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Manuscript Number: GCA-D-13-00516

Associate Editor: Dr. Marc Norman

Number of Figures: 14

Number of Tables: 2

Number of Electronic Appendices: 2

- EA1 = supplementary figures (11 figures)
- EA2 = supplementary tables (8 tables)

Petrography and geochronology and source terrain characteristics of lunar meteorites Dhofar 925, 961 and Sayh al Uhaymir 449

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Keywords: lunar-meteorite, lunar-regolith, South-Pole-Aitken-Basin, KREEP

Proposed running header: Lunar meteorites Dho 925 / 961 and SaU 449

Abstract

Dhofar (Dho) 925, 961 and Sayh al Uhaymir (SaU) 449 are brecciated lunar meteorites consisting of mineral fragments and clasts from a range of precursor lithologies including magnesian anorthositic gabbronorite granulites; crystalline impact melt breccias; clast-bearing glassy impact melt breccias; lithic (fragmental) breccias; mare basalts; and evolved (silica-rich) rocks. On the similarity of clast type and mineral chemistry the samples are likely grouped, and were part of the same parent meteorite. Phosphate Pb-Pb ages in impact melt breccias and matrix grains demonstrate that Dho 961 records geological events spanning ~500 Ma between 4.35 Ga to 3.89 Ga. These Pb-Pb ages are similar to the ages of 'ancient' intrusive magmatic samples and impact basin melt products collected on the lunar nearside by the Apollo missions. However, the samples' bulk rock composition is chemically distinct from these types of samples, and it has been suggested that they may have originated from the farside South Pole-Aitken impact basin (i.e., Jolliff et al., 2008). We test this hypothesis, and conclude that although it is possible that the samples may be from the South Pole-Aitken basin, there are other regions on the Moon that may have also sourced these complex breccias.

1. Introduction

The Moon is a witness plate to Solar System processes and preserves a record of the geological evolution of small planetary bodies (NRC, 2007). Manned and unmanned missions to the Moon returned ~382 kg of lunar rocks and soils (Vaniman et al., 1991). These were collected from within and around the nearside Procellarum KREEP Terrane (PKT) by the Apollo missions, and from equatorial latitudes on the eastern limb by the Luna missions. Therefore, interpretations of the Moon's

past have mostly been derived from a geographically restricted dataset on the lunar nearside. Lunar meteorites, which are sourced from potentially anywhere on the Moon's surface, however, provide a better global representation of the geological and chronological history of the Moon (Korotev, 2005; Joy and Arai, 2013). To date, there have been ~185 individual (named) lunar meteorites collected on Earth as hot and cold desert finds. These originated from perhaps as few as 40-50 source craters on the Moon (Basilevsky et al., 2010). Radiogenic isotope studies indicate that the majority of known lunar meteorites have been launched from the Moon in the last 10 Myr, and all have been launched in the last 20 Myr probably from small craters only a few kilometres or less in diameter (Warren, 1994; Head et al., 2002).

Remote sensing datasets provide information about the chemical and mineralogical diversity of the lunar surface (i.e., Lunar Prospector and the Kaguya gamma-ray spectrometer chemical data, Clementine, Chandrayaan-1 Moon Mineralogy Mapper, and the Kaguya Spectral Profiler spectral datasets). Although the spatial scales of these mapping efforts are often on the scale of hundreds of metres to tens of kilometre (depending on the method), many previous studies have used these datasets to test chemical and mineralogical similarities to lunar meteorite and infer potential source regions. For example, feldspathic lunar meteorites (i.e., samples with bulk rock FeO <7 wt%: Korotev et al., 2009) to origins in the highlands on the farside of the Moon (Palme et al., 1991; Korotev et al., 2003; Warren, 2005; Warren et al., 2005; Nyquist et al., 2006; Takada et al., 2006; Arai et al., 2008; Joy et al., 2010a; Yamaguchi et al., 2010; Fritz, 2012). Basaltic meteorites (i.e., bulk rock FeO >17 wt%) have been linked to different mare basalt lava flow units predominantly on the nearside of the Moon (Joy et al., 2008; Fernandes et al., 2009; Arai et al., 2010; Robinson et al., 2012). Meteorites of intermediate-Fe composition (i.e., with bulk compositions between 7 and 17 FeO wt%) and with high concentrations of thorium (>2 ppm) and other incompatible trace elements (ITEs) have previously been linked with high-Th regoliths on the nearside of the Moon in the Procellarum KREEP Terrane

(Gnos et al., 2004; Joy et al., 2011a), and tentatively with the South Pole-Aitken (SPA) impact basin on the farside of the Moon (Hill and Boynton, 2003; Korotev et al., 2007; Mercer et al., 2013).

Lunar meteorites Dhofar (Dho) 925 and 961 and Sayh al Uhaymir (SaU) 449 are breccias of intermediate-Fe composition (Demidova et al., 2005, 2007; Korotev et al., 2009; Korotev, 2012). Henceforth, this group of stones will be collectively referred to here as the Dhofar group. They were collected in Oman and are thought to have originated in the same meteorite fall, and are also grouped with the Dho 960 stone (Demidova et al., 2005, 2007; Korotev et al., 2010; Korotev 2012). All stones are formally classified as impact melt breccias (Russell et al., 2004, 2005; Connolly et al., 2007). The meteorites have elevated concentrations of Th (1-3 ppm: Table 1) compared with many other intermediate-Fe brecciated lunar meteorites, indicating inclusion of an ITE-rich component. Previous studies of Dho 961 (e.g., Jolliff et al., 2007, 2008, 2009; Korotev et al., 2007, 2009, 2010; Zeigler et al., 2010a, 2010b, 2013) report that the bulk rock composition is not consistent with Apollo samples sourced from the Procellarum KREEP Terrane. Zeigler et al., (2013 and refs. therein) argue that the meteorite may have originated from the South Pole-Aitken basin, which is the other notable Th-rich (i.e., ITE-rich) region of the Moon (Jolliff et al., 1998)

Here we report the composition, mineralogy and chronology of the Dhofar group of meteorites to investigate their geological history, and test the hypothesis that the samples represent South Pole-Aitken basin material. A launch locality in SPA would be significant, as geological samples from this massive impact basin are expected to hold the answer to several key lunar science questions (NRC, 2007; Jolliff et al., 2010) including: (i) the age of the basin, which is believed to be the largest and one of the oldest impact basins on the Moon (Wilhems et al., 1987; Spudis, 1993). Defining its age will help to constrain the early lunar bombardment record, which may help to anchor the early Earth-Moon impact flux chronology (NRC, 2007; Norman, 2009); (ii) determine the extent and nature of

products of the Moon's differentiation by studying igneous rock samples from the lunar farside (e.g., Arai et al., 2008; Ohtake et al., 2012; Gross et al., 2014); (iii) characterise products of impact melt sheet differentiation (e.g., Vaughan et al., 2012; Vaughan and Head, 2014; Hurwitz and Kring, 2014); (iv) determine the composition and timing of farside mare volcanism to shed light on the magmatic history of the Moon (e.g., Hagerty et al., 2011); (v) directly sample lunar mantle material, which may have been excavated during the SPA basin-forming event (Pieters et al., 1997; Yamamoto et al., 2010; Potter et al., 2012), helping to characterise the stratification of the mantle and address models of lunar differentiation and evolution (Elardo et al., 2011; Elkins-Tanton et al., 2011).

2. Samples and Method

We obtained three authenticated meteorite chips (EA1.1 to EA1.3) of Dho 925 (0.136 g), Dho 961 (0.331 g) and SaU 449 (0.764 g). Two 100 µm thick sections (named Dho 925,1, Dho 925,2 and Dho 961,1 and Dho 961,2) and a 30 µm thin section (named Dho 925,3 and Dho 961,3) were prepared from each of the Dho 925 and 961 stones using Buehler Epo-Thin resin. The SaU 449 sample was split into two chips, and the larger portion (0.535 g) was prepared as two 100 µm thick sections (named SaU 449,2 and SaU 449,3) and a 30 µm thin section (named SaU 449,1) using Buehler Epo-Thin resin. Water was not used during the section making process. The samples were studied and photographed using an optical Leica M205C microscope with a Leica DFC 490 camera at NASA Johnson Space Center (JSC).

The Dho 925,1, Dho 961,1 and SaU 449,3 sections (Fig. 1, EA1.1-EA1.3) were all analysed in further detail for mineral chemistry by electron microprobe. Sections Dho 925,1 and Dho 961,1 were also analysed for trace element composition using *in situ* laser ablation inductively coupled mass spectrometry (LA-ICP-MS). U-Pb chronology by ion microprobe was conducted on Dho 961,1.

