A dunite fragment in meteorite Northwest Africa (NWA) 11421: A piece of the Moon’s mantle

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ABSTRACT

A centimeter-sized fragment of dunite, the first recognized fragment of Moon mantle material, has been discovered in the lunar highlands breccia meteorite Northwest Africa (NWA) 11421. The dunite consists of 95% olivine (Fo83), with low-Ca and high-Ca pyroxenes, plagioclase, and chrome spinel. Mineral compositions vary little across the clast and are consistent with chemical equilibration. Mineral thermobarometry implies that the dunite equilibrated at 980 ± 20 °C and 0.4 ± 0.1 gigapascal (GPa) pressure. The pressure at the base of the Moon’s crust (density 2550 kg/m³) is 0.14–0.18 GPa, so the dunite equilibrated well into the Moon’s upper mantle. Assuming a mantle density of 3400 kg/m³, the dunite equilibrated at a depth of 88 ± 22 km. Its temperature and depth of equilibration are consistent with the calculated present-day selenotherm (i.e., lunar geotherm).

The dunite’s composition, calculated from mineral analyses and proportions, contains less Al, Ti, etc., than chondritic material, implying that it is of a differentiated mantle (including cumulates from a lunar magma ocean). The absence of phases containing P, Zr, etc., suggests minimal involvement of a KREEP component, and the low proportion of Ti suggests minimal interaction with late melt fractionates from a lunar magma ocean. The Mg/Fe ratio of the dunite (Fo83) is significantly lower than models of an overturned unmixed mantle would suggest, but is consistent with estimates of the bulk composition of the Moon’s mantle.

Keywords: Moon, mantle, dunite, meteorite, thermobarometry, NWA 11421, selenotherm

INTRODUCTION

The Moon’s mantle forms the greatest portion (by volume and mass) of the Moon, and figures prominently in all models of the Moon’s origin and evolution (Shearer and Papike 1999; Elkins-Tanton et al. 2002b; Wieczorek et al. 2006, 2013). Until now, understanding of the lunar mantle has been hindered by the absence of samples of mantle material. Lacking lunar mantle material to examine, the composition and physical state of the Moon’s mantle have been inferred from geophysical data and physico-chemical models: Apollo-era seismology (Kuskov and Kronrod 2009; Zhao et al. 2012; Matsumoto et al. 2015; Garcia et al. 2019); GRAIL measurements of gravity (Wieczorek et al. 2006; Matsuyama et al. 2016); and theoretical models based on those constraints and estimates of the Moon’s bulk composition (Elkins-Tanton et al. 2002a; Wieczorek et al. 2006; Elkins-Tanton et al. 2011). Here, we present the first undisputable sample of the lunar mantle, and its implications for its origin.

Lunar evolution: Background

The standing model of the evolution of the lunar mantle starts as a planet-encompassing lunar magma ocean (LMO), produced during a collision between the proto-Earth and a planetesimal (e.g., Wieczorek et al. 2006; Elkins-Tanton et al. 2011). As the LMO cooled, it crystallized mafic minerals, olivine followed by low-Ca pyroxene and then augite, which sank to form cumulate igneous rocks at the base of the ocean. The minerals became more ferroan (less magnesian) as crystallization proceeded. The mafic minerals accumulated in a chemically and mineralogically layered pile. Plagioclase was among the last minerals to crystallize, and it floated on the remaining LMO to form an anorthositic crust (Wood et al. 1970; Elkins-Tanton et al. 2011; Wieczorek et al. 2013). The last dregs of the LMO were rich in incompatible elements, titanium, and iron (the KREEP component), and were denser than the underlying mafic cumulates (Srivastava et al. 2022). The whole mantle was gravitationally unstable, with denser, ferroan material overlying lighter, magnesian material. Under the influence of gravity, the mantle overturned, bringing magnesian cumulates toward the surface and ferroan and Ti-rich materials to depth (Hess and Parmentier 1995; Elkins-Tanton et al. 2011). The overturned mantle could have been chemically layered, with the original stratigraphy essentially intact but inverted (Elkins-Tanton et al. 2011). Alternatively, the overturned mantle could have been mixed to various degrees (Boukaré et al. 2018; Zhao et al. 2019; Moriarty et al. 2021a; Schwingier and Breuer 2022). For a detailed summary see Gross and Joy (2016).

