

## Processes involved in the formation of magnesian-suite plutonic rocks from the highlands of the Earth's Moon

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**Abstract.** The earliest evolution of the Moon likely included the formation of a magma ocean and the subsequent development of anorthositic flotation cumulates. This primary anorthositic crust was then intruded by mafic magmas which crystallized to form the lunar highlands magnesian suite. The present study is a compilation of petrologic, mineral-chemical, and geochemical information on all pristine magnesian-suite plutonic rocks and the interpretation of this data in light of 18 "new" samples. Of these 18 clasts taken from Apollo 14 breccias, 12 are probably pristine and include four dunites, two norites, four troctolites, and two anorthosites. Radiogenic isotopic whole rock data also are reported for one of the "probably pristine" anorthositic troctolites, sample 14303,347. The relatively low Rb content and high Sm and Nd abundances of 14303,347 suggest that this cumulate rock was derived from a parental magma which had these chemical characteristics. Trace element, isotopic, and mineral-chemical data are used to interpret the total highlands magnesian suite as crustal precipitates of a primitive KREEP (possessing a K-, rare earth element (REE)-, and P-enriched chemical signature) basalt magma. This KREEP basalt was created by the mixing of ascending ultramafic melts from the lunar interior with urKREEP (the late, K-, REE-, and P-enriched residuum of the lunar magma ocean). The trace and major element compositions of nearly all magnesian-suite cumulates can be generated by 0-55% fractional crystallization of a primitive KREEP basalt combined with the trapping of varied proportions (generally  $\leq 20\%$ ) of instantaneous, residual, KREEP-basalt liquid. A few samples of the magnesian suite with extremely elevated large-ion lithophile elements (5-10x other magnesian-suite rocks) cannot be explained by this model or any other model of autometasomatism, equilibrium crystallization, or "local melt-pocket equilibrium" without recourse to an extremely large-ion lithophile element-enriched parent liquid. It is difficult to generate parental liquids which are 2-4 x higher in the REE than average lunar KREEP, unless the liquids are the basic complement of a liquid-liquid pair, i.e., the so-called "REE-fraction," from the silicate liquid immiscibility of urKREEP. Scarce age information on lunar rocks suggests that magnesian-suite magmatism was initiated at progressively more recent time from the northeast to the southwest on the lunar nearside from 4.45 to 4.25 Ga. This magmatic "event" could be due to melting of the lunar mantle beneath these regions and could have been generated either by latent heat during crystallization of the final, KREEP-rich (and, thus, Th- and U-rich), residual, lunar magma ocean liquid or heating due to radioactive decay of K, Th, and U.

### Introduction

Breccias from the Apollo 14 landing site have provided a wealth of information on the genesis of the lunar highlands. Several pristine (i.e., monomict and exhibiting no evidence of meteoritic contamination) rock types have been discovered in relative abundance, including rare ferroan anorthosites, and alkali-suite (highlands rocks with both evolved mineral-chemical, major and trace element character) and magnesian-suite rocks (highlands rocks with primitive mineral compositions yet evolved trace elements) [e.g., *Shervais et al.*, 1983, 1984; *Lindstrom et al.*, 1984]. The study of rocks from the Apollo 14 landing site has also led to important discoveries of evolved samples [e.g., *Warren et al.*,

1983b; *Shih et al.*, 1985] which have added to our understanding of such processes as liquid immiscibility and metasomatism in the lunar environment.