Petrography and mineral chemistry methodology. The sections were carbon coated and mapped using the NASA JSC JEOL 7600f Field Emission Gun Scanning Electron Microscope (FEG-SEM) with a beam current of 15 nA and an accelerating voltage of 20 to 30 kV. The SEM has a faraday cup, so we can set the sample current at a known value, and check at the end of each run to gauge beam stability. The system was coupled to a Thermo Scientific EDS (electron dispersive spectrometer) with NORAN System Six (NSS) software to derive <1 µm per pixel back-scatter electron (BSE) images and spatially resolved element data (~3 µm per pixel). Images were collected at a magnification of ×150. Each pixel of data that is collected retains a complete 0-20 KeV energy spectrum and, therefore, we were able to extract maps of C, O, Na, Mg, Al, Si, P, S, Cl, K, Ca, Ti, Cr, Mn, Fe, Co and Ni data. Element maps were normalised to the same brightness scale, colourised and recombined to make false-colour element maps using the ImageJ software package. We use the colour scheme (Joy et al., 2011b; Niihara et al., 2013) blue=silica, green=Mg, red=Fe, white=Al, yellow=Ca, pink=Ti, cyan=K to study qualitative compositional variability of individual clasts, and blue= Ca, red=Fe and green=P to distinguish phosphate phases (Fig. 1 and EA1.1 to EA1.3). Fragments of interest (EA1.4 to EA1.9) were then studied using back-scattered electron imagery with the NASA JSC JEOL 5910LV SEM at different magnifications.

Mineral and bulk fragment compositions were derived using the NASA JSC Cameca SX100 electron microprobe (EMP). Natural mineral and pure metal standards were used to calibrate the EMP measurements. For mineral phases, analyses were performed at 15 kV and 20 nA using a focused beam of 1 µm beam. Peak count times were 10 to 50 s. Cobalt concentrations were corrected for the Fe-kα and Co-kβ peak overlap. Detection limits are typically 0.03-0.04 wt% for Al_2O_3, K_2O, CaO and CoO, 0.05 wt% for Na_2O, MgO, SiO_2 and MnO, 0.06-0.09 wt% for P_2O_3, SO_2, TiO_2, Cr_2O_3, and NiO.

For bulk clast compositions the beam was defocused to a diameter of 20 µm and the same current and voltage settings noted above. For each clast an average composition is presented of data with totals between 95-105 wt% and where errors represent two standard deviation of the data points averaged together. Results of the defocused beam investigations were not corrected for unequal phase density effects (Warren, 1997; Lindstrom et al., 1999; Joy et al., 2010a). This is because many of the impact melt breccia clasts analysed have highly heterogeneous clast components and grain sizes, meaning that any attempt to correct for bulk clast mineral mode would not be significant for an individual 20 µm spot analysis. Therefore, major element bulk clast compositions should be viewed as quantitative to only within the two standard deviation quoted (Table EA2.4). We direct the reader to the study of Lindstrom et al. (1999) for a discussion about possible relative errors that may be incurred through using this technique.

We analysed apatite and merrillite also using the NASA JSC Cameca SX100 instrument. A defocused beam of 5 µm and beam settings of 15 KV and 10 nA were employed. We analysed for F, Cl, Ca, P, Si, Na, S, Al, Fe, Mg, Y, La, Ce, and Sm. Element peaks were analysed between 10 and 60 s. Fluorine counts were measured using a PC1 crystal and were determined using a time regression intercept analysis to mitigate F-migration effects following the methods of McCubbin et al. (2010, 2011) and Henderson (2011). We analysed F for 42 to 60 s in 6 s time bins (42 s gave us the most repeatable results). Apatite grains with positive time-intercept slopes were indicative of reliable F-measurements. Fluorine was calibrated using a F-bearing topaz, with each analysis made on a different location on the standard to reduce F-migration effects (Stormer et al., 1993). Chlorine was calibrated on a tugtupite crystal. P and Ca were calibrated to Durango apatite. Yttrium, La, Ce and Sm were calibrated to rare earth element (REE) phosphates. Sulphur was calibrated to a S-bearing hydroxylellestadite (called JSC standard LL_Apat: Onac et al., 2006). Magnesium and Fe were standardised to a kaersutite amphibole, and Na and Al to an oligoclase. Peak background offsets were

measured following the guidelines of Henderson (2011). The setup was tested on three phosphate standards as unknowns (Durango, LL_Apat, and Apatite-9). To calculate the amount of OH by stoichiometry an assumption about the REE-content of the apatite grains is necessary. We followed the method of McCubbin et al. (2011) in using EMP derived Ce to estimate the trivalent LREE and Y to estimate the trivalent HREE. If the recorded values were below the element detection limit then the detection limit itself was used as an upper limit estimate of REE value (see Table EA2.4). The OH group was then calculated as the difference between 1 anion per formula unit and the measured F and Cl anions, following the assumption of total X-site occupancy and that nothing other than F, Cl and OH entered the X-site. Analysis of standards showed that our technique produced acceptable measurements of F (e.g., 7 % relative error to known Durango standard value of 3.53 wt%) and Cl (e.g., 25 % relative error to known Durango standard value of 0.41 wt%), bearing in mind the low concentrations of these elements (particularly Cl). Likewise, calculated H-abundances in the LL_Apat standard were similar to known values (e.g., 16 % relative error to known LL_Apat standard value of 0.16 wt%: Onac et al., 2006), providing us with confidence that our approach is valid to quantify OH in terrestrial standard and lunar apatites.

Trace element analysis methodology. We measured minor and trace element abundances in individual mineral phases, and also estimated the bulk composition of impact melt breccia clasts and a melt vein in Dho 961. We measured 40 elements by LA-ICP-MS at the University of Houston using a Varian quadrupole ICP-MS and Photon Machines Analyte 193 laser ablation system over 2 analytical sessions. The laser was operated at 10 Hz with a fluence of 2.99 J/cm^2. We used two different beam sizes: 50 µm circular for individual mineral analyses, and either 50 or 150 µm circular for bulk clast measurements (where several measurements may have been averaged together to give a bulk clast composition: Table EA2.4). Background conditions were monitored for 20 s and the sample was ablated for 25 s. Data were reduced using the GEMOC Glitter software programme (http://www.glitter-gemoc.com/) where plots of counts per second versus time were examined for

each element per analysis, and integration intervals for the gas background and the sample analysis were selected manually. Reported errors for individual mineral phases are two sigma calculated by the Glitter software (Tables EA2.1-2.3). Reported uncertainties for bulk clasts represent two standard deviation of repeat analysis within the clast or, if only one measurement was made, two sigma analytical measurement errors (Table EA2.4).

For most analyses, calcium (^{42}Ca) was used as an internal standard using the wt% abundances of CaO in minerals and bulk clasts obtained by defocused EMP analysis. For analyses of olivine, ^{25}Mg was used as an internal standard with the wt% abundance of MgO in olivine determined by EMP analysis. Measurements were externally calibrated with the US Geological Survey (USGS) Basalt Hawaiian Volcanic Observatory (BHVO-2G) basaltic glass reference materials under the same conditions as the samples (i.e., when the laser was 50 µm for the lunar sample it was the same size for the standards, and likewise for the 150 µm analyses) (see TableEA2.8). A USGS Basaltic Icelandic basalt (BIR-1G) glass was monitored as an unknown standard (see Table EA2.8). In all cases, repeatability of the standard measurements and accuracy to GeoRem reference values (Jochum et al., 2005) was generally better for the larger 150 µm beam size. Repeatability of the BHVO measurements at both spot sizes for elements at concentrations > 0.1 ppm has a relative standard deviation range of between 1 and 7% for all elements analysed. Accuracy and precision of the BIR standard measurements are generally within < 10% of absolute values reported by GeoRem (most are < 5%) and a relative standard deviation range of < 7% (excluding Rb, P, K, Gd, and Ge, and elements < 0.1 ppm in concentration), respectively (Table EA2.8).

We observed that after the first day of data collection (of the Dho 925,1 section) the BIR measurements had anomalously ~33-45% low ^{157}Gd compared with the reference value of Gd, although the Dho 925,1 meteorite data itself do not appear to have low Gd-anomalies compared with other trivalent REE. We assessed the oxide production rate during instrument warm-up by monitoring

the CeO^+/Ce^+ ratio under the same analytical conditions used for the sample analyses. This procedure consisted of monitoring the ion intensity ratio of $^{140}Ce^{16}O$ (+^{156}Gd)/^{140}Ce during a line scan across NIST 612 glass. We tuned the system until the ratio was at the lowest value and we calculated the oxide production rate by subtracting the ^{156}Gd signal. We calculated an oxide production rate that was less than 1%, and typically less than 0.5%, making oxide interference corrections unnecessary. For the first session, it is likely that there was a higher oxide production rate that was not recognised during the instrument warmup. This issue was not observed during the second analytical session. As a further precaution for the second day of analysis (Dho 961,1 data collection), we monitored ^{155}Gd, ^{157}Gd and ^{160}Gd isotopes, and we report the data from ^{160}Gd isotope for Gd elemental abundance in the Dho 961 sample.

U-Pb methodology. Sample Dho 961,1 was cleaned using five minutes ultrasonic bath in ethanol followed by five minutes ultrasonic bath in distilled water and gold coated. We used a Cameca ims1280 ion probe housed by the Museum of Natural History in Stockholm to determine U-Pb isotopic compositions phosphates. A ~2 nA O_2^- primary beam with 23 kV energy (-13 kV primary, 10 kV secondary) was used in aperture illumination (Köhler) mode to produce a ~7 µm, flat-bottomed analytical pit. In order to minimise possible surface contamination (most likely from the sample polishing and cleaning process as well as the gold coating) each analysis was pre-sputtered for 30 s using a 20 × 20 µm raster. The mass spectrometer was operated at a mass resolution of 4860 (M/ΔM), sufficient to resolve Pb from molecular interferences using a 75 µm entrance slit and 250 µm exit slit. Secondary ions with 10 kV energy were centred in the field aperture and detected using four ion counting electron multipliers, with $^{204}Pb^+$, $^{206}Pb^+$, $^{207}Pb^+$ and $^{208}Pb^+$ isotopes detected simultaneously, followed by stepping of the magnet field through a background (also measured simultaneously, at a mass station 0.3 a.m.u. higher than the Pb peaks), and $^{238}U^+$, $^{232}Th^{16}O^+$ and $^{238}U^{16}O^+$ that were measured sequentially in the multicollector EM positioned on the axis. The electron multipliers were tuned for optimum yield and lowest noise using the pulse height analysis routine of the Cameca CIPS

software which gave relative yields within ±5% of unity. Actual relative detector yields were monitored using replicate analyses of a glass made from the BCR-2G basalt reference material assuming the Pb isotope composition determined by Woodhead and Hergt (2000).

