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► **To cite this version:**

James Day, Christopher A. Corder, Douglas Rumble Iii, Nelly Assayag, Pierre Cartigny, et al.. Differentiation processes in FeO-rich asteroids revealed by the achondrite Lewis Cliff 88763. *Meteoritics and Planetary Science*, 2015, 50, pp.1750-1766. 10.1111/maps.12509 . insu-01447679

HAL Id: insu-01447679

<https://insu.hal.science/insu-01447679>

Submitted on 12 Aug 2021

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Differentiation processes in FeO-rich asteroids revealed by the achondrite Lewis Cliff 88763

James M. D. DAY^{1*}, Christopher A. CORDER¹, Douglas RUMBLE III², Nelly ASSAYAG³, Pierre CARTIGNY³, and Lawrence A. TAYLOR⁴

¹Geosciences Research Division, Scripps Institution of Oceanography, La Jolla, California 92093–0244, USA

²Geophysical Laboratory, Carnegie Institution of Washington, Washington, District of Columbia 20015, USA

³Institut de Physique du Globe de Paris, Sorbonne Paris Cité, Univ. Paris Diderot, UMR 7154 CNRS, 1 rue Jussieu, 75238 Paris, France

⁴Department of Earth and Planetary Sciences, Planetary Geosciences Institute, University of Tennessee, Knoxville, Tennessee 37996, USA

*Corresponding author. E-mail: jmdday@ucsd.edu

(Received 01 December 2014; revision accepted 26 July 2015)

Abstract—Olivine-dominated (70–80 modal %) achondrite meteorite Lewis Cliff (LEW) 88763 originated from metamorphism and limited partial melting of a FeO-rich parent body. The meteorite experienced some alteration on Earth, evident from subchondritic Re/Os, and redistribution of rhenium within the sample. LEW 88763 is texturally similar to winonaites, has a $\Delta^{17}\text{O}$ value of $-1.19 \pm 0.10\text{‰}$, and low bulk-rock Mg/(Mg+Fe) (0.39), similar to the FeO-rich cumulate achondrite Northwest Africa (NWA) 6693. The similar bulk-rock major-, minor-, and trace-element abundances of LEW 88763, relative to some carbonaceous chondrites, including ratios of Pd/Os, Pt/Os, Ir/Os, and $^{187}\text{Os}/^{188}\text{Os}$ (0.1262), implies a FeO- and volatile-rich precursor composition. Lack of fractionation of the rare earth elements, but a factor of approximately two lower highly siderophile element abundances in LEW 88763, compared with chondrites, implies limited loss of Fe-Ni-S melts during metamorphism and anatexis. These results support the generation of high Fe/Mg, sulfide, and/or metal-rich partial melts from FeO-rich parent bodies during partial melting. In detail, however, LEW 88763 cannot be a parent composition to any other meteorite sample, due to highly limited silicate melt loss (0 to <<5%). As such, LEW 88763 represents the least-modified FeO-rich achondrite source composition recognized to date and is distinct from all other meteorites. LEW 88763 should be reclassified as an anomalous achondrite that experienced limited Fe,Ni-FeS melt loss. Lewis Cliff 88763, combined with a growing collection of FeO-rich meteorites, such as brachinites, brachinite-like achondrites, the Graves Nunataks (GRA) 06128/9 meteorites, NWA 6693, and Tafassasset, has important implications for understanding the initiation of planetary differentiation. Specifically, regardless of precursor compositions, partial melting and differentiation processes appear to be similar on asteroidal bodies spanning a range of initial oxidation states and volatile contents.

INTRODUCTION

Partially melted “primitive” achondrite meteorites offer insights into the earliest stages of planetary differentiation, as they provide evidence for limited melting and melt segregation in early-formed asteroids. The textural, lithological, and chemical diversity

displayed by partially melted achondrites emphasize the wide range of chondritic parental materials and the complex temporal relationship of melting and cooling experienced by their parent bodies (e.g., Mittlefehldt et al. 1998). In turn, the processes of melting, volatile-depletion, and metal-silicate equilibration on small asteroidal bodies have potential collateral effects on the

accretion of larger planets, which ultimately coalesce from smaller bodies during planetary growth (Day 2015). Partially melted achondrites include metamorphosed- and anatectic-chondritic materials, such as winonaites (Prinz et al. 1980; Benedix et al. 1998) and acapulcoites (e.g., Mittlefehldt et al. 1996), to lithologies that have witnessed variable degrees of partial melting, melt extraction, or are themselves partial-melt products, such as lodranites (e.g., Mittlefehldt et al. 1996; McCoy et al. 1997), ureilites (e.g., Warren et al. 2006; Goodrich et al. 2007; Bischoff et al. 2014), brachinites and brachinite-like achondrites (e.g. Mittlefehldt et al. 2003; Day et al. 2012; Gardner-Vandy et al. 2013; Keil 2014), or the oligoclase-rich meteorites GRA 06128/9 (e.g., Day et al. 2009, 2012; Shearer et al. 2010).

While many of these meteorite groups are defined, by among other things, the high $Mg/(Mg+Fe)$ of their silicates (e.g., winonaites, acapulcoite-lodranites, ureilites; e.g., Mittlefehldt et al. 1998), an increasing array of meteorites are being recognized with more ferroan compositions, suggestive of evolution from distinctive sources. These FeO-rich meteorite types include the brachinites, brachinite-like achondrites, and GRA 06128/9 meteorites (Day et al. 2012), the recrystallized primitive achondrite Tafassasset (Gardner-Vandy et al. 2012) and the poikilitic cumulate NWA 6693 (Warren et al. 2013). High-FeO partially melted achondrite meteorites are important materials for understanding inherent variation in oxygen fugacity and volatile abundances in the early solar system, as well as for examining the nature of precursor “chondritic” parental materials forming planets and planetesimals.

One particular FeO-rich meteorite, Lewis Cliff 88763, is an olivine-dominated stone that was found in the Antarctic ice in 1988. It was a small meteorite ($2 \times 1.6 \times 0.6$ cm) with an original mass of 4.1 g and has been described as a Brachina-like achondrite with a dark crystalline interior and fusion crust covering 95% of the sample (Lindstrom 1991). Further study of LEW 88763, notably by Swindle et al. (1998), found the meteorite to be mainly composed of olivine with accessory augite, pigeonite, plagioclase, chromite, whitlockite, ilmenite, troilite, and taenite. These authors considered that LEW 88763 was an “olivine-dominated primitive achondrite [whose] chondritic major element chemistry and high abundance of siderophile elements suggests that it is more primitive than brachinites.” In light of recent discoveries (e.g., GRA 06128/9; NWA 6693) and recognition of a variety of ferroan partially melted achondrite meteorites (e.g., brachinites and brachinite-like achondrites), we revisit the petrology and geochemistry of LEW 88763. In doing so, new insights

are revealed on differentiation initiated in FeO-rich, partially melted asteroids formed from carbonaceous chondrite-like sources.

SAMPLE AND METHODS

A polished thick section (LEW 88763, 14; total surface area = 0.27 cm^2) and a bulk-rock fragment (LEW 88763, 20; total sample weight = 0.202 g) of LEW 88763 were obtained from the Meteorite Working Group. Petrography of the polished thick section was performed using a Nikon petrographic microscope in plane-polarized, cross-polarized, and reflected light, including production of photomicrograph maps in all light formats. Modal abundances were calculated using the photomicrograph maps in conjunction with Image J processing software (methods similar to those in Day et al. 2006). Cropped maps contained $>4,000,000$ pixels and were converted to gray scale before being processed with Image J. Modes were obtained using a color contrast thresholding method, where thresholds were selected based on knowledge of the mineral phases in the sample. Threshold selections were confirmed by cross-calibration with petrography and major mineral chemistry.

Both major and minor element mineral compositions were obtained from the polished thick section at the University of Tennessee using a Cameca SX-100 electron probe micro-analyzer. All analyses were made with an accelerating potential of 15 keV and a beam size of $1 \mu\text{m}$. Beam currents were 30 nA for olivines, pyroxenes, and spinels; 20 nA for FeNi metals and Fe sulfides; and 10 nA for feldspars. Both natural and synthetic standards were used to calibrate the EMP and standards were measured throughout analytical sessions to ensure data quality. Background and peak counting times used were 20–30 s and standard PAP (ZAF) correction procedures were used. Detection limits for silicates and spinels were $<0.03 \text{ wt}\%$ for SiO_2 , TiO_2 , Al_2O_3 , MgO , CaO , and Na_2O ; $<0.04 \text{ wt}\%$ for K_2O , V_2O_3 , and Cr_2O_3 ; $<0.05 \text{ wt}\%$ for Cr_2O_3 , MnO , FeO , and P_2O_5 ; $<0.06 \text{ wt}\%$ for NiO ; and $<0.1 \text{ wt}\%$ for BaO . For FeNi metal and troilite, detection limits were $<0.03 \text{ wt}\%$ for Si, P, Mg, Al, and Ti; $<0.04 \text{ wt}\%$ for S, Fe, Cr, Mn, and Mg; and $<0.05 \text{ wt}\%$ for Co and Ni.