Several scenarios have been presented which attempt to explain evolved trace element compositions in lunar rocks. Alkali anorthosites exhibit both evolved trace element compositions and evolved mineral chemistry. *Warren and Wasson* [1979a] suggested that alkali anorthosites were either precipitates from magnesian-suite magmas or ferroan anorthosites which assimilated a late incompatible element-enriched magma ocean residuum (urKREEP). They favored a relationship between alkali anorthosites, KREEP basalts, and alkali gabbro-norites and noted that alkali anorthosites lie along the magnesian-suite trend in the typical An-in-plagioclase versus Mg#-in-mafic minerals plot. *Snyder et al.* [1995] have shown that alkali suite cumulates could have been precipitated from

pristine KREEP basalts after approximately 60% crystallization. However, magnesian-suite anorthosites are enigmatic in that they exhibit primitive mineral chemistry and evolved trace element compositions. Lindstrom *et al.* [1984] pointed out the problems with an urKREEP assimilation hypothesis for magnesian-suite anorthosites. These problems include their low abundances in other incompatible elements (besides the rare earth elements (REE)) such as Zr, Hf, Rb, and Cs, and the fact that interstitial mafic grains, which are demonstrably more enriched in Fe when compared to mafic grains found in the bulk sample, could have crystallized from either an evolved trapped liquid or a later fluid. However, their interpretation of the data, which are sparse, is considered by many to be suspect. These perceived problems led them to postulate a possible alternative scenario whereby REE-rich metasomatic fluids infiltrated primitive troctolitic to anorthositic plutons in the lunar crust. The presence of phosphates (especially whitlockite) in some magnesian anorthosites and troctolites was the only petrographic evidence of this metasomatic event [Lindstrom *et al.*, 1984].

During a survey of pristine western highlands samples, several clasts were chosen from Apollo 14 breccias for Nd and Sr isotopic analyses. Most of the clasts were alkalic in character; however, one "large" magnesian-suite clast was analyzed for major and trace elements by Neal *et al.* [1991] and enough sample remained in the breccia for allocation for radiogenic isotopic study. Mineral-chemical data for this clast also exhibited primitive olivine and plagioclase compositions consistent with its classification within the magnesian suite. This ~2-g clast, of which we were allocated 193 mg, constitutes one of the few magnesian troctolitic anorthosites analyzed for both its Nd and Sr isotopic composition. We combine radiogenic isotopic data for this troctolitic anorthosite with "new" data on 12 "possibly pristine" to "probably pristine" magnesian-suite rocks and compare these to previously published data on this important rock suite. The evaluation and modeling of this total data set leads to some intriguing conclusions: (1) most, if not all, magnesian-suite rocks are the products of fractional crystallization of a primitive KREEP basalt; (2) some magnesian-suite rocks are so elevated in the REE and other incompatible elements as to require a secondary enrichment process, possibly metasomatism; (3) magnesian-suite rocks are generally younger than ferroan anorthosites and older than alkali suite rocks; and (4) magnesian-suite magmatism appears to have been initiated first in the northeastern part of the lunar near-side and then swept southwestward over a period of 300-400 m.y.

## Analytical Methods and Data Presentation

Major element mineral chemistry of these samples was determined on a CAMECA SX-50 electron microprobe at the University of Tennessee (Table 1). Polished thin sections were analyzed with an accelerating voltage of 15 kV and a specimen current of 20 nA. Standard ZAF corrections were applied. Whole rock chemistry for trace elements and the major elements Ca, Fe, Na, and K in the samples reported within were determined by M. M. Lindstrom at Washington University and the Johnson Space Center (JSC), Houston, Texas, using instrumental neutron activation (INA) techniques [Lindstrom and Korotev, 1982; Lindstrom, 1984; Neal *et al.*, 1991]. Major element analyses were also performed by a broad-beam electron microprobe method similar to that outlined by Jezek *et al.* [1978]. This method includes the crushing of INAA samples and fusing on a Mo strip in an Ar atmosphere followed by quenching. The elements Ca, Fe, Na, and K, which were also measured by INA, served as a check for the fused-glass method.

Generally the agreement is excellent, although Fe is sometimes low in the fused-bead analyses, possibly due to FeNi grains present in the quenched glass.