U/Pb ratios in the phosphates have been corrected using a piece of apatite crystal with an age of 2058 Ma and U concentration of 67 ppm as characterised by ID-TIMS (see Nemchin et al., 2009 for details). Data for this reference material fall on parallel trends with the slopes of 1.80 in the Ln(^{206}Pb/^{238}U) vs. Ln(^{238}U^{16}O/^{238}U) coordinates. This slope was used for the correction of U-Pb data in the lunar apatites that followed the protocol utilised for the correction of ion probe zircon analyses (Williams, 1998).

3. Petrography and Mineral Chemistry Results

3.1. *Dhofar 925:* The Dho 925,1 section is ~7 × 6 mm and is a clast-bearing dark-grey glassy impact melt breccia (Fig. 1a and EA1.1). Clasts range from small (<10 µm) mineral and glass fragments up to 2 mm lithics, including magnesian granulites, basalts (quenched variolitic and ophitic texture), crystalline impact melt breccias and clast-bearing glassy melt breccias, lithic breccias, and rare Si-K-feldspar assemblages. The sample is cross-cut with fractures that are filled with terrestrially deposited minerals (picked out as yellow coloured veins in Fig. 1a).

3.2. *Dho 961:* The Dho 961,1 section is ~22 × 12 mm. Texturally, it is a lithic polymict breccia (we note that sometimes this texture is referred to in the lunar literature as a fragmental breccia). A few impact melt spherules have been previously reported by Zeigler et al. (2010a) in the Dho 961 sample, suggesting that it might contain a regolith component, or itself be a regolith breccia. However, we did

not located any spherules in our sample and cannot confirm this observation. A 300 µm wide melt vein is present in one corner of the section (Fig. 1b). The sample also contains a notably large (~4 × 2 mm) dark-grey, clast-bearing glassy impact melt breccia clast (top left in Fig. 1b) and a large (~5 × 3 mm) cream-coloured crystalline impact melt breccia clast (Fig. 1b and EA1.2). Dho 961,1 contains a more diverse textural range of clasts than seen in Dho 925,1. For example, coarse-grained granulite fragments are typically more mafic (20 to 60 % mafic phases), and basalts are more texturally diverse (variolitic, subophitic, ophitic textures). The sample is cross cut with fractures that are infilled with calcium carbonate (indicated as yellow veins in Fig. 1b) and some magnesian-phyllosilicates that originate from terrestrial alteration.

3.3. *SaU 449*: The SaU 449,3 section (~18 × 16 mm) is also a lithic (fragmental) polymict breccia with some large clasts up to 5 mm in size (Fig. 1c and EA1.3). The largest two clasts are a relatively coarse-grained (minerals up to 1.5 mm in size) granulite (~2.5 × 4 mm at top left in Fig. 1c) and a lithic monomict breccia (~4.5 × 2 mm at left centre in Fig. 1c). Other clast types include mineral fragments, glassy matrix impact melt breccias, crystalline impact melt breccias, monomict lithic breccias, granulites, basalts, rare granitic evolved clasts, and three large (100-600 µm) fragments of Fe-metal that have been partly oxidised, presumably by terrestrial alteration (red colour in Fig. 1c). Several large fractures run across the sample and are filled with terrestrial alteration minerals including calcium carbonate, oxidised iron products, calcium sulphate, and magnesian-phyllosilicates (green vein infill in Fig. 1c). This phyllosilicate is more common in SaU 449,3 than the Dho 925,1 or 961,1 sections.

3.4. *Lunar origin.* The stones are classified as lunar because their rock components and mineral compositions are similar to those in Apollo samples and other lunar meteorites (Russell et al., 2004, 2005; Connolly et al., 2007) and they host FeNi metal, indicative of origin from a low oxygen

fugacity planetary reservoir (Papike et al. 1991). FeO/MnO ratios of olivine (87±19 2stdev) and pyroxene (52±14 2stdev) in our sections are consistent with the lunar sample mafic mineral FeO/MnO trends (Fig. 2), supporting a lunar origin for the Dho 925, 961 and SaU 449 material we have studied.

3.5. *Clast description.* There are many lithic clast types ranging from 5 mm down to those scales that are indistinguishable from the sample matrix. Individual clast types and their petrography and mineral chemistry are described below and in supplementary files EA1.4 to EA1.9 and in Table EA2.

3.5.1. *Basalts.* Crystalline basaltic clasts are found in all three samples and have also been described by Deminova et al. (2005) and Zeigler et al. (2010a, 2013) in other portions of the stones. The fragments have typical mare basalt textures, ranging from plumose, to phenocrystic, ophitic, and subophitic (Fig. 3a-d, EA1.4), indicating a range of crystallisation histories.

Compositionally, the basalt fragments are dominated by zoned pigeonite ranging from $En_{0-68}Fs_{25-78}Wo_{6-31}$ (Figs. 4b and 4c). These compositions are similar to Apollo 17 very low-Ti (VLT) mare basalts (Papike and Vaniman, 1978). The predominance of low-Ca pyroxene in these type of basalts suggest plagioclase crystallisation occurred prior to pyroxene or co-precipitation of plagioclase and pyroxene. The pyroxene Ti# (atomic Ti/[Ti+Cr]) correlate positively with Fe# (atomic Fe/[Fe+Mg]) in two main trends (Figs. 5a and 5b) suggesting at least two sources for the parent magmas. The flatter part of the Al/Ti vs. Fe# trend (Fe# >0.4; Figs. 5c and 5d) can be explained by the co-crystallisation of pyroxene and plagioclase without ilmenite being a cotectic phase. These are similar to products of VLT magmas sampled by Apollo VLT basalts and lunar meteorites (Arai et al., 1996; Robinson et al., 2012). Two fragments have pyroxene that are Ti-richer (e.g., clasts Dho_961_1_Clast 15 and

Dho_961_1_Clast 71: Fig. 5b) and that are within the range of low-Ti Apollo and lunar meteorite basalt pyroxene (Arai et al., 1996; Robinson et al., 2012).

Plagioclase in basalt clasts are K-poor with compositions of $An_{94-98}Ab_{2-6}Or_{0-1}$ (Fig. 6), except one clast that has more evolved plagioclase crystals with $An_{88-94}Ab_{6-12}Or_{0-1}$ (SaU_449_3_Clast_81: EA1.4h), providing additional support that the majority of basaltic clasts in the Dhofar group are derived from at least two types of parental magmas. High-An# plagioclase in basalts (e.g., An# >96) have only been previously recorded in Luna 24 VLT mare basalts (Papike et al., 1991) and some lunar meteorite VLT basalts (e.g., Treiman and Drake, 1983; Joy et al., 2010b). Olivine occurs rarely in basalt clasts and where present (i.e., EA1.4b and EA1.4h) are Fo_{70-44} composition (Fig. 4a). These grains have >0.28 wt% CaO consistent with olivine in mare basalts (Ryder et al., 1997). The two olivine-bearing basalt fragments also host spinel with Mg-aluminous-chromite compositions (Fig. 7). Late-stage Si-phases occur in several of the basalt clasts (see EA1.4c, EA1.4h and EA1.4i for the darkest grey mineral phases) and, in the case of some clasts, this Si-rich phase has been replaced by terrestrial phyllosilicate. Other opaque phases include ilmenite, Cr-spinel and rarely small grains of FeNi metal (which are too small to collect stoichiometric EMP analyses).

3.5.2. *Coarse-grained granulites.* All three stones contain clasts that are coarsely grained with mineral phases are up to 1 mm (Figs. 3e-h and EA1.5) and a poikiloblastic texture. We term these clasts granulites as they have rounded to sub-rounded to sub-angular mineral boundaries, similar to meta-igneous clasts seen in granulitic lunar meteorite NWA 3136 (see Figure 4 of Hudgins et al., 2011). This poikiloblastic lithology is present as some of the largest clasts in the Dhofar group (e.g., a ~2 × 1 mm fragment in Dho 925,1 in ~1.5 × 1 mm fragment in Dho 961,1, and a ~2.5 × 4 mm fragment in SaU 449,1), suggesting it was a dominant precursor rock in the Dhofar group source region. Several other granulite clasts are also present that have smaller mineral fragments (<200 µm).

The coarse-grained granulites are typically dominated by granular An_{92-98} (Fig. 6) plagioclase crystals (rounded to sub-angular grain boundaries) that are ~20 µm to ~1 mm in size. Eighteen of these grains have positive Eu-anomalies (Eu/Eu* 1.2 to 2.3, where Eu/Eu* = Eu/√[Sm+Gd] normalised to the CI chondrite values of Anders and Grevesse, 1989), and trivalent REE at 0.1 to 9 × chondrite values (Fig. 8, Table EA2.1). REE abundances are intermediate to plagioclase from the lunar ferroan anorthosite suite and the Mg-Suite and high-Alkali Suite (Figs. 8c and 8e-g).