Bulk-rock major and trace element abundances were determined using pressure-assisted digestion of LEW 88763 powder in Teflon Parr bomb vessels at the Scripps Isotope Geochemistry Laboratory (SIGL). The powder was digested at $170 \text{ }^\circ\text{C}$ in Optima grade concentrated HF (4 mL) and HNO_3 (1 mL) for $>72 \text{ h}$ in a processing oven along with total procedural blanks, terrestrial basalt standards and the CV3 chondrite meteorite, Allende (the fragment measured was

originally obtained by D. Lal), used as an internal standard in the SIGL and for cross-comparison with other Allende aliquots presented in the literature. Samples were sequentially dried and taken up in concentrated HNO₃ to remove fluorides, followed by dilution and doping with indium to monitor instrumental drift during analysis. Major and trace element abundance analyses were obtained using a ThermoScientific iCAP Qc quadrupole inductively coupled plasma–mass spectrometer in low resolution mode. Analyses were standardized versus reference material BHVO-2 that was measured throughout the analytical run. In addition, four reference materials were analyzed as “unknowns” (Allende, BIR-1, BHVO-2, and BCR-2) in order to address matrix matching, external reproducibility, and accuracy. For major and trace elements, reproducibility, of the reference materials was generally better than 6% (RSD).

Osmium isotope and HSE abundance analyses were performed at the SIGL using methods described in Day et al. (2015). Homogenized powder aliquots of LEW 88763, Allende, and total procedural blanks were digested in sealed borosilicate Carius tubes, with isotopically enriched multi-element spikes (⁹⁹Ru, ¹⁰⁶Pd, ¹⁸⁵Re, ¹⁹⁰Os, ¹⁹¹Ir, ¹⁹⁴Pt), and 6 mL of a 1:2 mixture of multiply Teflon distilled HCl and HNO₃ that was treated with H₂O₂ to expunge Os. Samples were digested to a maximum temperature of 270 °C in an oven for 96 h. Osmium was triply extracted from the acid using CCl₄ and then back-extracted into HBr (Cohen and Waters 1996), prior to purification by microdistillation (Birck et al. 1997). Rhenium and the other HSE were recovered and purified from the residual solutions using standard anion exchange separation techniques. Isotopic compositions of Os were measured in negative-ion mode on a ThermoScientific Triton thermal ionization mass spectrometer at the SIGL. Rhenium, Pd, Pt, Ru, and Ir were measured using a Cetac Aridus II desolvating nebulizer coupled to a ThermoScientific iCAP q ICP-MS. Offline corrections for Os involved an oxide correction; an iterative fractionation correction using ¹⁹²Os/¹⁸⁸Os = 3.08271; a ¹⁹⁰Os spike subtraction; and finally, an Os blank subtraction. Reported precision for ¹⁸⁷Os/¹⁸⁸Os, determined by repeated measurement of the UMCP Johnson Matthey standard was better than ±0.2% (2 SD; 0.11390 ± 20; *n* = 5). Measured Re, Ir, Pt, Pd, and Ru isotopic ratios for sample solutions were corrected for mass fractionation using the deviation of the standard average run on the day over the natural ratio for the element. Reproducibility on HSE analyses using the iCAP q was better than 0.5% (2 SD) for 0.5 ppb solutions, and all reported values are blank corrected. The total procedural blanks (*n* = 2) run with the samples had ¹⁸⁷Os/¹⁸⁸Os = 0.401 ± 0.010, with

quantities (in picograms) of 4.2 ± 1.3 [Re], 2.8 ± 0.3 [Pd], 2.0 ± 0.9 [Pt], 13.3 ± 1.3 [Ru], 0.3 ± 0.2 [Ir], and 0.2 ± 0.1 [Os]. These blanks resulted in negligible corrections to samples (<1%).

Oxygen isotope analysis of LEW 88763 was performed at the Institut de Physique du Globe-Paris. Analytical methods are similar to those documented in Rumble et al. (1997). Briefly, prior to analysis, silicate grains were separated from each other under alcohol in a tungsten-carbide mortar and pestle, ultrasonicated in dilute HCl at room temperature for 240 s to remove weathering products, washed with MilliQ water, dried, and were then separated into magnetic and nonmagnetic fractions. Nonmagnetic fractions of the meteorite samples were analyzed using laser fluorination along with the San Carlos olivine and garnet standard UWG-2 from the Gore Mountain mine, Adirondack Mountains, New York (Valley et al. 1995). Oxygen isotopic ratios (^XO/¹⁶O, where *X* = ¹⁸O or ¹⁷O) are measured versus an international standard, V-SMOW (standard mean ocean water), and expressed in δ notation, according to the following equation: δ^XO = 1000 × ([^XO/¹⁶O]/[^XO/¹⁶O_{std}] – 1). From the delta notation values, Δ¹⁷O (in per mil, representing deviation from the terrestrial fractionation line and given by the following equation: Δ¹⁷O = δ¹⁷O – [(δ¹⁸O/1000 + 1)^{0.524} – 1] × 1000) was then calculated. Values of δ¹⁸O were normalized to UWG-2 garnet with δ¹⁸O = 5.75‰. Measurement of the San Carlos olivine and UWG-2 garnet standard aliquots that were measured gave (in per mil) 2 SD uncertainties on δ¹⁷O = ±0.09, on δ¹⁸O = ±0.17, and for Δ¹⁷O = –0.004 ± 0.018 (2σ; *n* = 18). In-run uncertainties for individual measurements were <0.06‰ for δ¹⁷O and <0.03‰ for δ¹⁸O.

RESULTS

Petrography and Mineral Chemistry

Petrography and mineral chemistry of polished section LEW 88763, 14 agrees well with a prior study of the meteorite (Swindle et al. 1998). The section represents a slice through the meteorite where two sides have an uneven distribution of fusion crust, one side of which is ~200 μm thick, and the other of which is ~50 μm thick (Fig. 1). The uneven distribution is consistent with LEW 88763 representing a fragment of a larger meteoroid that broke up during entry, or preferred aerodynamic orientation during atmospheric entry. Darkened regions of alteration, fine-grained matrix, and porosity make up nearly 50% of the polished section (Table 1). Major mineral phases observed were olivine (~80 modal %, after removal of the dark matrix and fusion crust), high- and low-Ca pyroxene (~9.5%), plagioclase (~6%), and opaque

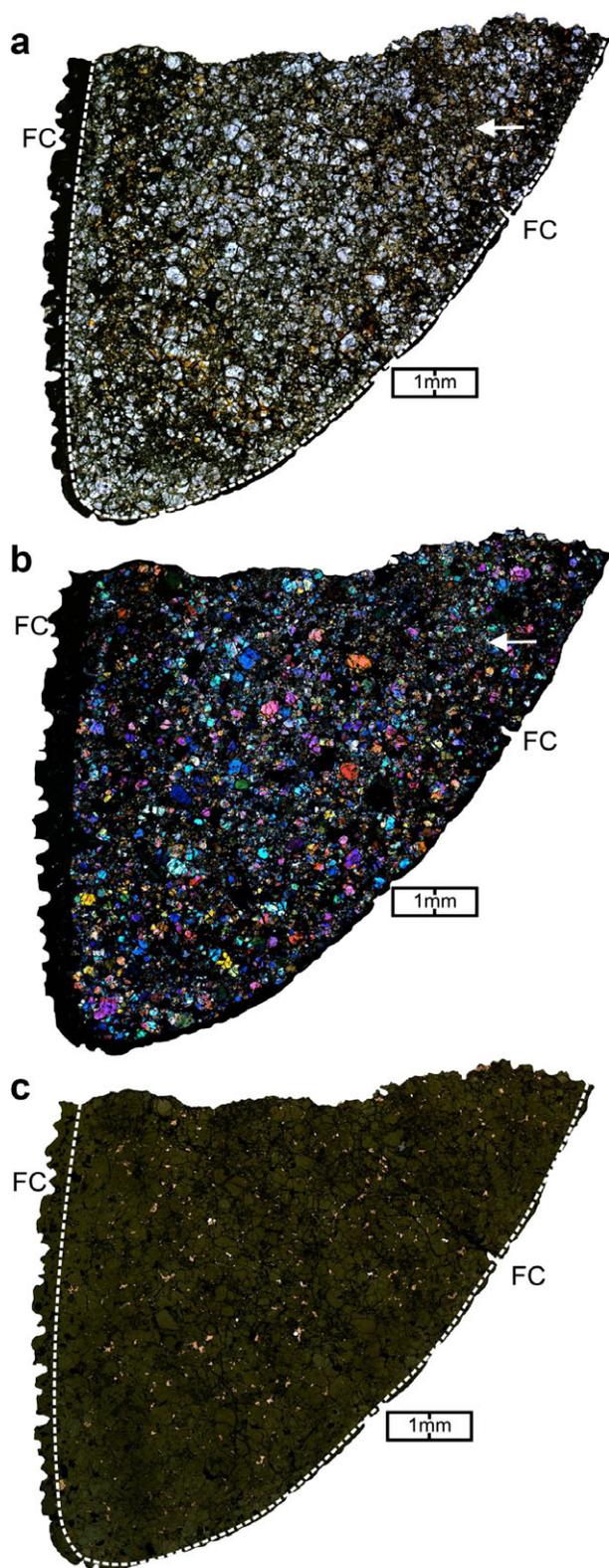


Fig. 1. Photomosaics of LEW 88763, 14 in (a) plane-polarized, (b) cross-polarized, and (c) reflected light. Boundary of the fusion crust (FC) is shown as a dashed white line in (a) and (c). Note the apparent darkening of the interior of the meteorite closest to the fusion crust (a, b) and the high concentration of bright sulfide/metals in the center of the meteorite (c). Arrows in (a) and (b) denote possible relict chondrule (see text for details).