The lunar crust, originally the plagioclase flotation cumulates, is less than 45 km thick as calculated from gravity and seismic data (Wieczorek et al. 2013). It is reasonable that larger impact basins would have penetrated the crust and exposed and/or excavated lunar mantle material (Morrison 1998; Potter et al. 2012; Vaughan et al. 2013; Miljković et al. 2015; Moriarty et al. 2021a). Many outcrops of olivine rich material have been identified around lunar basins, and could represent uplifted mantle material (Nakamura et al. 2009; Yamamoto et al. 2021).
al. 2010; Klima et al. 2011; Kramer et al. 2013; Moriarty et al. 2010; Bischoff et al. 2010; Mercer et al. 2013; Cao et al. 2020; Bechtold et al. 2021), which is inferred to have formed at depth and cooled slowly, as suggested by similar chemical zoning patterns in Al, Ti, and P; this zoning led Shearer et al. (2015a) to infer a shallow crustal origin. However, similar chemical zoning has recently been recognized in the troctolite 76535 (Nelson et al. 2021), which is likely to be fragments of crustal cumulate rocks (Taylor and Marvin 1971; Morris et al. 1990). A few dunite fragments are reported from Apollo 15 regolith breccias (Marvin et al. 1989a, 1989b, 1991). Marvin and colleagues suggested that these fragments formed at significant depth, but did not distinguish between a crustal and mantle origin. Finally, the Apollo 17 basalt 74275 contains xenoliths of dunite (Shearer et al. 2015a). The xenoliths’ olivine cores retain igneous-like zoning patterns in Al, Ti, and P; this zoning led Shearer et al. to infer a shallow crustal origin. However, similar chemical zoning has recently been recognized in the troctolite 76535 (Nelson et al. 2021), which is inferred to have formed at depth and cooled slowly, see Figure 4 here and McCallum and Schwartz (2001).

**Meteorites**

Among lunar meteorites, only a few dunitic and peridotitic clasts have been reported in regolith or melt breccias; no lunar meteoritic dunites (or dunitic peridotites) are known. In meteorite ALH 81005, despite extensive study of many thin sections, only a few peridotitic fragments have been reported (Kurat and Brandstätter 1983; Warren et al. 1983; Brum 2022; Brum et al. 2022). Dunitic and peridotitic material are absent to uncommon in other lunar meteorites (Warren et al. 1983; Arai et al. 2002; Nazarov et al. 2004; Sugihara et al. 2004; Hudgins et al. 2007; Bischoff et al. 2010; Mercer et al. 2013; Cao et al. 2020; Bechtold et al. 2021). Many of these fragments have moderate Mg*, and are likely related to the lunar Mg-suite (Shearer et al. 2015b).

### NWA 8046 clan

Here, we present a clast of dunite in lunar meteorite Northwest Africa (NWA) 11421, which is a member of the NWA 8046 clan.
A dunite clast is also present in NWA 14900 (Sheikh et al. 2022), a member of the NWA 8046 clan (Korotev 2022). This fragment consists only of olivine (Fo89) with a miniscule proportion of chromite. No other studies of NWA 8046 meteorites mention clasts of dunite, peridotite, or other ultramafic materials (Lunning and Gross 2019; Fagan and Gross 2020; Zeng et al. 2020; Treiman and Semprich 2021; Saini et al. 2022).

**SAMPLES AND METHODS**

**Samples**

A piece of NWA 11421, 11.67 g, was purchased from M. Cimala of Polandmet.com (Fig. 1). The properties of this fragment are consistent with the official description of the meteorite (Gattacceca et al. 2019). NWA 11421 is a member of the NWA 8046 clan of impact melt breccias, which consist of mineral and lithic fragments (mostly anorthositic troctolite or lherzolite) in dense black glass, see Figure 1 (Treiman and Coleff 2018). The dunite clast studied here, D1 (Fig. 1), was noted on a weathered surface by its color, and its extent determined with X-ray computed tomography, XCT, see Figure 2 (Treiman and Coleff 2018). Based on the XCT, the sample was cut to produce two thick sections (labeled NWA 11421_lpi1 and _lpi2) that expose the dunite clast (NWA 11421_lpi1_D1), leaving a significant portion of it remaining in the meteorite fragment. The results here are all from NWA 11421_lpi1.