A 193-mg sample was separated from the breccia matrix of 14303,301 at NASA-JSC. No adhering matrix was indicated by NASA-JSC or found by the authors. This 193-mg sample of 14303,347 was crushed in acetone in a boron carbide mortar under a flow of better than class 100 air. Clean-room-quality air in the labs was always better than class 10. The sample was then carefully mixed to obtain a representative, approximately 23 mg, split which was used for whole rock Rb-Sr and Sm-Nd isotopic analyses. The sample was dissolved in HF, HNO<sub>3</sub>, and HCl, and isotope dilution measurements made on a 10-15% split of this solution with <sup>87</sup>Rb-<sup>86</sup>Sr and <sup>149</sup>Sm-<sup>150</sup>Nd mixed spikes. Total-process blanks for chemical procedures were less than 10 pg Rb, 120 pg Sr, 10 pg Sm, and 50 pg Nd. Sr and Nd isotopic data were obtained by multidynamic analysis on a VG Sector multicollector mass spectrometer. Sr and Nd isotopic analyses are normalized to <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194 and <sup>146</sup>Nd/<sup>144</sup>Nd = 0.7219, respectively. Analyses of SRM987 Sr and La Jolla Nd standards were performed throughout this study and gave weighted averages (at the 95% confidence limit) of <sup>87</sup>Sr/<sup>86</sup>Sr = 0.710250 ± 0.000011, and <sup>143</sup>Nd/<sup>144</sup>Nd = 0.511854 ± 0.000011, respectively. All isotope dilution measurements utilized static mode analyses.

By convention, the Nd isotopic data are also presented in Table 2 in epsilon units, or ratios relative to a chondritic uniform reservoir (CHUR) [DePaolo and Wasserburg, 1976]:

$$\epsilon_{Nd} = \left[ \frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}} - (^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}}{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}} \right] \times 10^4$$

Decay constants ( $\lambda$ ) for the parent isotopes of <sup>87</sup>Rb and <sup>147</sup>Sm are 1.402 × 10<sup>-11</sup> [as per Minster *et al.*, 1982] and 6.54 × 10<sup>-12</sup>, respectively.

## Petrography and Mineral Chemistry

We report here the mineralogy and petrography of 18 Apollo 14 highland clasts. Original igneous textures have been obliterated by either meteorite bombardment or disaggregation upon "plucking" of the individual clast from the breccia matrix. These new clasts have been subdivided into dunites, norites (including gabbronorites), anorthosites (included troctolitic and noritic varieties), and troctolites. Throughout this paper, the thin-section number is given first followed by the INA number in parentheses.

The cataclasis, disaggregation, and small size of these samples precluded the determination of accurate and meaningful modes from the thin section. Therefore, in order to obtain the modal mineralogy of the samples which were used for whole rock chemical determinations, we used mineral and major element whole rock analyses to calculate a mode (Table 1). This approach could not be applied to two anorthosites, 14305 (.539) and 14305 (.451), where thin sections were not available.

**Dunites: 14303,261(.260); 14303,305(.306); 14303,307(.308); 14304,161(.160)**

These four samples have been classified as dunites because they contain no felsic minerals and are composed of mostly olivine. Due to the small clast sizes (0.5 to 1.0 cm in diameter), the possibility remains that these are unrepresentative samples of another rock type, say, troctolite. However, for the purposes of our discussion, we have grouped these samples as dunites. Sample

Table 1. Mineral Compositions and Modal Abundances in Magnesian-Suite Clasts from Breccias 14303, 14304, 14305, and 14321