Table 1. Mineral modes in LEW 88763 (in %).

	LEW 88763, 14	LEW 88763, 14 ^a	LEW 88763, 11 ^b
Olivine	42.2	81.6	71
Pyroxene	4.9	5.8	7
Plagioclase	3.0	9.5	10
Opaque phases	1.5	3.1	6
Chromite	0.6	1.3	–
FeNi	0.3	0.5	–
FeS	0.6	1.3	–
Fusion crust/matrix	46.8	–	–
Cracks/alteration	–	–	7
Total	100	100	101

^aNormalized to remove fusion crust.

^bFrom Swindle et al. (1998).

Olivine grains are subhedral to anhedral, generally do not exceed 0.4 mm in diameter, and are of generally uniform size. All other mineral phases are generally less than 0.25 mm in diameter and are typically interstitial to, or trapped within, olivine (Fig. 2). Swindle et al. (1998) measured higher modal abundances of plagioclase and opaque phases in polished section LEW 88763, 11, but also found that crystals were anhedral but of relatively uniform size. A few fine-grained regions with oval shape occur within LEW 88763, 14 that may correspond to relict chondrules (Fig. 1), as also observed by Gardner-Vandy (2012). The meteorite has an overall texture that is less equilibrated than many brachinites (e.g., Warren and Kallemeyn 1989; Mittlefehldt et al. 2003; Day et al. 2012; Keil 2014), and is texturally similar to some members of the winonaite meteorite group (e.g., Mount Morris, Winona; Benedix et al. 1998; Corder et al. 2014).

Silicate mineral compositions measured in LEW 88763, 14 confirm the FeO-rich nature of the meteorite (Swindle et al. 1998) with $FeO_{64.0} \pm 0.2$ olivine ($n = 14$), and low-Ca pyroxene ($Wo_{4.3} \pm 1.3En_{67.2} \pm 1.0$; $n = 7$) and high-Ca pyroxenes ($Wo_{39.0} \pm 0.5En_{45.6} \pm 0.2$; $n = 6$) with a high ferrosillite component, with olivine and pyroxene compositions being approximately in equilibrium (Table 2; Fig. 3). These average two pyroxene compositions correspond to an equilibration temperature of $\sim 1080 \pm 50$ °C (Anderson et al. 1993).

phases (~3%) including chromite (1.3%), FeNi metal (0.5%), and troilite (1.3%). No phosphates were observed in LEW 88763, 14.

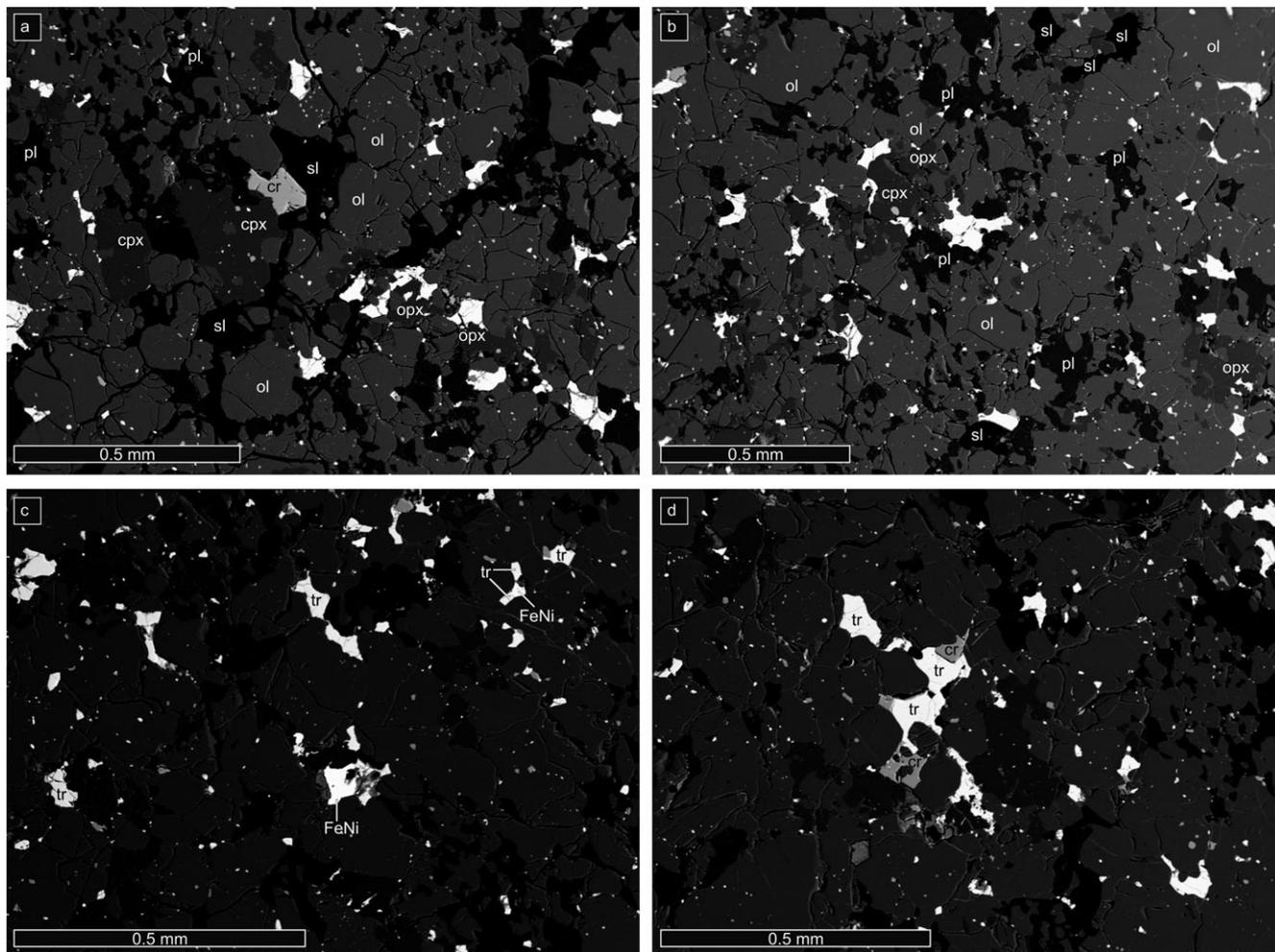


Fig. 2. Back-scatter electron images of regions of LEW 88763, 14, illustrating the presence of interstitial orthopyroxene (opx), clinopyroxene (cpx), plagioclase (pl), chromite (cr), troilite (tr), and metal (FeNi). Lower images (c) and (d) are darkened to emphasize distribution of FeNi metals and troilite in the meteorite.

Mineral compositions in LEW 88763, 14 for feldspar ($An_{30.1 \pm 6.2}Ab_{66.5 \pm 5.4}$; $n = 15$), Cr-spinel ($Cr\# = 81.0 \pm 0.9$; $n = 16$), taenite (44.2 ± 1.2 wt% Fe, 53.1 ± 0.8 wt% Ni; $n = 14$), and troilite (0.9 wt% Ni; $n = 15$; Table 3) are similar to mineral compositions measured in LEW 88763, 11 (Swindle et al. 1998).

Bulk-Rock Major and Trace Element Composition

In addition to reported values of major, minor, and trace elements in LEW 88763 by Nakamura and Morikawa (1993) and Swindle et al. (1998), we report 48 elemental abundances for LEW 88763, 20 (Table 4). The new data are generally consistent with previously published data (Fig. 4), with the exception of some of the rare earth elements (REE), Sr, Se, and Ni measured by Swindle et al. (1998). Differences in Se and Ni

probably reflect different proportions of metal and sulfide in portions of the meteorite that were measured (cf. ~3% versus ~6% modal opaques in LEW 88763, 14 and, 11; Table 1) whereas, the new REE data form a coherent, flat, CI chondrite-like pattern and are more similar to isotope dilution REE abundances measured by Nakamura and Morikawa (1993). Our analysis of LEW 88763 indicates a positive Eu anomaly ($Eu^*/Eu = 1.11$), whereas Nakamura and Morikawa (1993) measured a negative Eu anomaly, indicating uneven distribution of plagioclase \pm any phosphate in the meteorite. This is consistent with both the uneven distribution of plagioclase in LEW 88763 (Table 1) and the presence of phosphate in LEW 88763, 11 (Swindle et al. 1998), but its absence in LEW 88763, 14.