**Methods**

**Electron microbeam.** The dunite D1 and its surroundings in thick section NWA 11421_lpi1 were imaged in backscattered electron (BSE) mode using the JEOL 7600F at the Astromaterials Research and Exploration Science (ARES) Division, Johnson Space Center, Houston, Texas and with the PhenomXL SEM at the Lunar and Planetary Institute (USRA), Houston, Texas. Qualitative maps of element abundances (by energy-dispersive spectrometry, EDS), were acquired with the same instruments. Based on these element mappings, selected spots were chosen for quantitative chemical analysis using the JEOL 8530 FEG electron microprobe (EMP) at ARES. Analytical conditions were nominal for the instrument and laboratory. Peak intensities were measured for Kα radiation of these elements using well characterized standards: Si, diopside; Ti, rutile; Al, olivoclas; Cr, chromite; Fe, fayalite; Ni, NiO; Mn, rhodonite; Mg, diopside or forsterite; Ca, diopside; Na, olivoclas; K, orthoclase; and S, anhydrite. The incident electron beam was at 15 kV and 10 nA (for plagioclase) or 25 nA (for other minerals) measured in a Faraday cup, and was focused on surfaces of standards and samples. Peak X-ray intensities were counted for 30–60 s, and backgrounds were counted for the same total durations. Analytical standards were run as unknowns to validate the calibrations.

All mineral analyses and their locations on the thick section are given in the Online Materials.  

**Mineral proportions, densities, and bulk composition.** Mineral proportions in the dunite clast D1 were calculated using a supervised classification routine on element abundance images from the SEM. The classification was done using the Multispec program (Landgrebe and Biehl 2011) following the protocols of Maloy and Treiman (2007). X-ray element images were masked to include only the D1 clast, mineral classification training was based on EMP quantitative point analyses. Mineral densities (at 1 bar) were calculated assuming linear mixing from the densities of end-member compositions.

D1’s bulk composition was calculated from the area proportion of each phase in it (olivine, orthopyroxene, clinopyroxene, plagioclase, chromite), the point EMP analyses of each phase, and the calculated densities of each.

**X-ray computed tomography (XCT).** An X-ray tomography image stack of the whole meteorite piece was acquired at the ARES division of Johnson Space Center in May 2018 (Treiman and Coleff 2018). The XCT instrumentation and methods are as described in Zeigler et al. (2017) and Eckley et al. (2020); see Online Materials’ Appendix I.

**Thermobarometry.** Details of the thermobarometry calculations are given in the text below and in Online Materials’ Appendix II, including mineral analyses on which they are based.

**Results**

The analyzed section of NWA 11421 is a lunar highlands melt breccia (Figs. 1, 2, and 3) consistent with its classification (Gattacceca et al. 2019) and its pairing into the NWA 8046 clan (Korotev 2022). Most of the lithic fragments in the section are troctolitic or lherzolitic anorthosites (Figs. 2d and 3a); there are also clasts of anorthositic impact melt and breccia. No basaltic or KREEPy fragments have been noted, although the rock contains rare small fragments of evolved, silica-rich material (Treiman and Semprich 2019). Mineral grains include those from the anorthositic and dunitic lithologies, and other types including exsolved low-Ca and high-Ca pyroxenes, Fe-sulfide, Fe-metal, and Mg-Al spinel (Fig. 3a).

The thick section _lpi1 (and the meteorite in general) shows minor evidence of terrestrial weathering (Korotev 2022). One crack in the D1 dunite and its surroundings contains K-rich material, tentatively identified as clay (Fig. 3c). Another crack contains a Ca-rich grain (Fig. 3c), without other elements detectable by EMP, that is likely a Ca-carbonate. That same crack also contains a sulfur-bearing grain, again without elements detected in our EMP maps (Fig. 3c). This could be a grain of barium sulfate (barite), such as occurs in other NWA meteorites (Korotev et al. 2009).

