**SUMMARY**

Miller Range 13317 is a mingled regolith breccia of lunar origin comprised of fragments of basalts, impact melts, and feldspathic material (including clasts with an affinity to the Apollo FANs). It is one of a handful of lunar meteorites available that have sampled ancient (>3800 Ma) basaltic volcanism. It is distinct from any of the other lunar meteorites collected to date and there is no substantial evidence that MIL 13317 is paired with any other Th-rich lunar meteorite, although further investigation is needed on Lynch 002 to confirm this. As a result, MIL 13317 provides new perspectives on the diversity of rocks on the lunar surface, and in particular lunar volcanism through time. MIL 13317 shows a complex regolith history with a range in trapped and cosmogenic noble gas isotope concentrations. However, all parts of the meteorite have resided in the top millimeter of the lunar surface at some point in history, where trapped gases from the solar wind and lunar atmosphere were acquired. The $^{40}\text{Ar}/^{36}\text{Ar}$ antiquity indicator suggests that compaction from the parent soil into breccia components occurred between 2610 ± 780 and 1570 ± 470 Ma, which closed the fragments off from further space weather processes. MIL 13317 then experienced exposure to the cosmic environment for a period of approximately 835 ± 84 Myr, where it resided in the regolith at an average depth of 30 cm.

Comparing Lunar Prospector data with the bulk composition of the Moon, it is clear that the basaltic clasts in MIL 13317 provide evidence of ancient (~4340 Ma) VLT, high-Al basaltic volcanism that was not sampled by the Apollo missions, possibly derived from cryptomare deposits that were buried by subsequent lava flows and/or impact ejecta. These ancient ages suggest that basaltic volcanism, in particular high-Al, VLT basaltic volcanism, on the Moon started as early as 4350 Ma and before the bulk onset of lunar mare volcanism at ~3850 Ma, providing further constraints on the earliest volcanic activity of the Moon.

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SUPPORTING INFORMATION

Additional supporting information may be found in the online version of this article:

Data S1. Analytical techniques.

Fig. S1. Backscattered electron images of grains used in SIMS U-Pb dating work. Image directory and corresponding data for each grain analysis can be found in Table S3. Zirc = zircon; badd = baddeleyite; phos = phosphate.

Fig. S2. Thin section images of MIL 13317 in plain polarized light (ppl) and crossed polarized light (xpl).

Fig. S3. Cathodoluminescence images of basalt Clast 1 in Miller Range (MIL) 13317,7. Unshocked regions of the plagioclase grains show up as yellow. Plagioclase regions that are shocked show up as deep red-purple.

Fig. S4. $^{134}$Xe/$^{132}$Xe versus $^{136}$Xe/$^{132}$Xe for MIL 13317,10. The data for MIL 13317,10 plot in the solar dominated region with the $^{238}$U, $^{235}$U, and $^{244}$Pu endmembers plotting far off to the right of the plot.

Fig. S5. a) Comparison of FeO (wt%) versus Al$_2$O$_3$ (wt%) showing the fusion crust composition for MIL 13317,7. MIL 13317 is compared to other Miller Range and the YAMM group of basaltic meteorites (data from Korotev and Ziegler 2014). MIL 13317 is also compared to another Th-rich meteorite Calcalong Creek (Hill and Boynton 2003) and Kalahari 009. b) FeO (wt%) versus Th (ppm). Plot shows a comparison of lunar meteorite (blue dots and markers) and Apollo and Luna soils (represented by the colored fields, Korotev 2018) with the fusion crust data for MIL 13317 (Table 1). Note the log scale for the Th concentrations.

Fig. S6. Chips of Miller Range 13317,10 and Miller Range 13317,11. Subsplits from both chips were used for noble gas analyses. A polished block was made of MIL 13317,11b. MIL 13317,10 represents a regolith dominated chip and MIL 13317,11 represents a large brecciated basaltic clast.

Table S1. Summary and types of analysis conducted on clasts in MIL 13317,7 and MIL 13317,11a.

Table S2. Trace elements abundances (ppm) for clasts and matrix minerals and bulk compositions in MIL 13317,7. Errors are 1σ SD from the averages of several measurements (number of analysis displayed in brackets with the mineral, where px = pyroxene and plag = plagioclase).

Table S3. U-Pb isotope data for phases (zircon, baddeleyites, apatites, merrillites) in MIL 13317,7 determined by SIMS analysis.

Table S4. Plagioclase compositional summary data for MIL 13317,7 and MIL 13317,11a.

Table S5. Olivine compositional summary data for MIL 13317,7.

Table S6. Pyroxene compositional summary data for MIL 13317,7 and MIL 13317,11a.

Table S7. Fusion crust and vein compositional summary data for MIL 13317,7.

Table S8. Bulk values and averages of compositional data for fine-grained clast in MIL 13317,7.

Table S9. Summary and raw data of Neon and Argon noble gas data for bulk rock chips of MIL

Table S10. Summary and raw data of Xenon noble gas data for bulk rock chips of MIL 13317,10. Raw data include blank corrected and mass spectrometer sensitivity corrected data. Errors are to 1σ.

Table S11. Summary of endmember noble gas components, production rates, cosmic ray exposure ages, and antiquity ages in MIL 13317,10 and MIL 13317,11b.