Sample	Plagioclase			Low-Ca Pyroxene			High-Ca Pyroxene			Olivine		Modal %					
	An	Wo	En	Wo	En	Mg#	Wo	En	Mg#	Fo	Plag	Oliv	Lo-Ca Px	Hi-Ca Px	Cr-Usp	Orth	Silica
<i>Dunites</i>																	
14303,261	-----	1-3	73-89	-----	75-90	-----	-----	-----	-----	81-87	-----	81	18.7	-----	0.3	-----	-----
14303,305	-----	-----	-----	-----	-----	-----	-----	-----	-----	76-88	-----	100	-----	-----	-----	-----	-----
14303,307	-----	-----	-----	-----	-----	-----	-----	-----	-----	86-89	-----	100	-----	-----	-----	-----	-----
14304,161	-----	-----	-----	-----	-----	-----	-----	-----	-----	83-89	-----	100	-----	-----	-----	-----	-----
<i>Norites</i>																	
14304,153	90-92	2-3	77-78	-----	77-80	-----	-----	-----	-----	68-70	71	3	25.9	-----	0.1	-----	-----
14304,175	94-95	1-3	77-87	-----	78-88	-----	-----	-----	-----	67-70	68	1	31	-----	-----	-----	-----
14304,450	96-98	1-3	86-89	-----	86-89	-----	-----	-----	-----	88-90	28	1	71	-----	-----	-----	-----
14304,489 <sup>a</sup>	85-95	2-10	48-76	38-39	53-71	33-34	54-55	-----	-----	66-67	56	1.5	30.5	-----	-----	3	-----
14304,9023	88-92	3-4	64-66	37-43	66-68	43-45	72-76	-----	-----	66-67	56	1.5	30.5	-----	-----	3	6
<i>Troctolites</i>																	
14303,323	93-96	-----	-----	-----	-----	-----	-----	-----	-----	81-88	83	17	-----	-----	-----	-----	-----
14304,171	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	68	1	31	-----	-----	-----	-----
14305,453	94-97	1-2	89-90	47-48	90-91	48-49	93-94	-----	-----	-----	81	-----	16	-----	-----	3	-----
14305,538 <sup>a</sup>	84-89	10-18	56-60	-----	66-69	-----	-----	-----	-----	57-58	84	16	-----	-----	-----	-----	-----
14305,9025	92-94	-----	-----	-----	-----	-----	-----	-----	-----	83-85	84	16	-----	-----	-----	-----	-----
14321,1379	94-96	-----	-----	-----	-----	-----	-----	-----	-----	86-88	72.5	27	-----	-----	-----	0.5	-----
<i>Anorthosites</i>																	
14304,165	95-96	-----	-----	-----	-----	-----	-----	-----	-----	87-89	92	8	-----	-----	-----	-----	-----
14304,173	93-96	-----	-----	-----	-----	-----	-----	-----	-----	77-84	97	3	-----	-----	-----	-----	-----

Mineral and major-element whole-rock analyses were used to calculate wt.% modes, and then converted to vol.% (see text).

<sup>a</sup> Modes could not be calculated for norite 14304,489 and troctolite 14305,538 (see text).

**Table 2.** Rb-Sr and Sm-Nd Isotopes of Mg-Suite Anorthosite 14303,347

Element/Isotope	Value
Rubidium, ppm	0.494
Strontium, ppm	257
$^{87}\text{Rb}/^{86}\text{Sr}$	0.00556
$^{87}\text{Sr}/^{86}\text{Sr}$	0.699492±13
$^{87}\text{Sr}/^{86}\text{Sr}(t)$	0.69915
Samarium, ppm	12.3
Neodymium, ppm	44.0
$^{147}\text{Sm}/^{144}\text{Nd}$	0.1683
$^{143}\text{Nd}/^{144}\text{Nd}$	0.511792±6
$\epsilon_{\text{Nd}(t)}$	-1.0

14303,261,(260) shows extreme cataclasis and exhibits grain sizes from <0.1 to 0.5 mm. Interstitial, opaque glass is also present in this sample and was presumably formed by impact melting (no evidence of glass injection from without is present; the glass was likely formed *in situ* at grain boundaries). The remaining three dunites are composed of disaggregated grains (up to 1.5 mm). The grain sizes are generally small (~0.5 mm) due to cataclasis and annealing. Olivine compositions range from  $\text{Fo}_{76}$  to  $\text{Fo}_{89}$ , with only negligible intergrain variation; the range of Fo contents for 14303,305,(306) is 76 to 88 and is due to core-to-rim zonation (Figure 1). All but 14303,261,(260) are composed of 100% olivine, although 14303,305,(306) contains a single grain of chromite which