Pyroxene grains interstitially infill the spaces between plagioclase grains. Many of these pyroxene are not zoned (i.e., they are compositionally equilibrated) with distinct low-Ca ($En_{58-77}Fs_{17-28}$,Wo_{3-53}) and high-Ca ($En_{45-57}Fs_{8-17}Wo_{29-44}$) grains (Fig. 4b). Comparison with lunar temperature two-pyroxene isotherm (Lindsley and Andersen, 1983) of adjacent high and low-Ca grains suggests a range of equilibration temperatures from ~1050 to ~900°C. A low-Ca equilibrated pyroxene (Wo_8) has trivalent REE with values of 2 to 9 × chondrite values with a negative Eu-anomaly (Eu/Eu* 0.03), and two high-Ca (Wo_{32-40}) pyroxene have trivalent REE with values of 10 to 75 × chondrite values with negative Eu-anomalies (Eu/Eu* 0.03-0.06: Fig. 9c, Table EA2.2). Some clasts have pyroxene that exhibit lamellar intergrowths between the Ca-rich and Ca-poor varieties (n.b., some of the analyses shown in Figure 4b include mixing of these co-existing phases).

Rounded olivine grains (Fo_{66-82}; typically with a variation of ~$Fo_{<5}$ per clast) are included within some of the pyroxene, and also less commonly within the plagioclase grains (EA1.5). At Fo_{72-78}, concentrations of Mn, Cr, Co and Ni (Fig. 10, Table EA2.3) are most similar to Apollo igneous plutonic Mg-Suite gabbronorites (Shearer and Papike, 2005), and CaO concentrations are most similar to Apollo granulites and Mg-suite intrusive lithologies, rather than shallow level magmas/lavas (Ryder et al., 1997).

The proportion of mafic to felsic phases varies from fragment to fragment (EA1.5), though plagioclase is typically >60% (by area estimated using imaging software), and mafics typically <40% (by area). Thus, most of the fragments are anorthositic gabbronorites. The fragments have mineral compositions (plagioclase An_{92-98}, olivine Fo_{66-82}) that are intermediate to ferroan anorthosite and Mg-Suite samples (Warren, 1993), and most similar to magnesian anorthosites and magnesian granulites (Lindstrom and Lindstrom, 1986; Takada et al., 2006; Treiman et al., 2010; Gross et al., 2014).

The texture (Figs. 3e-h and EA1.5) and mineral chemistry of these clasts suggests that they are mildly to moderately annealed meta-igneous rocks (i.e., they preserve some little to some degree of slowly-cooled intrusive rock textures, although their mineral chemistry has often been equilibrated). Zeigler et al. (2010a) describe these types of clasts in the Dhofar stones as granulite/plutonic material. However, we do not consider this clast group to be derived from igneous intrusive sources as we have found FeNi metal present in some of the clasts (0.49 to 0.55 wt% Co, 5 to 6 wt% Ni: Table EA2.7), which is more similar to the compositional range of meteoritic metal rather than indigenous lunar metal (Papike et al., 1991). As their precursor rocks contained a meteoritic metal component it is most likely related to an impact melt sheet, which underwent mild-moderate post-emplacement annealing to enhance chemical equilibration and mineral grain rounding.

3.5.3. Lithic (fragmental) breccias: Numerous types of lithic breccias are included in the three samples (EA1.8) that represent material that was cataclastically broken apart and aggregated back together. A monomict and a polymict lithic breccia are both notably large (> 500 µm) clast types in SaU 449,3 (Fig. 1c). The monomict breccia (EA1.8k) appears to be a sheared, brecciated granulite with similar mineral compositions to clasts discussed in section 3.5.2. The polymict breccia (EA1.8l) includes several different granulite clast types, and individual mineral fragments including two

ferroan exsolved pyroxenes >100 µm, fragments of FeNi metal and several magnesian mafic phases (olivine Fo_{74-81}, low-Ca pyroxene $En_{78-90}Fs_{7-18}Wo_{3-9}$) akin to material from the Apollo Mg-Suite. These clasts are supported by a groundmass of smaller grained mineral fragments.

3.5.4. Crystalline impact melt breccias. These clasts are texturally similar to crystalline impact melt rocks that formed from large-degree impact melting events (Taylor et al., 1991). They occasionally contain clasts of mineral fragments inherited from the disrupted impact target rock(s). The crystalline melt breccias also often contain metallic components, probably originally sourced from the projectile that caused the impact event, but that were probably modified through chemical exchange with the lunar melt sheet (Wanke et al., 1978; Norman and Nemchin, 2014).

Several large clast-poor fragments with ophitic to subophitic textures are present in all three meteorites (EA1.6). The largest of these in Dho 961 (see details in EA1.7) is composed of zoned pigeonitic pyroxene ($En_{41-72}Fs_{17-36}Wo_{5-38}$, 51% area) enclosing plagioclase crystals (An_{86-97}, 44% by area), with late-stage assemblages composed of aggregates of plagioclase with olivine grains (Fo_{51-67}, 2.5 %), ilmenite (1 %), K-feldspar (~0.6 %), and phosphate and Zr-bearing phases (~0.7%), with accessory silica (often terrestrially replaced by magnesian phyllosilicates), troilite and FeNi metal. This metal has 0.5 to 0.6 wt% Co and 5 wt% Ni, more similar to that reported in meteorites than indigenous lunar rocks. By phase area modal recombination the clast is a gabbronorite, and has bulk ~15.9 wt% Al_2O_3, 9.8 wt% FeO, 11.7 wt% MgO and Mg#68 (Table EA2.4), which is similar to Apollo mafic impact melts (Fig. 12a). Several other clasts in the Dhofar group have similar textures (Fig. 3i-j) and mineral compositions (Figs. 4 to 11) to this clast. This material is likely the dominant carrier of ITE-rich phases as they are K-feldspar, phosphate and Zr-phase bearing. We collectively term this type of crystalline impact melt breccia 'type 1'. Plagioclase (An_{94-95}) phases have trace element concentrations intermediate to Mg-Suite and FAN rocks are unlike mare basalts (Figs. 8e-f,

Table EA2.1). Two low-Ca pyroxene grains (Wo_{7-10}) and one high-Ca pyroxene grain (Wo_{32}) have REE concentrations and element ratios (Fig. 9b, Table EA2.2) more akin to lunar highland lithologies than mare basalts. Olivine (Fo_{51-52}) grains have >1.5 CaO wt%, similar to shallow level / erupted lunar lavas (Ryder et al., 1997). Olivine in one of the clasts (EA1.6b and EA1.6c) includes small blebs of troilite that appear to be replacing the olivine phase, similar to that observed in some components of lunar breccias (Shearer et al., 2012 and refs. therein). Sulphide-free olivine grains have trace element concentrations (Table EA2.3) similar to those reported in mare basalts clasts, although the impact melt olivine have comparatively lower Co abundances (Figs. 10b-10d).

An ITE-rich mafic gabbronoritic impact melt also dominates the clast population of the Dho 961 sample described by Jolliff et al. (2008). Our type 1 crystalline impact melt breccia differs from the Jolliff et al. (2008) impact melt in that it has lower bulk FeO (~9.8 wt% compared with ~12.5 to 14 wt% FeO: Fig. 12a) and contains no olivine clasts. Where olivine is present in our sample as groundmass crystals it is more ferroan and has less Ni (Fo_{51-67}, 33-60 ppm Ni) compared with the Jolliff et al. (2008) olivine (clasts Fo_{55-80}, matrix olivine Fo_{50-73}, ~200 ppm Ni).

Other types of crystalline impact melt breccias are also seen in our samples. One example we term crystalline impact melt breccia 'type 2' (Dho_961_1_Clast_70, Fig. 3k) is has a similar texture to the late-stage assemblages seen in the type 1 impact melt, but is composed solely of fayalitic olivine (Fo_{46-48}), anorthositic plagioclase (An_{94-95}), with minor K-feldspar, silica and ilmenite, and accessory merrillite and apatite.

A finer grained feldspathic poikilitic type we term crystalline impact melt breccia 'type 3' (Fig. 3l and close up in EA1.6f, and see similar texture types in EA1.6g and EA1.6h) has anorthitic plagioclase

(An_{94-97}) where larger chadacryst grains (>40 µm) with subrounded mineral boundaries are inherited clasts, and the finer groundmass plagioclase likely crystallised from the impact melt itself. Also present are Ca-rich ($En_{30-49}Fs_{19-45}Wo_{25-41}$) and Ca-poor ($En_{67-70}Fs_{22-26}Wo_{7-10}$) oikocryst pyroxene and accessory troilite and ilmenite. Its bulk composition (Table EA2.4, Fig. 12) has greater REE concentrations than the clast-bearing glassy impact melt breccias (see section 3.5.6), but lower REE concentrations and different element ratios than Apollo mafic impact melt groups (Fig. 12).