**Dunite mineralogy and composition**

The D1 dunite fragment, before cutting, was approximately 10 × 7 × 4 mm (Figs. 1 and 2). The thick section analyzed here exposes a 5 × 4 mm surface of the dunite (Fig. 3). The thick-section surface appears representative of the whole fragment, except that it does not expose an apparent stringer of high-density...
Mineralogically, the D1 dunite is simple; it consists only of olivine, low-Ca pyroxene, high-Ca pyroxene, anorthite plagioclase, and chromite (Fig. 3a). No other phases were detected (Fig. 3), such as Fe ± Ti oxides, Ca-phosphates, zircon/baddeleyite, alkali feldspar, or garnet. A small proportion of K-bearing material on a fracture (Fig. 3c) is interpreted as clay produced during terrestrial weathering. Minerals in the dunite are nearly of constant compositions and lack zoning in major or minor elements (Fig. 4); Table 1 gives average mineral compositions, and a calculated bulk composition of the dunite (based on mineral proportions, analyses, and densities; Table 2); full analyses are in the Online Materials.

The silicate minerals are magnesian; the olivine is Fo83 (Fig. 4; Table 1), and the pyroxenes are slightly more magnesian (Wo0.3En0.8Fs0.9; and Wo0.4En0.5Fs0.1, Table 1), consistent with Fe-Mg equilibria (Baker and Herzberg 1980; Lindsley 1983). The olivine has FeO/MnO = 84, consistent with a lunar provenance (Karner et al. 2003). The pyroxenes contain minor non-quadrilateral components (e.g., Al, Ti, Cr, Na), and so are represented well on a standard pyroxene quadrilateral (Fig. 4). The augite is slightly sub-calcic, and the orthopyroxene contains a small proportion of Ca (Table 1; Fig. 4). D1’s plagioclase is An96.5 (see Table 1 for values and abbreviations), such as is abundant in lunar anorthosites and most lunar rocks. Its chromite is a complex solid solution, with significant Al substitution for Cr, Mg substitution for Fe2+, and a small proportion of Ti. The analytical sums for chromite are low, which we attribute to its small grain size and irregular surfaces near grain edges; the chromite standard analyzed well. It is also possible that the dunite’s chromite contains an unanalyzed element or a bit of ferric iron.

Mineral proportions and their calculated densities are given in Table 2, along with a calculated bulk density for the D1 dunite.

**Dunite texture**

Macroscopically, the D1 dunite has a granoblastic-polygonal texture, and lacks apparent preferred mineral orientations. This absence of preferred orientations is also seen on the weathered meteorite surface (Fig. 1), in the different colors of the olivine, augite, and orthopyroxene grains. Note that the weathered and cut surfaces are approximately perpendicular to each other. Likewise, the XCT scan shows no alignments or preferred ori-
entations of the pyroxene and plagioclase grains (Fig. 2). Thus, we infer that the dunite lacks linear and/or planar structures; i.e., it is structurally isotropic.

Olivine grains can be distinguished from each other, at least in part, by the presence of gaps (from grinding/polishing) along grain boundaries and cracks representing cleavage in individual grains (Figs. 3 and 5). From this view, the D1 olivines are all of approximately the same size, ~100 µm across, and show no obvious preferred elongation direction. Boundaries between grains of olivine (as can be discerned) are generally planar (Fig. 5) and are consistent with inferred constraints of equilibrium surface energy (Spry 1969; Barker 2013).

Symplectite

Chrome and some augite in the D1 dunite are exceptions to this textural equilibrium. Nearly all the chromite occurs either as symplectic intergrowths with augite (Figs. 5c and 5d) or elongate grains, sandwiched between silicate mineral grains. Much of the augite is in equant anhedral grains (e.g., Fig. 3a, upper right side of dunite clast), but some augite is in symplectic intergrowths with chromite (Figs. 5c and 5d), and some occurs as elongate grains between other mineral grains (e.g., at the center of Fig. 5b).
The largest example of elongated grains in the thick section is in the lower left corner of the dunite in Figure 3a. There, an elongate augite grain and an augite-chromite symplectite define a short veinlet cutting the dunite. This veinlet could be an example of the high-density veinlets observed in XCT (Figs. 2b and 2c).