Abundances of Al, Fe, Mg, Ca, P, and Cr in LEW 88763 are similar to carbonaceous chondrite Allende,

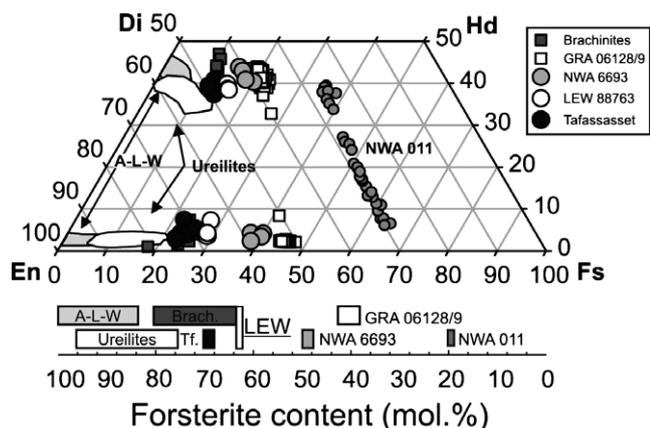


Fig. 3. Pyroxene quadrilateral and olivine forsterite content diagrams for LEW 88763 versus primitive achondrites including: acapulcoite-lodranite-wononaites (A-L-W), ureilites, brachinites (Brach.), Tafassasset (Tf.), NWA 6693, the GRA 06128/9 meteorites, and NWA 011. Other meteorite data from Benedix et al. (1998), Mittlefehldt et al. (1998), Yamaguchi et al. (2002), Floss et al. (2005), Day et al. (2012), Gardner-Vandy et al. (2012), and Warren et al. (2013).

whereas Na and Mn abundances are factors of approximately two to three times higher, and Co and Ni contents are a factor of approximately two lower, than in chondrites. The Mg/(Mg+Fe) ratio of LEW 88763 is 0.39 and is within the range of Mg/(Mg+Fe) for bulk samples of Allende (0.38–0.41; Table 4). These results are similar to those obtained by Swindle et al. (1998), who argued for limited differentiation of a chondritic precursor to produce LEW 88763. Some volatile and moderately volatile elements show strong depletions relative to CI chondrite compositions, including Pb, Cs, Sn, Ge, and Zn.

Highly Siderophile Element Abundances and Re-Os Isotopes

Highly siderophile elements (HSE: Os, Ir, Ru, Pt, Pd, Re) are within the lower range of chondrites and within approximately a factor of two of the abundances measured in CI chondrite Orgueil (Table 5; Fig. 5). Slight depletions, relative to Os, Ir, and Ru, are evident for Pt, Pd, and Re. The measured $^{187}\text{Os}/^{188}\text{Os}$ for LEW 88763 is similar to carbonaceous chondrites, including Allende, at 0.1262 (Fig. 5b). However, the low measured $^{187}\text{Re}/^{188}\text{Os}$ (0.26) leads to supra-chondritic initial $^{187}\text{Os}/^{188}\text{Os}$, assuming an ancient crystallization age >4.5 Ga (e.g., Swindle et al. 1998; assumed at 4.56 in Table 5), implying recent disturbance of the Re-Os isotope system. Consequently, calculated [Re*] concentration, based on the assumption of a chondritic initial $^{187}\text{Os}/^{188}\text{Os}$ at the time of sample formation, is higher than measured Re in LEW 88763.

Oxygen Isotopes

New oxygen isotope values obtained for LEW 88763 (Table 6; Fig. 6) are consistent with values reported in previous studies indicating $\Delta^{17}\text{O}$ of -1.14 to -1.26‰ , with a mean of $-1.19 \pm 0.10\text{‰}$ based on six analyses of the meteorite from this study, Clayton and Mayeda (1996), and Greenwood et al. (2012). The oxygen isotope systematics of LEW 88763 are, therefore, most consistent with the oxygen isotope signatures of the acapulcoite-lodranite clan ($\delta^{18}\text{O} = 3.4 \pm 0.8\text{‰}$, $\delta^{17}\text{O} = 0.7 \pm 0.5\text{‰}$, $\Delta^{17}\text{O} = -1.1 \pm 0.4\text{‰}$; 2σ ; data from Clayton and Mayeda 1996), and with NWA 6693 ($\Delta^{17}\text{O} = -1.08\text{‰}$; Warren et al. 2013).

DISCUSSION

Effects of Weathering on LEW 88763

Lewis Cliff 88763 shows indications of terrestrial weathering, with minor alteration of some sulfide grains and the effect of frictional heating, leading to a concentric discoloration to the interior of the meteorite (Fig. 1). LEW 88763 has been shown to contain no excess ^{129}Xe , but has a K-Ar age ≥ 4.5 Ga and a cosmic-ray exposure age of 26–28 Ma (Swindle et al. 1998); the residence time on Earth is unconstrained. Assuming derivation from a reservoir with uniform initial $^{187}\text{Os}/^{188}\text{Os}$, it would be anticipated that LEW 88763 would plot near a primordial $^{187}\text{Re}-^{187}\text{Os}$ isochron (e.g., Smoliar et al. 1996), as is the case for the portions of Allende that were measured (Fig. 5b). However, this is not true, and, instead, LEW 88763 lies to the left of the isochron, with chondritic measured $^{187}\text{Os}/^{188}\text{Os}$ unsupported by low Re/Os. This non-isochronous behavior is similar to that observed for some ureilites (Rankenburg et al. 2007), brachinites, and brachinite-like achondrites (Day et al. 2012). In the case of brachinites and brachinite-like achondrites, it has been demonstrated that low Re/Os reflects mobility of Re within the meteorites, due to terrestrial alteration (Day et al. 2012; Hyde et al. 2014). Rhenium, rather than Os mobility with the meteorites, is indicated by the variability in Re concentrations and Re/Os measured among different splits of achondrite meteorites (e.g., Zag [b]), from the alteration of sulfide and metal. Consequently, this would mean portions of LEW 88763 are likely to have high Re/Os due to addition of Re, and both high and low Re/Os has been observed in aliquots of ureilite and brachinite meteorites (Rankenburg et al. 2008; Day et al. 2012).

The susceptibility of sulfides and metals—the primary hosts of the HSE in the studied samples—to terrestrial aqueous alteration is well documented in

Table 3. Average compositions of metal and sulfide mineral phases (in wt%) in LEW 88763.

		Si	S	Fe	Co	Ni	P	Mn	Mg	Al	Ti	Total	Ni/Co	Fe/Co	Fe/S
FeNi	Av.	0.05	0.02	44.1	1.79	53.1	<0.03	0.01	0.02	0.17	<0.03	99.3	29.7	24.7	–
<i>n</i> = 15	2 σ	0.03	0.01	1.22	0.05	0.76	–	0.01	0.03	0.09	0.01	0.8	1.03	0.96	–
FeS	Av.	0.01	36.1	62.1	0.14	0.90	<0.03	0.01	<0.03	<0.03	<0.03	99.2	–	–	1.72
<i>n</i> = 15	2 σ	0.01	0.24	0.63	0.06	0.57	–	0.01	–	–	–	0.4	–	–	0.04

chondrites (e.g., Horan et al. 2010) and sulfide- and metal-rich achondrites (e.g., Shearer et al. 2010; Day et al. 2012; Hyde et al. 2014). However Os, Ir, Ru, Pt, and Pd do not appear to have been modified within the meteorite. From our measurements and those of Swindle et al. (1998), U is anomalously elevated in LEW 88763 relative to CI chondrite suggesting potential modification of U through terrestrial alteration. For other elements, there is no clear indication for modifications from terrestrial alteration, such as during hot-desert alteration, where LREE modification can be particularly problematic (e.g., Barrat et al. 1999; Crozaz et al. 2003). For example, Sr and Ba are in chondritic abundances in LEW 88763 and there is no pronounced Ce-anomaly. Therefore, it is assumed that other geochemical systematics are also unaffected by weathering in the meteorite.

Partial Melting and Metamorphic Constraints on LEW 88763

Lewis Cliff 88763 experienced limited melt loss and essentially represents a metamorphosed and anatectic-chondritic composition. Pyroxene thermometry indicates that temperatures >1000 °C were reached during formation, consistent with the near-absence of structures indicative of accreted nebular components. For example, relict chondrules were not observed in the study of Swindle et al. (1998), have been observed by Gardner-Vandy (2012), and some oval structures in LEW 88763, 14, which are ≤ 1 mm in diameter, could be tentatively assigned as relict chondrules (Fig. 1). These observations suggest temperatures sufficient to induce Fe-Ni-S melting (~988 °C) and, potentially, enabled the onset of silicate melting (>1050 °C; e.g., Jurewicz et al. 1991, 1995; Feldstein et al. 2001; Gardner-Vandy et al. 2013; Usui et al. 2015), but may not have completely destroyed relict structures.

Modeling of the REE abundances reveals extremely limited partial melt loss (0 to <<5%) to explain the composition of LEW 88763 (Fig. 7). The uncertainty on the estimate of partial melt loss comes from the choice of model parameters and starting compositions. It is therefore likely that any onset of silicate melting led to localized redistribution of silicate phases, rather than extraction of a significant silicate melt component. This

is consistent with the REE pattern for LEW 88763 that is nearly flat and similar to CI chondrites and the Brachina meteorite. In contrast, some other FeO-rich partially melted achondrites (e.g., NWA 6693; Tafassasset) appear to be REE-depleted cumulates or strongly depleted melt residues, rather than metamorphosed chondritic compositions (Fig. 7).