Fluoroapatite (Fig. 11, Table EA2.5) and merrillite (EA1.6b and EA1.6c) are accessory phases in many of the crystalline impact melt breccia clasts (e.g., EA1.7d). Type 1 impact melt merrillite crystals are more magnesian (Mg#83-90, 1.2 to 2.3 wt% Y_2O_3, 0.5 to 1.1 wt% La_2O_3: Table EA2.6) than type 2 (Mg#73-83, 1.3 to 1.4 wt% Y_2O_3, 0.5 to 0.6 wt% La_2O_3: Table EA2.6), suggesting that they were sourced from different parent melts.

3.5.5. Glassy impact melt breccias. These rocks are clast-bearing and are composed of mineral and lithic clasts of a range of sizes (e.g., micron to mm size) enclosed in a glassy matrix (see clast classification of Taylor et al., 1991, and references therein). This clast type presents variable textures in all three samples. Dho 925,1 is in itself a clast-bearing glassy impact melt breccia (Figs. 1a and 3m). Most glassy impact melt breccia clasts (Figs. 3m-o and EA1.8a-j) have 20 to 25 wt% bulk Al_2O_3 and 5 to 9 wt% FeO (Fig. 12a and Table EA2.4), with compositions similar to the Dho 925 bulk rock (Table 1) and straddling the boundary between lunar feldspathic rocks and rocks of intermediate-Fe composition. These clasts are alkali poor (<0.6 wt% Na_2O + K_2O; Table EA2.4) and have relatively low concentrations of incompatible trace elements (trivalent REE ×4 to ×23 chondrite abundances) compared with Apollo mafic impact melt groups (trivalent REE ×40 to ×420 chondrite abundances: Fig. 12e, Table EA2.4).

There are some clast-bearing melt breccias that are different from the types discussed above. For example, one variety (EA1.8e) is highly magnesian, with sub-rounded >100 µm fragments of Ca-poor pyroxene ($En_{83-85}Fs_{12-13}Wo_3$), Mg-rich olivine (Fo_{82-86}) and plagioclase (An_{93-96}) in a microcrystalline groundmass. The target rocks melted to form this sample must have been magnesian lithologies that were also included into some of the lithic breccias described in section 3.5.4.

3.5.6. Granitic (evolved) clasts. There are rare occurrences of siliceous and K-feldspar bearing lithic clasts in the Dhofar group (Fig. 3o and EA1.9). The largest fragment (~200 × 150 µm: EA1.9a) contains silica with K-feldspar ($An_{17-23}Al_{3-7}Or_{71-80}$; Fig. 6). These fragments could represent material from High Alkali Suite intrusive magmatics, or late-stage precipitates of erupted basalts.

3.5.7. Melt vein. The top left portion of the Dho 961,1 section (~1 % of the total section area: Fig. 1b) contains a glassy vein containing <10% of small (<50 µm) mineral fragments and a single (100 × 150 µm) clast of 'type 1' impact melt breccia. The bulk composition of the vein itself (not including the clast) is listed in Table 1: it is more similar to the bulk composition of Dho 925 (Korotev et al., 2009) and the averaged Dhofar group (Demindova et al., 2007; Korotev, 2012), than to the bulk composition of Dho 961 (Korotev et al., 2009) (Fig. 12f). Korotev et al. (2009) observed a melt vein in a larger slab of Dho 961 that also had a composition akin to the Dho 925 bulk rock.

3.5.8. Mineral fragments. Most silicate minerals analysed in the meteorite matrix have similar compositions to those in clasts (Figs. 4-9) and it is likely that they are derived from the same types of comminuted precursor rocks. A few exotic mineral types have also been sampled. For example, two anorthitic plagioclase fragments (An_{97}) in Dho 925 occur with LREE concentrations lower than

plagioclase from the Apollo ferroan anorthosite suite (Fig. 8a, Table EA2.1) and the magnesian anorthosites sampled by lunar meteorite Dho 489 (Takeda et al., 2006).

Fragments of FeNi metal are distributed throughout the matrix of all three samples. In terms of Co and Ni concentrations all are similar to metal in asteroid-sourced meteorites rather than indigenous lunar rocks (Papike et al., 1991). Some also have elevated concentrations of phosphorous (> 0.04 wt% P; Table EA2.7), suggesting that these metals may have been re-crystallised in KREEP-bearing lunar impact melts or that the delivered meteoritic metal was P-bearing (see Nazarov et al., 1981 and references therein). SaU 449,3 in particular has three fragments of FeNi metal >100 µm in size (bright red coloured phases in Fig. 1c and EA1.3) that have been partially replaced by Fe-oxides and have textures that suggest alteration and oxidation by terrestrial fluids.

4. Chronology Results

Pb-Pb and U-Pb isotope data was collected from 38 phosphate grains (apatites and merrillites) in Dho 961,1 (Table 2, EA1.10 and EA1.11). Some of these were measured in discrete grains with no to little petrographic context (i.e., they occur as mineral fragments in the Dho 961,1 matrix: EA1.11), and others were in crystalline impact melt breccias (EA1.10 and EA1.11) and an equilibrated granulite (EA1.11). The U-Pb data (Table 2), both uncorrected for initial Pb, and when corrected using the Stacey and Kramers (1975) model composition (assuming that most of the non-radiogenic Pb is the result of surface contamination introduced during sample preparation), scatter widely when plotted on the concordia diagram (Figs. 13a and 13b), with both normally and reversely discordant analyses visible in the data set. This scatter is most likely a result of analysis of relatively small grains in the thin sections, where analytical artefacts associated with relief present in the samples that are difficult to control and correct for. While U-Pb ratio can be very sensitive to the changing analytical conditions

and surface imperfections, Pb istope ratios are not affected by these changes. Consequently $^{207}Pb/^{206}Pb$ data can be used to constrain ages of phosphates, even though this restricts our ability to distinguish grains with partially reset U-Pb systems. In addition, taking into account the significant spread of $^{207}Pb/^{206}Pb$ ages indicating variable Pb loss from the phosphate grains all ages should be interpreted as limits of true ages, i.e., oldest ages in a population can be interpreted as the minimum age of this population, while youngest ages represent maximum age for the thermal overprint(s) that caused Pb loss.

Crystalline impact melt breccia type 1. Phosphates in three of these clasts have a continuous spread of $^{207}Pb/^{206}Pb$ ages between 4032±18 Ma (2σ) and 3839±50 Ma (2σ) (Table 2, Fig. 13c). Together these give a weighted mean age of 3945±11 Ma (23 datapoints 95% conf. to MSWD 32), however, this is not significant as there is so much variability from grain to grain. The youngest age likely represents the maximum age of a resetting event. The oldest date of 4032±18 Ma places a minimum time limit for crystallisation of the impact melt type 1, but could represent either the timing of impact melt crystallisation or the older recorded resetting event for this clast.

Crystalline impact melt breccia type 2. Six phosphate grains were examined and one grain has an old age of 4258±15 Ma (Table 2, Fig. 13c, EA1.11), which could represent the minimum formation age for the clast, or could represent an inherited relict grain. Five of the grains have younger ages that range from 3994±4 Ma to 3877±16 Ma (Fig. 13c), which is within the same spread of ages measured in phosphates from the crystalline impact melt breccia type 1, suggesting that the two clast types have similar time-temperature histories.

These phosphate ages in the type 1 and 2 impact melt breccias were measured within a spatial scale of just a few hundred micron to mm in each clast. Nemchin et al. (2009) discuss various models of U-Pb

and Pb resetting in lunar impact melt breccia apatites and these studies imply that as different regions of the Dho 961 crystalline impact melt cooled at different rates they were likely not trapped under thick thermally insulating lid(s). This indicates that isotopic closure or later resetting of these impact melts occurred within a few meters to tens of meters of the Moon's surface, rather than in a thick impact melt pond.

Coarse-grained granulite. One of the equilibrated granulite clasts contains a P-bearing (0.01-0.03 wt% P: Table EA2.7) FeNi metal fragment hosting a sub-rounded OH-bearing chlorapatite grain (Fig. 11, EA1.5d and e, EA1.11, Table EA2.5). It is possible that this apatite formed from the metal breaking down under oxidising slow cooling conditions <750°C (as discussed by Friel and Goldstein, 1977). This apatite yielded a Pb-Pb age of 4088±15 Ma (Table 2, Fig. 13c), implying that the apatite formed at, or before, this time.

Matrix fragments. We analysed eight unique phosphate fragments within the matrix of the Dho 961,1 lithic breccia (EA1.11). These include two merrillites and six apatites (some of which are OH-bearing and others not: Fig. 11, Table EA2.5). These have a range of older Pb-Pb ages from 4343±10 Ma to 4009±10 Ma (Table 2, Fig. 13c), which are generally older than phosphates in the impact melt clasts (Fig. 13c), suggesting that they may have been derived from different type(s) of parent lithologies.

Assembly of the Dho 961 breccia. The youngest dates recorded in Dho 961,1 are found in type 1 and type 2 crystalline impact melt breccia. The weighted mean of the four youngest ages in the two clasts is 3890±18 Ma (2σ), and this provides an upper limit age of the time that the Dho 961 breccia was possibly lithified.