Inferences

Thermobarometry

The mineral compositions in the NWA 11421_lpi1_D1 dunite appear to represent a state of chemical equilibrium, so we can apply thermobarometry to determine its equilibrium temperature and pressure. We consider the minerals to be in chemical equilibrium first because the compositions of each mineral are consistent across the dunite fragment (Fig. 4; Online Materials); second because the Mg* of the olivine, augite, and orthopyroxene are consistent with equilibrium, see Table 1 and Figure 4 (Baker and Herzberg 1980; Lindsley 1983); and finally because the Ca contents of the augite and orthopyroxene are consistent with equilibrium (Lindsley 1983).

With this evidence of chemical equilibrium, we can apply established mineral thermobarometers to determine the temperature and pressure at which the dunite’s minerals equilibrated: 980 ± 20 °C and 0.4 ± 0.1 GPa. We calculated temperatures and pressures for six different sets of minerals (Ol + Pl + Cpx + Opx) in direct or nearly direct contact, Figure 6 (see Online Materials). Details of the temperature and pressure calculation are given in Online Materials Appendix II. Equilibration temperatures were calculated from two-pyroxene thermometry (Ca distribution between augite and orthopyroxene) using the calibration of Brey and Köhler (1990) and two calibrations from Putirka (2008). For each set of minerals, these temperatures are within 20 °C of each other. The resulting minimum and maximum temperatures were then used as input to calculate pressures using THERMOCALC’s avP algorithm (Powell and Holland 1994, 2008), selecting the temperature that produced the P result with the lowest residuals. The calculated pressures rely primarily on the Al contents of pyroxenes, e.g., the Mg- and Ca-Tschermak’s, or kushiroite (Kimura et al. 2009), components. To validate the procedure, we calculated temperatures and pressures for the lunar troctolite 76535 (Fig. 6; Online Materials Appendix II); our results are comparable to those in earlier studies (McCallum and Schwartz...
2001; Elardo et al. 2012). For the six sets of minerals in the D1 dunite, calculated equilibrium temperatures range from 940 to 990 °C (Fig. 6), with an average of 980 °C. Calculated equilibrium pressures range from 0.27 ± 0.1 to 0.51 ± 0.1 GPa (Fig. 6), with a best estimate of 0.4 ± 0.1 GPa.

Equilibration depth: The upper mantle

To understand the original geologic setting of the NWA 11421_lpi1_D1 dunite, it is crucial to know how the calculated equilibrium pressure corresponds to depth. We use the Wieczorek et al. (2013) model of the lunar crust and upper mantle: a porous (fragmented) anorthositic crust with average density of 2550 kg/m³ and thickness from 34 to 43 km, overlying a peridotitic upper mantle of density 3400 kg/m³ (nearly identical to that calculated for the dunite, Table 2). With those constraints (and lunar surface gravity of 1.62 m/s²), pressure at the base of the crust is calculated to be 0.14 to 0.18 GPa. The equilibration pressure for D1 is greater than these pressures, which places D1’s equilibration in the Moon’s upper mantle. Using the Wieczorek et al. (2013) model, the D1 dunite equilibrated at a depth of 88 ± 22 km in the Moon’s upper mantle.

The dunite’s mineral equilibration is consistent with formation on a “normal” present-day lunar geotherm (i.e., selenotherm), see Figure 7. Within 2σ uncertainty, the dunite’s equilibration is consistent with the present-day selenotherm calculated by Khan et al. (2014) from seismic data. That selenotherm includes consideration of a partially molten mantle at depths >1200 km, and a porous crust of low thermal conductivity. The nominal pressure and temperature of equilibration plot at slightly higher temperature (or lower pressure) than Khan’s selenotherm (Fig. 7), consistent also with equilibration along an ancient, slightly hotter, thermal gradient.

Texture

The texture of the D1 dunite (excepting the symplectites) arose during its chemical and thermal history in the Moon, and so reveals some of that history. As described above, olivine grains in D1 are all of approximately the same size (Fig. 3), and show no obvious elongations or preferred orientations. None of the minerals in D1 show their own crystal forms (i.e., are not idiomorphic); rather, grain boundaries are straight or curved as consistent with equilibria of mineral surface energies (Spry 1969).