The HSE are sensitive monitors of Fe-Ni-S melt loss and parameterization has been produced to describe HSE fractionation during differentiation of a chondritic composition (Chabot and Jones 2003). Fe-Ni-S melt loss, while evident in LEW 88763, to describe the factor of approximately two lower absolute siderophile (e.g., Se, Ni, Co) and HSE abundances compared with chondrites, was not extensive and did not greatly fractionate Pd/Os, Ir/Os, or Pt/Os (Fig. 8). These ratios in LEW 88763 are remarkably similar to carbonaceous chondrites, especially CR chondrites, which are characterized by typically higher Ir/Os and lower measured $^{187}\text{Os}/^{188}\text{Os}$ than in enstatite or ordinary chondrites (e.g., Walker et al. 2002; Horan et al. 2003; Fischer-Gödde et al. 2010). Modeling illustrates that, at most, a few percent of Fe-Ni-S melt loss was experienced by LEW 88763 during anatexis.

A Carbonaceous Chondrite Precursor Composition

LEW 88763 is a residue after limited partial melting, similar to the origin of some winonaites (e.g., Benedix et al. 1998), with which it shares strong textural similarities. On the other hand, LEW 88763 has lower Mg/(Mg+Fe) compared with winonaites and, while similar to brachinite meteorites, it has a more negative $\Delta^{17}\text{O}$ value than any of these meteorite groups. The low $\Delta^{17}\text{O}$ (–1.19‰) and $^{187}\text{Os}/^{188}\text{Os}$ (0.1262) values of LEW 88763 also preclude possible genetic links with ordinary ($\Delta^{17}\text{O} = >0\text{‰}$; $^{187}\text{Os}/^{188}\text{Os} = \sim 0.128$) or enstatite chondrite groups ($\Delta^{17}\text{O} = \sim 0\text{‰}$; $^{187}\text{Os}/^{188}\text{Os} = \sim 0.128$). Many aspects of the chemistry of LEW 88763 indicate that the chondritic source involved in its genesis was similar to carbonaceous chondrites. The low Mg/(Mg+Fe) of LEW 88763 (0.39) is similar to the CV3 carbonaceous chondrite Allende. Interelement ratios of the HSE, including long-term Re/Os, as measured by $^{187}\text{Os}/^{188}\text{Os}$, are similar to carbonaceous chondrites. Finally, the oxygen isotope systematics of LEW 88763

Table 4. Major- and trace-element abundances determined in LEW 88763, Allende and terrestrial standards.

Sample	LEW 88763	Allende UCSD	Allende Literature	BHVO-2 (<i>n</i> = 3)			BCR-2 (<i>n</i> = 3)			BIR-1 (<i>n</i> = 3)			TPB (<i>n</i> = 4)	
				Sub-ID	Av.	RSD	Rec.	Av.	RSD	Rec.	Av.	RSD	Rec.	Average
Al	12849	13627	17359	71654	2%	71600	71141	81813	82034	6%	82034	112	45%	
Fe	225000	204000	221000	86583	2%	86300	96387	96500	79036	5%	79036	4290	6%	
Mn	2394	1254	1490.0	1294	2%	1290	1517	1520	1355	2%	1355	9.6	11%	
Mg	147000	123000	154000	43513	2%	43600	21589	35900	58494	2%	58494	87	45%	
Ca	16197	14684	20000	82023	2%	81700	51230	50900	95055	2%	95055	1202	9%	
Na	5988	3118	3400	15947	2%	16400	22760	23400	13502	3%	13502	360	8%	
Cr	3704	2928	3612	277	2%	280	25	18	370	4%	370	13	5%	
Co	429.7	583.4	591	44.4	3%	45	31.9	37.0	52	3%	52	1.1	11%	
Ni	6660	11650	13660	120.6	3%	119	17.8	167.6	170	7%	170	13.4	16%	
Li	2.26	1.46	1.40	4.69	3%	4.80	9.11	9.00	3.60	2%	3.60	0.017	42%	
P	1018	937	1067	1168	3%	1200	1450	1500	100	4%	100	57.076	8%	
Sc	9.1	7.9	11.0	31.8	4%	32	31.0	43.1	44	6%	44	0.081	6%	
Ti	615	570	1000	16339	5%	16300	12580	5573	310	6%	310	0.450	110%	
V	76.9	58.9	82.4	317.4	6%	317	387.3	416	125	7%	125	0.350	130%	
Cu	80.5	88.9	97.0	128.1	5%	127.0	19.2	19.0	70	8%	70	0.312	62%	
Zn	65.4	74.9	106.0	104.1	5%	103.0	122.0	127.0	16	8%	16	1.370	13%	
Ga	6.04	4.85	5.90	22.10	5%	22.00	21.26	23.00	18	5%	18	0.003	60%	
Ce	2.38	5.74	17.90	1.64	6%	1.60	1.66	1.22	0.60	8%	0.60	0.011	17%	
Se	0.52	0.46	1.32	0.11	2%	0.10	0.11	0.02	0.043	4%	0.043	0.005	10%	
Rb	1.81	1.16	1.81	9.95	3%	9.11	47.14	48.00	0.021	3%	0.021	0.021	69%	
Sr	12.5	13.4	15.7	403.8	3%	396.0	323.1	346.0	110	3%	110	0.021	69%	
Y	2.15	2.12	2.63	25.83	3%	26.00	33.04	37.00	16	3%	16	0.001	121%	
Zr	6.34	5.60	6.77	170.3	3%	172.0	173.7	188.0	0.60	3%	0.60	0.062	57%	
Nb	1.97	1.72	1.45	3.71	28%	4.00	232.36	248.00	0.038	2%	0.038	0.026	36%	
Mo	0.48	0.64	0.30	1.84	6%	1.70	2.30	0.80	0.277	7%	0.277	0.026	36%	
Sn	0.062	0.076	0.098	0.10	3%	0.10	1.07	0.01	0.001	3%	0.001	0.001	117%	
Cs	4.30	4.27	4.74	139.21	2%	131.00	653.40	683.00	7.00	2%	7.00	0.033	71%	
Ba	0.361	1.10	0.490	15.13	2%	15.20	22.96	25.00	0.63	1%	0.63	0.015	167%	
La	0.953	5.44	1.24	37.20	2%	37.50	48.88	53.00	1.90	1%	1.90	0.035	169%	
Ce	0.136	0.213	0.192	5.33	3%	5.35	6.33	6.80	0.36	3%	0.36	0.0008	156%	
Pr	0.671	1.071	0.967	24.35	3%	24.50	26.46	28.00	2.50	3%	2.50	0.0028	156%	
Nd	0.221	0.281	0.313	6.04	3%	6.07	6.04	6.70	1.10	1%	1.10	0.0001	42%	
Sm	0.089	0.090	0.115	2.06	3%	2.07	1.94	2.00	0.55	2%	0.55	0.0001	63%	
Eu	0.272	0.366	0.401	6.21	2%	6.24	6.38	6.80	1.80	2%	1.80	0.0005	145%	
Gd	0.051	0.062	0.072	0.92	3%	0.93	0.96	1.07	0.0001	2%	0.0001	0.0001	104%	
Tb	0.368	0.411	0.479	5.29	3%	5.31	5.92	6.40	4.00	4%	4.00	0.0002	87%	
Dy	0.081	0.081	0.100	0.97	3%	0.98	1.19	1.33	0.00003	1%	0.00003	0.00003	94%	
Ho	0.241	0.241	0.299	2.53	3%	2.54	3.38	3.50	0.0001	1%	0.0001	0.0001	112%	
Er	0.037	0.044	0.053	0.33	3%	0.33	0.48	0.54	0.00003	2%	0.00003	0.00003	107%	
Tm	0.245	0.267	0.319	1.99	3%	2.00	3.10	3.50	0.26	2%	0.26	0.0001	120%	
Yb	0.037	0.038	0.046	0.27	3%	0.27	0.45	0.51	0.60	4%	0.60	0.0003	103%	
Lu	0.168	0.158	0.188	4.34	3%	4.36	4.49	4.80	0.0012	1%	0.0012	0.0012	85%	
Hf	0.027	0.027	0.033	1.13	2%	1.14	0.71	0.04	0.0012	2%	0.0012	0.0012	48%	
Ta	0.217	0.253	0.200	0.21	6%	0.21	0.42	0.10	0.070	3%	0.070	0.014	21%	
W	0.139	1.286	1.520	1.76	4%	1.60	8.95	11.00	3.00	2%	3.00	0.014	49%	
Pb	0.048	0.057	0.078	1.24	5%	1.22	5.45	6.20	0.001	1%	0.001	0.001	77%	
Th	0.220	0.0123	0.0174	0.41	3%	0.40	1.51	1.69	0.004	2%	0.004	0.004	25%	

Published and recommended data for Allende are from Jarosewich et al. (1987) and Stracke et al. (2012), and for BHVO-1, BCR-2, and BIR-1 are from GeoREM. TPB = total procedural blank.