5. Remote Sensing Data and Potential Source Regions

It may be possible to constrain the source region of the meteorites using remote sensing geochemical datasets (see approach of Jolliff et al., 2009; Corrigan et al., 2009). We searched the Lunar Prospector gamma-ray spectrometer datasets using a method similar to Joy et al. (2010a, 2011a) and Mercer et al. (2013), assuming that meteorites were derived from a compositionally homogeneous terrane exposed on the scale of individual pixels. We used the bulk FeO and Th and TiO_2 composition (Table 1) of both the Dho 961 meteorite (data from Korotev et al., 2009) and the average Dho 925/960/961 group composition (data from Demidova et al., 2007) as two compositional input parameters. Using these values ± the maximum Lunar Prospector dataset element errors as search thresholds, we searched the high spatial resolution 2° per pixel dataset (Prettyman et al., 2006 calibration) to locate regions of interest (regolith) with similar compositions (Fig. 14; see caption for further details). We only used Th, FeO and TiO_2 in our search analysis as other elements in the Lunar Prospector 2° per pixel dataset have large statistical errors, and previous authors have identified some inconsistencies between the Si and Mg Lunar Prospector data and returned lunar sample composition (e.g., Wohler et al., 2010). As such we do not use these other key rock forming elements in our analysis.

Our results (Fig. 14) show that Dhofar 961 is compositionally similar to regoliths in the north western region of the South Pole-Aitken basin (akin to those area identified in Jolliff et al., 2009), and also to regoliths around the western, southern and eastern periphery of the Procellarum KREEP Terrane; notably in regions to the north of Lacus Somniorum and south of the eastern region of Mare Frigoris (Figs. 14a-d). However, we note that both Korotev et al. (2009) and our study observed that impact melt veins cross-cutting the Dho 961 sample have compositions that are more like the Dho 925, 960 and SaU 449 stones than the Dho 961 stone (Table 1). Therefore, it is possible the Dho 961 bulk composition reported by Korotev et al. (2009) is anomalous as it contains a higher component of mafic impact melt clasts than other portions of the meteorite (see also discussion in Korotev, 2012) and is, thus, not best suited for this exercise.

The mean bulk rock Dho 925/960/961 composition (Demidova et al., 2007) matches a wide range of lunar regoliths with similar Th, FeO and TiO_2 compositions (Figs. 14e-h). This search suggests that the Dhofar group may have been sourced from the eastern limb of the Moon, north-eastern regions of the South Pole-Aitken basin (e.g., Apollo basin area), south-polar regions to the south of SPA (between Schrödinger and Zeeman basins), cryptomaria in the Schickard-Schiller region, and regions around Mare Moscoviense and Mare Australe.

6. Discussion

6.1. Are the meteorites from the South Pole-Aitken basin?

It has been argued (see Zeigler et al., 2013 and references therein) that the Dhofar group originated from a moderately Fe- and Th-rich region of the Moon that is not within the Procellarum KREEP Terrane. An alternative source region has been suggested to be the South Pole-Aitken basin (Jolliff et al., 2007, 2008, 2009; Korotev et al., 2007, 2009, 2010; Zeigler et al., 2010a, 2010b; 2013). We weigh up the evidence for a PKT or an SPA origin further here.

6.1.1. Evidence from remote sensing datasets: We note that the meteorites in the Dhofar group are not formally classified as regolith breccias, and so their compositions should be used with caution when comparing with present day surficial regoliths. Although our results (Fig. 14) show that the meteorites are not compositionally similar in terms of Th, FeO and TiO_2 to central regions of the PKT, Dho 961 is similar in composition to some northern, southern and western regoliths within SPA (Fig. 14a-d).

These results are consistent with those of Jolliff et al. (2008, 2009) and Zeigler et al. (2013), and so together this could support an origin in SPA. However, (i) there are regoliths around the periphery of the PKT, notably near the south and east of Mare Frigoris, that are also similar in composition to Dho 961 (Fig. 14a-d); (ii) the Dho 961 bulk composition is anomalous compared with its grouped Dho 925 / 960 and SaU 449 bulk stones (see section 5, and Korotev, 2012). Therefore, regolith searches for this sample alone might not be significant (Fig. 14a-d) and, when similar searches are performed using the average composition of the Dho 925/960/961 meteorites (Fig. 14e-h), these results are less consistent with SPA being the only possible areas of the Moon that are compositionally similar to the Dhofar meteorite group (see section 5).

6.1.2. Evidence from bulk rock chemistry and impact melts: The South Pole-Aitken basin is thought to have a large impact melt sheet, which possibly differentiated into cumulate horizons (Morrison, 1998; Vaughan and Head, 2014; Hurwitz and Kring, 2014). These differentiated products are thought to be dominated by a surficial noritic signature (Isaacson and Pieters, 2009; Nakamura et al., 2009) associated with surficial Th-rich lithologies (Hagerty et al., 2011), and with underlying gabbroic and olivine-rich cumulate piles (Hurwitz and Kring, 2014). If the Dhofar group of meteorites were sourced from SPA, then impact melt breccia clasts and also possibly the coarse-grained granulite clasts in the Dhofar group stones could have been derived from the SPA basin-forming event, or secondary events into such a melt sheet.

Impact melts sampled by the Apollo 12, 14, 15, 16 and 17 missions (Jolliff, 1998; Korotev et al., 2004) represent melted products of basin(s) within the Procellarum KREEP Terrane. These Apollo impact melt types represent mixing between feldspathic rocks and varieties of upper and lower crust KREEP-bearing rocks (i.e., High Alkali Suite and Mg-Suite samples and KREEP basalts: see Figs.

12a-d). Meteorites that were sourced from the PKT like SaU 169 (Gnos et al., 2004; Liu et al., 2012) and NWA 4472 (Joy et al., 2011a) have impact melts with similar compositions to the Apollo melt groups. Korotev et al. (2009) argue that the bulk composition of Dho 961 has key compositional differences from Apollo impact melt samples, and Jolliff et al. (2008) reported olivine-bearing mafic gabbronoritic impact melts in Dho 961 that have unusual compositions (e.g., high bulk P/REE, P/K, Sc/Cr ratios) compared with KREEP-rich mafic Apollo melt breccias. They used these observations as evidence that the meteorite has a non-PKT origin. We are unable to petrographically investigate or test the chemistry or age of this type of mafic impact melt as we have not located fragments of this material in our samples (see section 3.5.5). However, our type 3 feldspathic crystalline impact melt breccia clasts and the clast-bearing glassy impact melt breccias are also different in composition compared with the Apollo mafic impact melts. For example, our glassy breccias are less mafic and have different Sc/Sm and Eu/Sm ratios (Fig. 12). Compositionally, they appear to reflect mixing between a feldspathic end-member and a ferroan high Sc/Sm rock type with minor KREEP. These melt breccias do not appear to have melted an Apollo-like Mg-Suite or High Alkali-Suite KREEP-rich rock component, and are compositionally more similar to lunar meteorite feldspathic and intermediate-Fe impact melts (see Joy et al., 2010a and refs. therein) (Figs. 12a-d). These differences from Apollo-like lithologies are evidence that the samples do not have a central PKT-origin, and were sourced from another moderately Th-rich, Fe-rich region of the Moon, which may or may not be the SPA basin.

6.1.3. Evidence from impact melt breccia chemistry age: The South Pole-Aitken basin formation age is still unknown from direct geological sampling. From crater size-distribution models it is inferred to be >4 Ga, with ages ranging from ~4.26 Ga (Hiesinger et al., 2012a, 2012b) to >4.33 Ga (deduced from the oldest regions of the basin floor: Morbidelli et al., 2012 after Marchi et al., 2012). However, these ages are not well constrained due to the lack of calibration of the >4 Ga part of the lunar crating

curve (Stöffler et al., 2006). Younger, overprinted, impact basins within SPA have model crater count ages between 3.91 and ~4.09 Ga (Hiesinger et al., 2012a, 2012b).

The 4.35-4 Ga (Fig. 13c) ages of phosphates in the Dho 961,1 matrix and a coarse-grained granulite are within the range of the proposed ages of the SPA basin and younger basins therein. The type 2 impact melt breccia in Dho 961 has an older Pb-Pb age of 4.26 Ga, which is similar to the Hiesinger et al. (2012a, 2012b) model age of SPA (~4.26 Ga), but not Morbidelli et al. (2012) model age (>4.33 Ga). The type 1 and type 2 crystalline impact melt breccias in Dho 961 both have a range of younger phosphate Pb-Pb ages from 4 to 3.9 Ga (Fig. 13c, Table 2). These ages are younger than the proposed age of SPA (>4.26 Ga), but consistent to when many Apollo impact melts were formed and isotopically reset (Bogard, 1995). These 4 to 3.85 Ga ages in Apollo samples are attributed to crater and/or basin-forming events at the latter end of basin-formation (i.e., Imbrium and Serenitatis basin-forming events: Bogard, 1995). They are also consistent with the model crater count ages of some of the younger basins within SPA (~4.09 and 3.91 Ga: Hiesinger et al., 2012a, 2012b). If any of the impact melt products in Dho 961 were formed by the South Pole-Aitken basin (>4.26 Ga), they have been reset by overprinting younger thermal events.

6.1.4. Evidence from mare basalt fragment chemistry: Mare basalts on the nearside of the Moon are variable in composition from VLT to low-Ti to high-Ti, with a wide range of model crater count ages (e.g., Wilhems et al., 1987). No high-Ti mare basalt clasts were found in the Dhofar group, so it is likely that the samples were not sourced from high-Ti regions of Mare Tranquillitatis, Oceanus Procellarum and some interior regions of Mare Moscoviense on the lunar farside. Many basalt clasts in the Dhofar group originate from VLT lavas, and basalts with similar compositional affinities dominate regions on the lunar nearside in Mare Frigoris and the northern regions of Mare Imbrium and many cryptomaria (buried ancient lava flows).