These textures of D1 are consistent with those of a granoblastic-polygonal metamorphic rock—one in which grain sizes and boundaries have adjusted to equilibrium shapes during extensive thermal metamorphism without deformation. Granoblastic-polygonal textures are common among mantle rocks from the Earth (Mercier and Nicolas 1975), although Earth mantle rocks tend to have larger grains (e.g., ~1 mm vs. the 0.1 mm of D1). Such textures are not characteristic of igneous cumulate rocks (Wager et al. 1960; Wager and Brown 1967).

Symplectite formation

The chrome- augite symplectites in the D1 dunite require explanation in the context of long-standing controversies about symplectite formation in other lunar rocks. In the still-current summary, Bell et al. (1975) described six varieties of lunar symplectites and four general formation mechanisms. The symplectites in D1 fall into Bell’s category C, “... 10–1000 μm elongated masses along grain boundaries...” (Figs. 5c and 5d). Bell and coauthors agreed that type-C symplectites are formed by reactions between olivine and plagioclase. Dymek et al. (1975) inferred that similar symplectites in 72415 were formed by in-
terated with a silicate fluid (i.e., in an open system). Elardo et al. (2012) confirmed this inference, showing that symplectites of this type in troctolite 76535 (their Fig. 1) formed by the addition of Cr and Fe in an open-system process. The D1 symplectites are similar enough to those in 76535 (Elardo et al. 2012) in texture and in composition that a similar open system origin seems reasonable. An origin by garnet breakdown seems unlikely for the D1 symplectites because garnet in peridotites tends to form euhedra (Spry 1969; Dégi et al. 2010; Barker 2013) and not intergranular pods and films (Figs. 2b, 2c, 3a, 5c, and 5d).

**The lunar upper mantle**

It seems presumptuous to extrapolate from a single clast in a breccia to the Moon’s whole mantle, yet such assumptions have proven useful (Wood et al. 1970). If the D1 dunite clast in NWA 11421 is representative of a portion of the lunar mantle, it could help constrain models of the Moon’s early history.

The Al, Ca, and Ti abundances in the D1 dunite seem most consistent with formation in a differentiated lunar mantle that was well mixed after its overturn (see Introduction). Estimated compositions of the bulk, undifferentiated lunar mantle have 3–7% Al$_2$O$_3$, 3–5% CaO, and 0.2–0.4% TiO$_2$ (Elkins-Tanton et al. 2011), while the dunite contains only 0.55% Al$_2$O$_3$, 0.57% CaO, and 0.07% TiO$_2$ (Table 1). Thus, the dunite composition is consistent (in general terms) with a primitive mantle composition that was depleted in components that partition into silicate melt (e.g., Al, Ca, Ti).

The dunite’s Mg* (i.e., Fo) of 83 is consistent with most models of the bulk primitive lunar mantle, which have Mg* = 80–85; see Table 1 of Elkins-Tanton et al. (2011). A lunar mantle that differentiated during a magma ocean episode would retain that average bulk Mg* and be stratified with the highest Mg* at its base (according to mineral-melt element partitioning). In some models of LMO crystallization, Fo$_{83}$ olivine is calculated to form only after ~65–75% of LMO crystallization (depending on the model starting composition), and is nearly the last olivine to crystallize (Elkins-Tanton et al. 2011; Lin et al. 2020; Johnson et al. 2021). In other models of LMO crystallization Fo$_{83}$ olivine does not crystallize (Snyder et al. 1992; Rapp and Draper 2018); low-Ca pyroxene would be the only silicate with Mg* = 83.

This cumulate pile from a crystallizing LMO would have been gravitationally unstable, having the Fe-rich denser materials near the top. This pile would have overthrown, bringing denser material to the mantle base with some degree of mixing (Hess and Parmentier 1995). If there had been no mixing after overturn, e.g., Figure 5b of Elkins-Tanton et al. (2011), olivine at the depth inferred for the NWA 11421 dunite would be ~Fo$_{88}$, significantly more magnesian than observed (Table 1). This mismatch in Fo number implies that at least some of the lunar mantle had been mixed during overturn. However, the presence of augite-chromite symplectites that post-date dunite formation (see below) allows the possibility that the original dunite might have been somewhat more magnesian than what we now see, having equilibrated with the Fe-Cr-bearing material responsible for the symplectites.