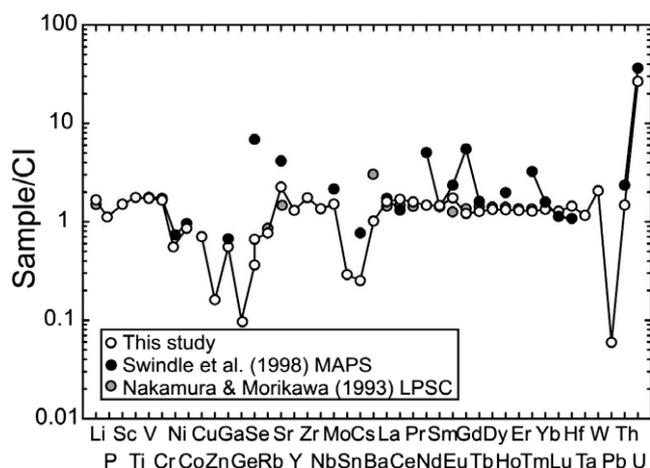


Fig. 4. CI chondrite normalized multi-element diagram for LEW 88763, 20, measured in this study relative to aliquots of LEW 88763 measured in previous studies (Nakamura and Morikawa [1993] and Swindle et al. [1998]). CI chondrite normalization is from McDonough and Sun (1995). Note, the normalization value for U is $0.0074 \mu\text{g g}^{-1}$, whereas measured U in the Allende CV3 chondrite from this study was $\sim 60\%$ higher ($0.0123 \mu\text{g g}^{-1}$).

are within the range of CR, CH, CB, and possibly CV-CK, or CM carbonaceous chondrites groups (e.g., Clayton and Mayeda 1996). For example, comparison of the elemental abundances of LEW 88763 ordered by condensation temperature, illustrates that it is similarly volatile- and moderately volatile element-rich compared with Allende, as well as with other FeO-rich partially melted achondrite meteorites (Fig. 9). The preponderance of evidence is that LEW 88763 represents a metamorphic product from a carbonaceous chondrite source composition, with the strongest existing similarities to CR chondrites.

Implications for Melting Processes in FeO-Rich Parent Bodies

Due to the differing compatibility of Fe and Mg during partial melting, Mg is generally retained in the residue, relative to Fe. This simple relationship is a foundation for the fundamental understanding of melting processes in planetary bodies and explains the lower $\text{Mg}/(\text{Mg}+\text{Fe})$ in melts relative to (mantle) source residues in Earth, Mars, or the Moon (e.g., Fig. 10). The observation of similar $\Delta^{17}\text{O}$ values between brachinites and the GRA 06128/9 oligoclase-rich achondrites has led to models of the generation of olivine-rich residues and feldspar-rich melts during 13–30% partial melting of a FeO-rich parent body (Day et al. 2009, 2012). Similar relationships have been invoked for FeO-rich achondrite Tafassasset and the CR chondrites, although they cannot be directly related by partial melting (Gardner-Vandy

et al. 2012), with the “eucrite-like” meteorite NWA 011 perhaps representing an example of a potential extrusive melt product of such a system (Fig. 10).

As noted by Warren et al. (2013), LEW 88763 and NWA 6693 share similar oxygen isotope characteristics, with LEW 88763 representing an achondrite that experienced limited partial melting, whereas NWA 6693 is interpreted as an igneous cumulate. It is impossible, however, for NWA 6693 to be a direct melt product of LEW 88763. This is because LEW 88763 has experienced partial melting that is too limited to generate a NWA 6693-like composition, and instead, LEW 88763 represents the least-modified FeO-rich achondrite source composition recognized to date. Thus, while NWA 6693 and LEW 88763 may originate from the same (genetic) parent body, they indicate that complex melting processes are potentially exhibited on individual parent bodies. For example, the origin of LEW 88763 and NWA 6693 from the same parent body can potentially be reconciled if melting was restricted due to relatively late parent-body construction as ^{26}Al was almost extinct, or due to impact disruption (Rubin 2007). A possible scenario includes the variable heating of an asteroidal parent body sufficient to lead to nascent (sulfide-rich?) core formation, partial-melt residues, and high Fe/Mg cumulates and melts, but equally to leave portions of the asteroid partially heated and metamorphosed or, potentially, completely unmetamorphosed. Many aspects of models for melting processes on asteroidal parent bodies appear to be common to both FeO-rich, as well as more reduced partially melted achondrites.

Whatever the exact relationship of FeO-rich primitive achondrites to one another, the identification of LEW 88763 as a metamorphosed and anatectic melt from a carbonaceous chondrite source has some important implications for general models of planetary differentiation. First, the distinction between chondritic and achondritic materials subjected to metamorphism and anatexis becomes yet more diffuse. Lewis Cliff 88763 has chemical similarities to some carbonaceous chondrites, yet the mineralogy of LEW 88763 is dominated by olivine. The relationship is similar to winonaites, which are interpreted as metamorphosed chondrites and have high $\text{Mg}/(\text{Mg}+\text{Fe})$ (Benedix et al. 1998). Second, the possible relationship between LEW 88763 and NWA 6693, suggested from $\Delta^{17}\text{O}$, indicates that yet another FeO-rich asteroidal parent body generated Fe-rich partial melts on an asteroid (e.g., Day et al. 2009, 2012; Gardner-Vandy et al. 2012). It is remarkable that an increasing number of FeO-rich achondrites are being recognized with variable and low $\Delta^{17}\text{O}$ values. Given that carbonaceous chondrites have generally lower $\Delta^{17}\text{O}$ values than ordinary or enstatite chondrites, this relationship implies that asteroidal

Table 5. Rhenium-osmium isotopes and highly siderophile element abundances (in ng g^{-1}) in LEW 88763 and Allende.

Sample	Mass (g)	Os	Ir	Ru	Pt	Pd	Re	Re*	$^{187}\text{Re}/^{188}\text{Os}$	$\pm 2\sigma$	$^{187}\text{Os}/^{188}\text{Os}$	$\pm 2\sigma$	$^{187}\text{Os}/^{188}\text{Os}_i$	$\pm 2\sigma$	ΔDOs^1
LEW 88763, 20	0.059	371.3	399.5	541.5	588.6	388.8	20.94	32.98	0.272	0.007	0.12620	0.00007	0.1047	0.0010	95.3
Allende	0.101	737.6	717.7	1030.9	1502.7	637.7	60.31	64.76	0.394	0.015	0.12586	0.00007	0.0947	0.0018	-4.6
	0.101	666.4	657.4	994.9	1384.7	627.2	54.54	58.55	0.394	0.014	0.12588	0.00007	0.0947	0.0017	-4.6
Literature data ²															
Allende ^a		729	684	1007	1357	674	60.5	59.9	0.393		0.12595				-2.5
Allende ^b		763	700	1140	1379	786	61.2	0.386			0.12638				7.0
Allende ^c		785	720	1118	1421	682	63.5	0.390			0.12596				-0.1
Allende ^d		758	712	1016	1348	674	60.9	0.387			0.12615				3.6
Allende ^e		746	730	843	1290		60.2								
Allende ^f			694	1058	1395	598									
Allende ^g		833	776			705	68.4								
Allende ^h			740												
LEW 88763 ⁱ		490	445												

¹ ΔDOs (after Fischer-Gödde et al. 2010) refers to the combined deviation in the determined $^{187}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Re}/^{188}\text{Os}$ for a given sample from the IIIA iron meteorite reference isochron, $\Delta\text{DOs} = 10^4 (^{187}\text{Os}/^{188}\text{Os}_{\text{chondrite}} [0.09524 + 0.07887 \times ^{187}\text{Re}/^{188}\text{Os}_{\text{chondrite}}], ^{187}\text{Os}/^{188}\text{Os}_{\text{chondrite}})$, and $^{187}\text{Re}/^{188}\text{Os}_{\text{chondrite}}$ are the values determined for chondrites, 0.09524 is the initial $^{187}\text{Os}/^{188}\text{Os}$ and 0.07887 is the slope of the IIIA iron meteorite isochron (Smoliar et al. 1996).

²a = [ID CT/HPA] Fischer-Gödde et al. (2010); b = [ID CT] Becker et al. (2006); c = [ID CT] Brandon et al. (2005); d = [ID CT] Walker et al. (2002) and Horan et al. (2003); e = [Spark source mass spectrometry] Jochum (1996); f = [NiS] Tagle and Berlin (2008); g = [RNAA] Takahashi et al. 1978; h = [recommended values] Jarosewich et al. (1987); i = [Average NAA] Swindle et al. (1998); ID = isotope dilution; CT = Carius tube; HPA = high pressure ashier; NiS = nickel sulfide fire assay; RNAA = radiochemical neutron activation analysis; NAA = neutron activation analysis.

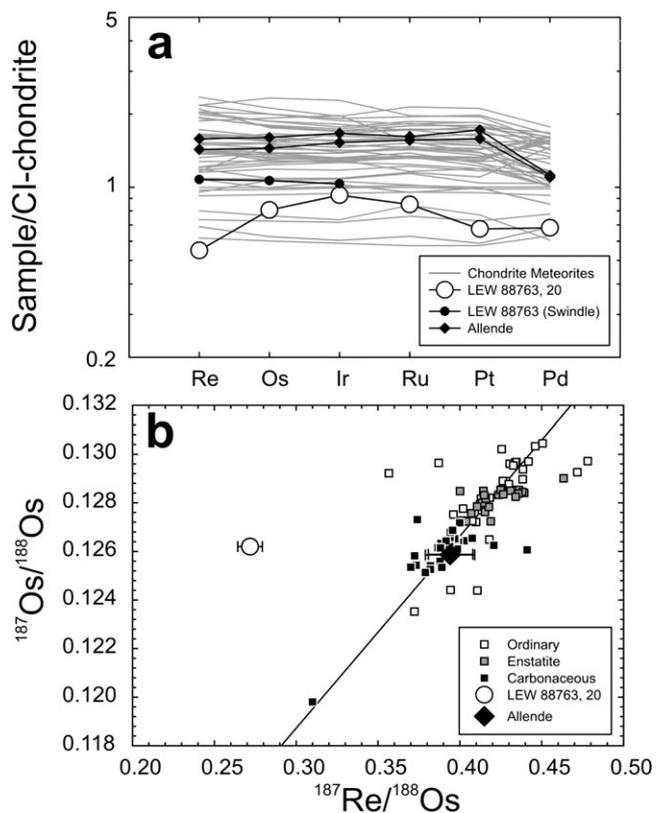


Fig. 5. a) CI chondrite normalized highly siderophile element abundances for LEW 88763 (this study and Swindle et al. 1998) and Allende versus chondritic meteorites and b) Re-Os isotope diagram showing the 4.568 Ga IIIAB iron reference isochron (Smoliar et al. 1996). Chondrite normalization and chondrite data from Horan et al. (2003).