Mare basalts occur on the floor of the South Pole-Aitken basin, within superimposed basins and crater floors. These basalt ponds have Late Imbrian model crater count ages of 3.8 to 3.2 Ga (Yingst and Head, 1999 and refs. therein). These ages are younger than any ages recorded by rocks and mineral fragments of the Dhofar group (Fig. 13c, Table 2). Hagerty et al. (2011) found that five main SPA basin mare basalts are low-Ti to high-Ti types (~4-10 wt% TiO_2) (see also Yingst and Head, 1999), which is dissimilar from the dominant VLT basalt types found in the Dhofar group (Fig. 5b). This could argue against a SPA origin for the Dhofar group, although there may be older buried cryptomaria basalts within SPA with different compositions that are not currently exposed at the surface.

7. Conclusions

The Dho 925, 961 and SaU 449 samples are lithologically diverse (Fig. 1). The group represent products of impact cratering event(s), which affected several types of target rocks and mixed them together as an impact melt breccia. Collectively, the breccias contain fragments of at least 4 different mare basalt textural types from very low-Ti and low-Ti lavas; two main impact melt breccia types including crystalline and clast-bearing glassy, which both have several different sub-varieties; two main types of granulites include coarse grained clasts (typically >200 µm grain size), and finer grained types (typically <200 µm grain size); lithic (fragmental) breccias including both monomict and polymict fragments; and fragments of evolved lithologies that include silica and/or K-feldspar minerals. This gives a minimum of at least eleven distinct lithic clast types, some of which have many different types of textural variation. Similarity of clast chemistry and texture between the individual samples is strong evidence that the samples are grouped and originate from the same parent meteorite

that fragmented into separate stones during the meteorite's fall to Earth (Demidova et al. 2007; Korotev et al., 2009; Korotev, 2012).

Dhofar 961 records at least half a billion years of lunar history (Fig. 13c). Phosphate fragments found within the Dho 961,1 matrix have 4.34 to 4 Ga ages. These dates overlap with the ages of ancient KREEP-driven magmatic episodes and Pre-Nectarian (>3.92 Ga) basin-formation on the nearside of the Moon (Wilhelms et al., 1987; Stöffler et al., 2006; Nemchin et al., 2008, 2009). A phosphate in crystalline (type 2) impact melt breccia in Dho 961,1 has an old 4.26 Ga formation/reset age that is consistent with several impact resetting events in the Pre-Nectarian at ~4.2 Ga (Garrick-Bethell et al., 2008; Fischer-Gödde and Becker, 2012; Fernandes et al., 2013; Norman and Nemchin, 2014). There are also a series of younger 4 Ga to 3.89 Ga events that are recorded in Dho 961,1 crystalline impact melt breccias that are consistent with late basin-forming events like the formation of Imbrium (sample ages 3.85 to 3.92 Ga; Stöffler et al., 2006; Liu et al., 2012) and Schrödinger (crater count absolute model age ~3.92 Ga; Hiesinger et al., 2012b). The youngest reset ages in the Dho 961,1 sample are at ~3.89 Ga and they represent an upper limit for the time of formation. This possibly records the timing of the final impact, which mixed and consolidated several generations of precursor rocks into the Dhofar meteorite group, although further age dating of all the stones is required to test this hypothesis.

Evidence from our study and previous studies (Zeigler et al., 2013 and refs. therein), suggest that the meteorites were probably not sourced from directly within the nearside Procellarum KREEP Terrane. For example, the bulk chemistry of the Dhofar group does not match present day Procellarum KREEP Terrane regoliths (Fig. 14), and the bulk and rock components differ in chemistry from impact melts sampled by the Apollo missions (Fig. 12). The meteorite may have been sourced from the South Pole-Aitken basin (Jolliff et al., 2008). However, other source regions on the lunar surface cannot be ruled out (Fig. 14), especially as basalts within the Dhofar group are mostly VLT in composition, which

does not match well with low-Ti to intermediate-Ti types that outcrop within SPA (Yingst and Head, 1999; Hagerty et al., 2011). If the meteorites did originate from the South Pole-Aitken basin, then evidence of basin formation age may have been numerously overprinted by younger crater and basin forming events, especially at ~3.9 Ga (Fig. 13c), consistent with an observed increase in impact flux at the latter end of the lunar basin-forming epoch recorded by Apollo samples (Bogard, 2005) and also KREEP-rich lunar meteorites that originated from within the Procellarum KREEP Terrane (Gnos et al., 2004; Joy et al., 2011a; Liu et al., 2012).

Acknowledgements

This research was funded by NASA Lunar Science Institute contract NNA09DB33A, David A. Kring PI. KHJ acknowledges funding from the Leverhulme Trust, UK (grant 2011-569). We thank David Mann for sample preparation. We acknowledge the resources of Dr. Randy Korotev's Lunar Meteorite List, and NASA's Apollo and Lunar Meteorite sample compendium. We thank Drs. Axel Wittmann, Tomoko Arai and Romain Tartese for thoughtful reviews, and Dr. Marc Norman for his AE handling and suggestions which greatly improved this paper. This is LPI contribution number XXXX.

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Figure Captions

Figure 1. False colour montaged element maps of the (a) Dho 925,1, (b) Dho 961,1 and (c) SaU 449,3 thick sections. Colours shown in the electronic version represent distribution and qualitative concentration (i.e., brighter the colour, more of that element is present) of different elements where Al=white, Ca=yellow, Fe=red, Si=blue, Mg=green, Ti=pink and K=cyan. Typically anorthitic plagioclase appears white; apatite crystals and terrestrial $CaCO_3$ infill is yellow; Fe-rich metal and sulphides are bright red; silica and Si-rich glass is bright blue; K-rich feldspar is cyan; and olivine and pyroxene can be a range of colours from brown (Ca-richer) to green/ blue-green (Mg-richer) and to purple (Fe and Ti-richer) depending on composition.

Figure 2. Mn vs. Fe atoms per formula unit in (a) olivine, and (b) pyroxene in the Dho 925, 961 and SaU 449 samples. Data are compared with mafic phases in other lunar meteorites (MacAlpine Hills 88104/05: Joy et al., 2010a; Dar al Gani 400: Joy, 2007, Joy et al., 2010a; Meteorite Hills 01210: Joy, 2007, Joy et al., 2010a; North West Africa 4472: Joy et al., 2011a; Miller Range 07006: Joy et al., 2010b; Robinson et al., 2012; La Paz 02205 and pairs: Joy et al., 2006; Joy, 2007; and Miller Range 05035: Joy et al., 2008), where the Moon lines are linear fits to these data (olivine: Mn = [0.0114×Fe]-0.0003; pyroxene: Mn = [0.0116×Fe]-0.0038). Apollo mineral data are plotted for comparison (Takeda et al., 1975; Papike et al., 1998; Ryder et al., 1997; Jolliff et al., 1999; Shervais and McGee, 1998; Schnare et al 2008; Taylor et al., 2012). Planetary trend lines for Earth basalts, Mars (martian meteorites) and Vesta (HED-meteorites) are from Papike et al. (2009), and ordinary chondrite (OC) and carbonaceous chondrite (CO) trend lines are from Berlin et al. (2011).

Figure 3. Back-scattered electron (BSE) images of clasts and components in Dho 925,1, Dho 961,1 and SaU 449,3. (a) Phenocrystic basalt in Dho 925 (Dho_925_1_Clast_5). (b) Zoned basalt in Dho 961 with a large olivine grain hosting a melt inclusion (Dho_961_1_Clast_15). (c) Ophitic basalt in Dho 961 (Dho_961_1_Clast_39). (d) Zoned basalt clast in Dho 961,1 with variolitic plagioclase

(Dho_961_1_Clast_74). (e) Part view of a poikioblastic granulite, which is the largest fragment in Dho 925 (Dho_925_1_Clast_16). (f) Granulite clast (Dho_925_1_Clast_53). (g) Granulite with large area of FeNi metal including an apatite grain (Dho_961_1_Clast_84). (h) Granulite with high proportion of mafic phases (Dho_961_1_Clast_34). (i) Type 1 crystalline impact melt breccia clast in Dho 925,1 (Dho_925_1_Clast_13). (j) Type 1 crystalline impact melt breccia in SaU 449 (SaU_449_3_Clast_4). (k) Type 2 crystalline impact melt breccia (Dho_961_1_Clast_70). (l) Type 3 fine-grained poikilitic crystalline impact melt breccia (SaU_449_3_Clast_58). (m) Close-up texture of the Dhofar 925 sample matrix, which is a clast-rich (olivine, pyroxene and plagioclase mineral fragments) glassy impact melt breccia. (n) Clast-bearing impact melt breccia in Dho 925 where large plagioclase clasts sit in a groundmass of plagioclase laths in a mafic glass matrix (Dho_925_1_Clast_20). (o) Clast-bearing impact melt breccia in SaU 449 (Sau_449_3_Clast_85) with a Mg-rich Fo_{91} olivine fragment and clasts of plagioclase held in a fine-grained groundmass of plagioclase microcrysts and mafic glass. (p) Fragment of silica and K-feldspar intergrowth with an apatite grain present in the top left corner (SaU_449_3_Clast_94).