If a stratified differentiated lunar mantle had homogenized during or after overturn (Boukaré et al. 2018; Zhao et al. 2019), then it would maintain its bulk Mg* across all depths and so would be consistent with the Mg* and inferred depth for the NWA 11421 dunite (Table 1). Likewise, abundances of Al and Ti in the dunite are consistent with a differentiated mantle, one from which igneous, incompatible elements had been partially removed to form the lunar anorthositic crust and incompatible-enriched late LMO melts. So, the NWA 11421 dunite is most consistent with a lunar mantle that was mixed well after (or during) its overturn (Boukaré et al. 2018; Moriarty et al. 2021a).

**Discussion: Other possible samples of the lunar mantle**

To our knowledge, D1 in NWA 11421 is the first lunar sample known to have equilibrated at pressures consistent with the lunar mantle. It is possible that other lunar dunites and peridotites are samples of the lunar mantle, but few are reported to have mineral assemblages (olivine–plagioclase–augite–low-Ca pyroxene) that could provide equilibration temperatures and pressures (see Online Materials’ Appendix II). See Online Materials’ Appendix III for comments about thermobarometry of lunar spinel cataclasites.

However, many lunar symplectites have bulk compositions consistent with mixtures of garnet ± olivine (Bell et al. 1975), which suggests that they were originally those minerals and decomposed to augite + chromite on decompression (Bell et al. 1975; Schmitt 2016). Specifically, symplectites in dunite 72415 have been interpreted as products of the decomposition of garnet (Schmitt 2016; Bhanot et al. 2022). If so, the garnet must have originated in the lower lunar mantle at pressures greater perhaps than 2.3 GPa (Schmitt 2016). The garnet would then have been transported, perhaps during the overturn of the LMO cumulates, to the shallow mantle (Bhanot et al. 2022), where it could have decomposed to symplectites and then would have been transported to the surface.

**Implications**

The D1 dunite clast in NWA 11421 _lpli _equilibrated last at ~980 °C and 0.4 ± 0.1 GPa, at a depth of 88 ± 22 km, firmly in the Moon’s upper mantle. This temperature and pressure are consistent with estimates of the present-day selenotherm (Khan et al. 2014). Its chemical composition (Mg*, Al content) is consistent with estimates of the bulk composition of the lunar mantle, suggesting that the dunite formed after mantle differentiation (separation of anorthositic crust and Fe-Ti-rich residua) and after density-driven overturn re-homogenized mantle. This interpretation of the D1 dunite’s origin is not unique—a similar chemistry and texture could form from an undifferentiated mantle composition by the removal of a partial melt or perhaps a garnetiferous peridotite.

The veinlets and masses of augite and augite-chromite symplectite represent a fluid-based metasomatic event after the dunite host had achieved textural equilibrium (presumably still in the mantle). Similar metasomatic products occur in other lunar and asteroidal samples (Elardo et al. 2012; Vaci et al. 2021), and their origin remains unclear, especially the nature and origin of the metasomatic fluid.

The D1 dunite is the first recognized sample of the lunar mantle, although mantle rock is inferred to have been brought to the surface by large impact events (Yamamoto et al. 2010; Miljković et al. 2015; Bretzfelder et al. 2020; Moriarty et al. 2021).
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2021a). It is puzzling that no other bits of mantle dunite have been recognized, despite the relative abundance of crustal-intrusive rocks in the meteorite and Apollo collections, e.g., McCallum and Schwartz (2001) and Elardo et al. (2012). Finding other fragments of lunar mantle rock would be very useful, and the search should be widened. The clast described here was recognized first because it was exposed on a weathered surface; where possible, XCT scans of other lunar breccias could reveal more fragments of the lunar mantle.

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Endnote:

1Deposit item AM-23-128911. Online Materials are free to all readers. Go online, via the table of contents or article view, and find the tab or link for supplemental materials.