Table 6. Oxygen isotope composition of LEW 88763.

Sample type	$\delta^{18}\text{O}$ (‰)	$\delta^{17}\text{O}$ (‰)	$\Delta^{17}\text{O}$ (‰)	Reference and remarks
LEW 88763, 20	2.389	0.115	-1.136	This study
Repeat	2.380	0.111	-1.135	
Repeat	2.449	0.061	-1.221	
Average	2.406	0.096	-1.164	
2 SD	0.075	0.060	0.099	
LEW 88763	2.47	0.10	-1.19	Untreated, 2 replicates (Greenwood et al. 2012)
LEW 88763	2.37	-0.03	-1.26	Clayton and Mayeda (1996)
Grand mean	2.41	0.07	-1.19	This study,
2 SD	0.08	0.11	0.10	Greenwood et al. (2012), Clayton and Mayeda (1996)

parent bodies sampled as meteorites today, were constructed from a range of chondritic precursor compositions. Alternatively, evidence for generation of

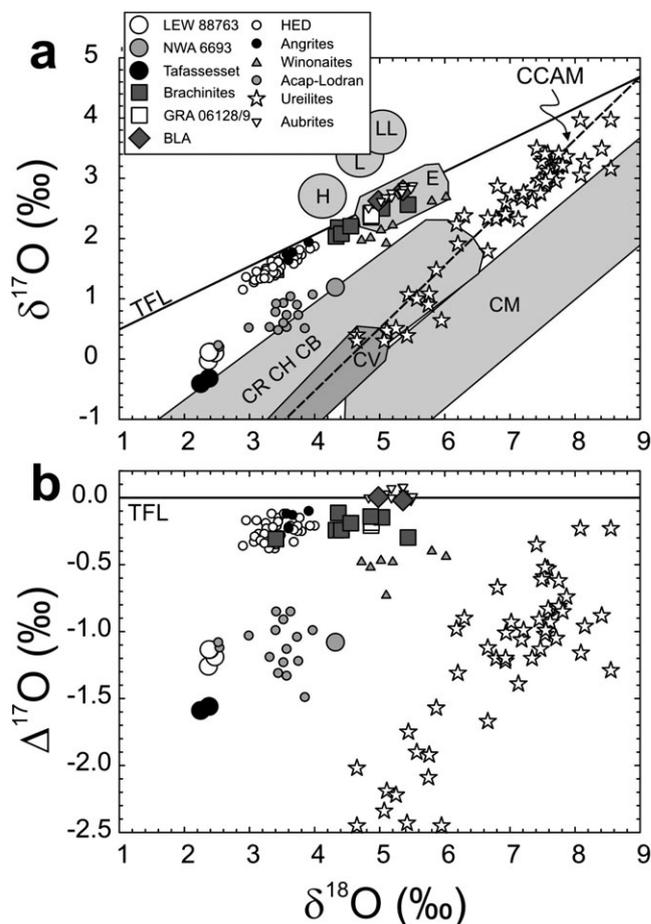


Fig. 6. Oxygen isotope systematics of LEW 88763 compared with other primitive achondrites plotted as (a) $\delta^{18}\text{O}$ versus $\delta^{17}\text{O}$ and (b) $\delta^{18}\text{O}$ versus $\Delta^{17}\text{O}$. BLA = brachinite-like achondrites; HED = howardite-eucrite-diogenites; Acap-Lodran = acapulcoite-lodranites. Terrestrial fractionation line (TFL; $\delta^{17}\text{O} = 0.524 \times \delta^{18}\text{O}$); carbonaceous chondrite anhydrous mineral like (CCAM; $\delta^{17}\text{O} = 0.941 \times \delta^{18}\text{O} - 4$); and ordinary (H, L, LL), enstatite (E), and carbonaceous chondrite (CM, CV, CR, CH, CB) fields from Clayton (2010). CI chondrite compositions plot off the top right of (a). Published data are from Clayton and Mayeda (1996), Day et al. (2009, 2012), Greenwood et al. (2012), Gardner-Vandy et al. (2012), Warren et al. (2013), and references therein.

Fe-rich melts and Mg-rich residues on asteroidal parent bodies with variable FeO, and hence variable oxygen fugacity and initial source compositions, indicates that partial melting and Fe-rich melt segregation is a common process associated with differentiation initiation on all planetary bodies.

A Note on the Classification of LEW 88763

Despite long-known and extensive evidence to the contrary (e.g., Swindle et al. 1998), LEW 88763 remains classified in the Meteoritical Bulletin Database as a

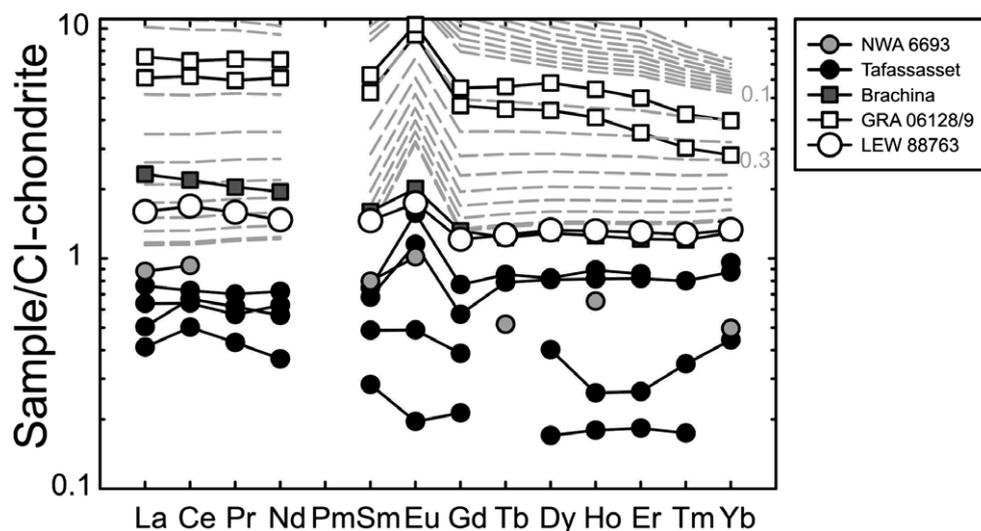


Fig. 7. CI chondrite rare earth element diagram for LEW 88763 versus REE patterns obtained for GRA 06128/9 and Brachina (Day et al. 2009), the average composition of NWA 6693 (Warren et al. 2013), and Tafassasset (Gardner-Vandy et al. 2012; Göpel et al. 2015). Also shown is a CI chondrite melting model (fig. 12 of Day et al. 2012) illustrating 13–30% partial melting required to explain the GRA 06128/9 REE compositions. For LEW 88763, very limited silicate melting (0 to <<5%) is required to explain the heavy REE compositions. Model uses modal proportions of olivine, orthopyroxene, clinopyroxene, and plagioclase of 0.67/0.18/0.04/0.11 and subequal melting of phases, up to plagioclase exhaustion.