Figure 4. Dho 925,1, Dho 961,1 and SaU 449,3 mafic mineral phase compositions. (a) Olivine Fo composition (100 × atomic Mg/[Mg+Fe]). Pyroxene quadrilateral showing variation of pyroxene composition in (b) all clasts and (c) detailed in basaltic clasts only. IMB = impact melt breccia.

Figure 5. (a) Diagram of atomic Ti/Ti+Cr (Ti#) vs. Fe# (atomic Fe/Fe+Mg) for pyroxene varieties in main Dho 925,1, Dho 961,1 and SaU 449,1 clast types. (b) Details of Ti# versus Fe# of pyroxene for basalts clasts only. Low-Ti and very-low Ti (VLT) mare basalt crystallisation trends are indicated (after Arai et al., 1996; Robinson et al., 2012). (c) Fe# vs. cation ratio of Al/Ti in pyroxene in main clast types. (d) Same as (c) but detailing pyroxene compositions for basalts clasts only.

Figure 6. Feldspar composition in different clast types in the Dho 925,1, Dho 961,1 and SaU 449,3 samples. (a) End member feldspar composition ternary where Ca% (anorthite end member) = atomic 100×Ca/(Ca+K+Na), Na% (albite end member = atomic 100×Na/(Na+K+Ca) and K% (orthoclase end member) = atomic 100×K/(K+Na+Ca). (b) Frequency histogram of plagioclase An# compositions.

Figure 7. Spinel compositions in clasts in Dho 925,1, Dho 961,1 and SaU 449,3 compared in (c) with spinel in other lunar meteorites (La Paz basalt group: Joy et al., 2006; Miller Range 05035: Joy et al., 2008; and North West Africa 4472: Joy et al., 2011a). Spinel fields have been taken from Haggerty (1973). Abbreviations in (c): S = spinel *sensu stricto*, H = hercynite, C = chromite, PC = picro-chromite.

Figure 8. Trace element concentrations in plagioclase in Dho 925,1 and Dho 961,1 (Table EA2.1). Measurements have been divided into clast type where (a) shows plagioclase chondrite normalised rare earth element (Anders and Grevesse, 1989) in the samples matrix; (b) plagioclase REE in 'type 1' crystalline impact melt breccia clast Dho_961_1_Clast_52 (EA1.7). (c) REE in 18 plagioclase grains from 9 different equilibrated granulite clasts and; (d) plagioclase REE in granulite clast Dho_925_1_Clast_51. (e) – (h) Plagioclase composition of the Dhofar group compared with Apollo rock suites (ferroan anorthosite (FAN), high-Alkali (HAS), Mg-Suite and mare basalts: Papike et al., 1996; 1997; Shervais and McGee, 1998, 1999; Floss et al., 1998; Joy et al., 2006, 2008; Schnare et al., 2008). Trends to FAN and HAS plagioclase in (f) are power law fits where the difference is caused by the enhanced presence of Na and K-rich plagioclase in the HAS Suite, compared with Ca-rich plagioclase in FAN rocks (see Joy, 2013). ITEs = incompatible trace elements.

Figure 9. REE chondrite normalised (Anders and Grevesse, 1989) composition of pyroxene in Dho 925,1 and Dho 961,1 sorted by clast type (Table EA2.2). (a) Matrix Ca-poor pyroxene fragment. (b) A

Ca-rich pyroxene and two Ca-poor pyroxene in 'type 1' crystalline impact melt clasts (Dho_961_1_Clast_52 and Dho_925_1_Clast_13). (c) Ca-rich and Ca-poor pyroxene compositions in granulite clasts (Dho_925_1_Clast_39 and Dho_961_1_Clast_46). Also shown is the bulk composition of high-K KREEP for comparison (Warren, 1989).

Figure 10. Minor and trace elements in olivine in Dho 925,1 and Dho 961,1 (Table EA2.3). (a) Chondrite normalised (Anders and Grevesse, 1989) REE patterns of olivine in granulite clasts and in type 1 crystalline impact melt breccias. (b) Cr/Mn ratio vs. Fo, (c) Co ppm vs. Fo, and (d) Ni ppm vs. Fo content of olivine compared with Apollo Mg-Suite samples (Papike et al., 1998; Shervais and McGee, 1999; Shearer and Papike, 2005; and where rock type labels in (d) are taken from Longhi et al., 2010), mare basalt samples (Papike et al., 1976 *and refs therein*; Papike et al., 1998; Anand et al., 2006; Shearer and Papike, 2005; Schnare et al., 2008; Borg et al., 2009) and ferroan anorthosite samples (Bersch et al., 1991; Papike et al., 1998; Shearer and Papike, 2005).

Figure 11. Apatite OH-F-Cl compositions (Table EA2.5) projected onto a ternary diagram (note that only the bottom right corner of the ternary is shown). Also shown is the average detection limit for the missing OH component using the EMPA method (calculated by McCubbin et al., 2011).

Figure 12. Composition of clasts in Dho 925,1 and Dho 961,1 (Table EA2.4), and bulk composition of the grouped meteorites (Table 1). (a) Bulk clast Al_2O_3 vs. FeO wt% compared with Apollo high-Mg Suite and high-Alkali Suite and bulk lunar meteorite compositions (compositions taken from the electronic appendix to Wieczorek et al., 2006, and supplemented with data from Joy et al., 2006, 2008, 2010), bulk high-K KREEP (Warren, 1989), average composition of Apollo impact melt breccias (IMB) (Jolliff, 1998; Korotev et al., 1994) and lunar meteorites impact melt breccia

compositions (Joy et al., 2010). Also shown are the bulk composition of Dho 925, 961 and SaU 449 (Korotev et al., 2009), the Dhofar grouped meteorites (Demidova et al., 2007; Korotev, 2012) as listed in Table 1, and the bulk Al_2O_3 and FeO of the mafic gabbronorite impact melt in Dho 961 described by Jolliff et al. (2008). (b) Clast and bulk rock Mg# vs. Ti/Sm (ppm); (c) Sc/Sm vs. Eu/Sm (ppm); (d) Clast and bulk rock Mg# vs. Eu/Sm (ppm) and (e) CI-normalised REE as measured in clasts (clast-bearing glassy impact melt breccias, a type 3crystalline impact melt breccia and a granulite – see Table EA2.4 for reported concentrations) and a melt vein of Dho 961 measured by this study compared with Apollo impact melt compositions. (f) Bulk rock compositions of the Dhofar group and a melt vein of Dho 961 measured by this study (data are listed in Table 1).

Figure 13. Dho 961,1 phosphate U-Pb isotope concordia plot constructed using Isoplot (Ludwig, 2012, we used v4.1 of this software package). Data shown have been corrected for initial Pb using the modern Stacey and Kramers (1975) composition. Error ellipses are 2σ and decay constant errors are included in the error calculation. (a) Matrix phosphates, and apatite in metal fragment of equilibrated granulite clast Dho_961_1_Clast_84 (EA1.5d). (b) Crystalline impact melt breccia type 1 (EA1.6a-c) and type 2 (EA1.6d). (c) Phosphate Pb-Pb ages (range of two σ errors) shown from different clasts and matrix fragments in Dho 961,1.

Figure 14. Identification of lunar regoliths with similar compositions to Dho 961 and the average of the Dho 925/960/961 stones. All data are shown in cylindrical projections of the map of the Moon with 0° latitude and 0° longitude at centre. For Dho 961 the bulk meteorite Th (2.86 ppm), FeO wt% (11.14 wt%) and TiO_2 (0.63 wt%) were used to define the input composition parameters (data from Korotev et al. 2009: Table 1). Values for Th (0.93 ppm), FeO wt% (8.77 wt%) and TiO_2 (0.27 wt%) were used to define the input composition parameters for the average of Dho 925/960/961 (Demidova et al., 2007: Table 1). Both the bulk sample compositions and Lunar Prospector data have internal errors which need to be considered and in all cases the maximum Lunar Prospector 2°/pixel data error

(Prettyman et al., 2006) was larger than the reported 1 sigma bulk sample analytical errors. Therefore, we used compositional ranges of meteorite Th ppm concentration ± 0.398 ppm (max. LP data Th ppm error), meteorite FeO wt% concentration ± 2.998 wt% (max. LP data FeO error), and TiO_2 wt% concentration ± 1.918 wt% (max. LP data TiO_2 error) to define the range of regolith compositions searched. Results are shown for searches just using FeO and Th in (a) and (e), and using FeO, Th and TiO_2 (b) and (f). Grey tones indicate identification of matching elements in each search and on a black (no element range matches) to white scale (all elements satisfied). We interpret these white coloured localities as the most compositional similar regolith to the input meteorite composition (± error) and these regions of interest are shown in (c), (d), (g) and (h) with the underlying Moon Clementine albedo image for geographic context. Major terranes (SPA basin and the PKT) are annotated along with Schödinger basin (S), Mare Moscoviense (MM) and Mare Australe (MA), and cryptomaria in the Schickard-Schiller region (SS).

Table captions

Table 1. Bulk compositions of lunar meteorites Dho 925, 960, 961 and SaU 449 summarised from the literature (Demidova et al., 2007; Korotev et al., 2009; Korotev, 2012). Average Dho 961,1 melt vein composition determined by this study where the * denotes CaO measured using EMP.

Table 2. U-Pb and Pb-Pb data collected from phosphates in the Dho 961,1 thick section (see text for method details).