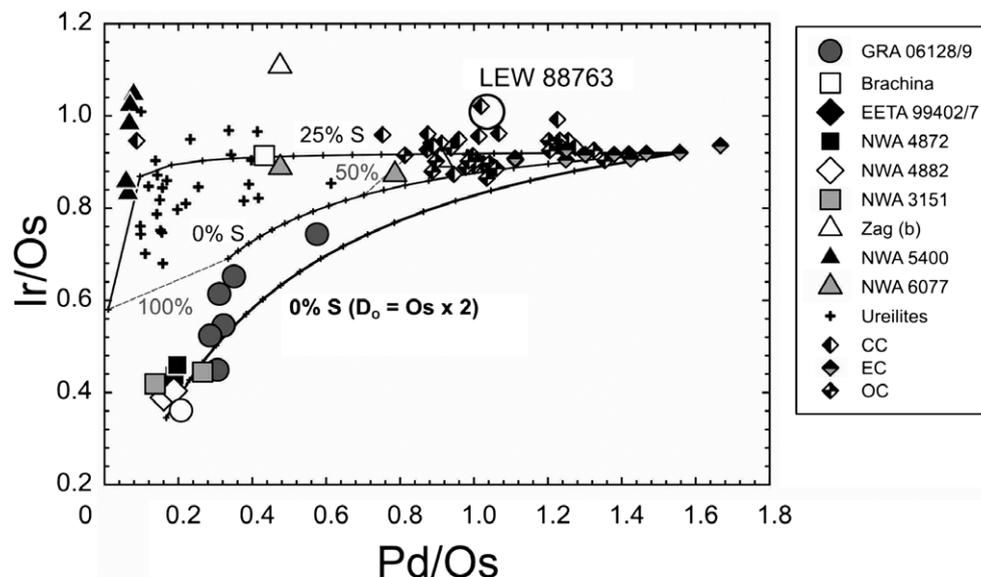


Fig. 8. Plot of Pd/Os versus Ir/Os for LEW 88763 with calculated compositions of metal residues. Also shown are chondrite (Horan et al. 2003; Fischer-Gödde et al. 2010), GRA 06128/9, brachinite, brachinite-like achondrite (Day et al. 2009, 2012), and ureilite HSE compositions (Rankenburg et al. 2008). The melt calculations model the composition of residues that result from single episodes of batch melting. The starting HSE composition is the bulk composition of Orgueil with concentrations adjusted assuming that all of the HSE were originally in metal and that metal comprises 5% of the bulk. Curves show compositions of residues resulting from no sulfur, and 25% sulfur. Fractions of residue are labeled and are in increments of 5%. Solid metal-liquid metal D_o values were calculated using the parameterization of Chabot and Jones (2003). The thick curve denotes D_o for Os that is two times greater than parameterized values required to explain brachinite and GRA 06128/9 Pd/Os and Ir/Os. LEW 88763 has Pd/Os, Ir/Os, and Pt/Os (data not shown) similar to some carbonaceous chondrites, notably CR chondrites (e.g., EET 92042). CC = carbonaceous chondrites; EC = enstatite chondrites; OC = ordinary chondrites. Details of the models are provided in Day et al. (2012).

brachinite or Brachina-like meteorite, with a note that “the meteorite is an achondrite, very similar to Brachina in texture and mineral compositions.” However,

observations of a texture similar to winonaites; nearly chondritic refractory elemental abundances; and oxygen isotope compositions strongly dissimilar to brachinites

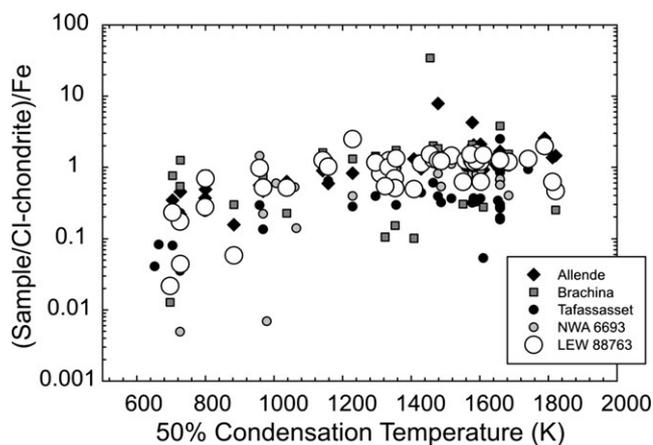


Fig. 9. 50% condensation temperature (K) from Lodders (2003) versus elemental concentration in LEW 88763, Allende (this study), Brachina (Day et al. 2012), Tafassasset (Gardner-Vandy et al. 2012), and NWA 6693 (Warren et al. 2013) double-normalized to CI chondrite (McDonough and Sun 1995) and Fe content.

($\Delta^{17}\text{O} = \sim -0.2$; Day et al. 2012) demand that LEW 88763 should be reclassified, likely as an anomalous achondrite that experienced limited metal/sulfide loss. While further work to elucidate Cr and Ti isotopic composition in LEW 88763 may provide a definitive link with other achondrite meteorites, including NWA 6693 (Warren et al. 2013), the message from these meteorites are that common processes are evident between parent bodies formed from different initial source compositions.

CONCLUSIONS

Achondrite meteorite Lewis Cliff (LEW) 88763 is an olivine-dominated stone containing ~6 modal % pyroxene; ~10% modal plagioclase; and ~3–6 modal % chromite, FeNi metal, and troilite. Exposure to the terrestrial atmosphere led to a variable thickness fusion crust, as well as concentric alteration of the interior of the meteorite from the fusion crust. The strongest evidence for recent terrestrial modification is the low Th/U and Re/Os of the meteorite, implying mobilization of U and Re within the meteorite during residence on the Antarctic ice. Achondrite LEW 88763 has something of a “split-personality,” having (1) textures similar to winonaites; (2) a low $\Delta^{17}\text{O}$ value (-1.19‰) and a low bulk-rock Mg/(Mg+Fe) ratio (0.39), similar to the FeO-rich cumulate Northwest Africa (NWA) 6693; and (3) near-chondritic abundances of many major, minor, and trace elements, including the rare earth elements. The similar bulk-rock major, minor, and trace element abundances of LEW 88763, relative to some carbonaceous chondrites (e.g., Allende),

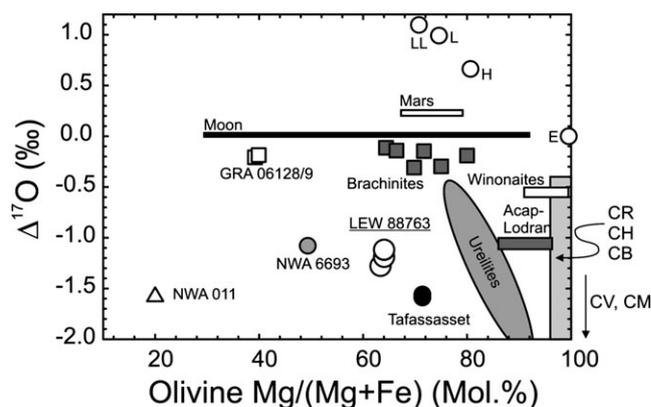


Fig. 10. Forsterite composition versus $\Delta^{17}\text{O}$ diagram. Forsterite contents in lunar, terrestrial, and martian igneous products reflect residues or cumulates (typically, high Mg/[Mg+Fe]), or differentiated melts (typically, lower Mg/[Mg+Fe]) that fall on the same mass-dependent three-oxygen isotope fractionation line. The same relationship, of residues and differentiated melts, respectively, has been suggested for brachinites and the oligoclase-rich meteorites, GRA 06128/9 (Day et al. 2009, 2012). LEW 88763 has similar $\Delta^{17}\text{O}$ to NWA 6693, which is interpreted as a cumulate rock from a ferroan parental composition (Warren et al. 2013), whereas LEW 88763 can be interpreted as a metamorphosed chondrite that experienced limited Fe-Ni-S loss, perhaps offering another potential link between asteroidal melt products and partial residues. Similar relationships are also possible for CR chondrites, Tafassasset, and possibly even the differentiated melt product, NWA 011 (e.g., Yamaguchi et al., 2002; Floss et al. 2005; Gardner-Vandy et al. 2012). Oxygen isotope data for chondrite meteorites are from Clayton (2010) and data for Mars, the Moon, ureilites, winonaites, and acapulcoite-lodranites are from Warren et al. (2013).

including Pd/Os, Pt/Os, Ir/Os, and measured $^{187}\text{Os}/^{188}\text{Os}$ (0.1262) ratios, implies a volatile-rich precursor composition, similar to some carbonaceous chondrite groups. Lack of strong fractionation of the rare earth elements, but a factor of approximately two lower highly siderophile element abundances in LEW 88763, compared with carbonaceous chondrites, implies limited loss of a sulfide melt during metamorphism and anatexis. These results support the generation of high Fe/Mg partial melts from FeO-rich parent bodies during partial melting.

In detail, LEW 88763 cannot be a parent composition to any meteorite recognized in the terrestrial collection, due to limited silicate melt loss (0 to $\ll 5\%$) and an extracted melt that was likely dominated by Fe-Ni-S. This meteorite represents the least-modified FeO-rich achondrite source composition recognized to date in the meteorite collection. Combined with evidence of partial melting processes in FeO-rich meteorites such as brachinites, brachinite-like achondrites, the Graves Nunataks 06128/9 meteorites, Northwest Africa 6693, and Tafassasset, the origin of

LEW 88763 has important implications for the initiation of planetary differentiation. First, regardless of precursor compositions (e.g., ordinary, enstatite, or carbonaceous chondrite source composition), partial melting and differentiation processes appear to be similar on asteroidal bodies spanning a range of initial oxidation states and volatile contents. Second, the range of $\Delta^{17}\text{O}$ in FeO-rich achondrites (~ -0.2 to -1.8 ‰) implies the generation of volatile-rich and oxidized asteroids through the combination of both ^{17}O -poor (carbonaceous) and ^{17}O -rich (ordinary/enstatite) chondritic sources.

Acknowledgments—We thank J.-A. Barrat, K. Gardner-Vandy, and T. Swindle for their constructive reviews and I. Franchi for editorial handling. We are grateful to the Meteorite Working Group for provision of LEW 88763, 14 and 20. US Antarctic meteorites are recovered by the Antarctic Search for Meteorites (ANSMET) program, which has been funded by the NSF and NASA and curated by the Smithsonian Institution and NASA Johnson Space Center. Support for this work from the NASA Cosmochemistry program (NNX12AH75G to JMDD and NNX11AG586 to LAT) is gratefully acknowledged. The authors declare no conflicts of interest.

Editorial Handling—Dr. Ian Franchi

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SUPPORTING INFORMATION

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

Fig S1: Comparison plot of concentrations measured by ICP-MS (this study) versus concentrations measured using neutron activation analysis (Swindle et al. 1998).