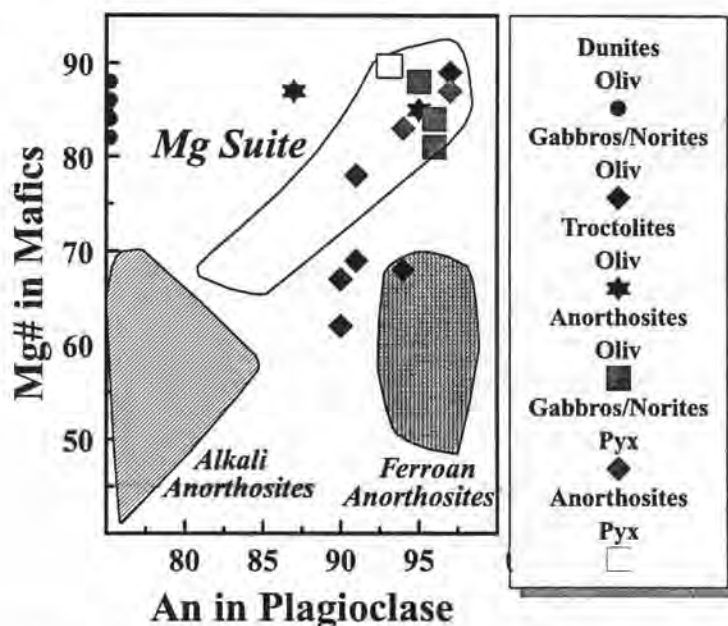
could not be analyzed, as it was below the surface of the olivine. 14303,261,(260) contains 18.7% low-Ca pyroxene ( $\text{Wo}_{1-3}\text{En}_{77-85}$ ) as well as 0.3% chromite (Table 1). The Mg# of the pyroxene in this sample is somewhat more varied (75-90) than the Fo content of the accompanying olivine (81-87). Much of this can be accounted for by core-to-rim zonation toward lower Fo content. These core-to-rim zonations in olivine suggest a relatively shallow emplacement depth [i.e., *Ryder, 1992*].

**Norites:** 14304,153,(154); 14304,175,(174); 14305,450,(450); 14305,489,(489); 14305,9023,(443)

Two samples, 14305,489,(489) and 14305,9023,(443) contain both Ca-rich and Ca-poor pyroxenes (Table 1), but are not classified as gabbronorites, as they contain <10 modal % high-Ca pyroxene.

Sample 14305,489,(489) is a grain mount of two pieces ~1.5 x 1.5 mm. The texture is cataclastic with a variable grain size (<0.1 to 0.5 mm). Armalcolite (~0.2 mm) is present as a relatively early crystallizing phase in 489, corresponding to the Ca-rich type 2 composition of *Haggerty* [1973]. It is rimmed with ilmenite, indicating a back reaction with the interstitial melt (Figure 2a). The constituent minerals exhibit a large range of intergrain variability (Table 1) and the whole rock major element abundances could not be recreated using these observed mineral compositions.

Norites 14305,9023,(443) and 14305,450,(450) are collections of disaggregated mineral grains (0.2-0.3 mm and 0.3-0.4 mm, respectively). Monomineralic grains predominate, but occasionally biminerallitic grains of plagioclase and pyroxene are seen. 14305,9023,(443) also contains evolved minerals, such as orthoclase, silica, and high-Ca pyroxene. We consider this mineral assemblage to represent the crystallized intercumulus liquid and that such compositions are required in order to generate the major



**Figure 1.** Plot of An content of plagioclase versus Mg# in mafic minerals for "new" magnesian suite clasts. Also shown are fields for lunar alkali anorthosites, ferroan anorthosites, and the magnesian-suite.

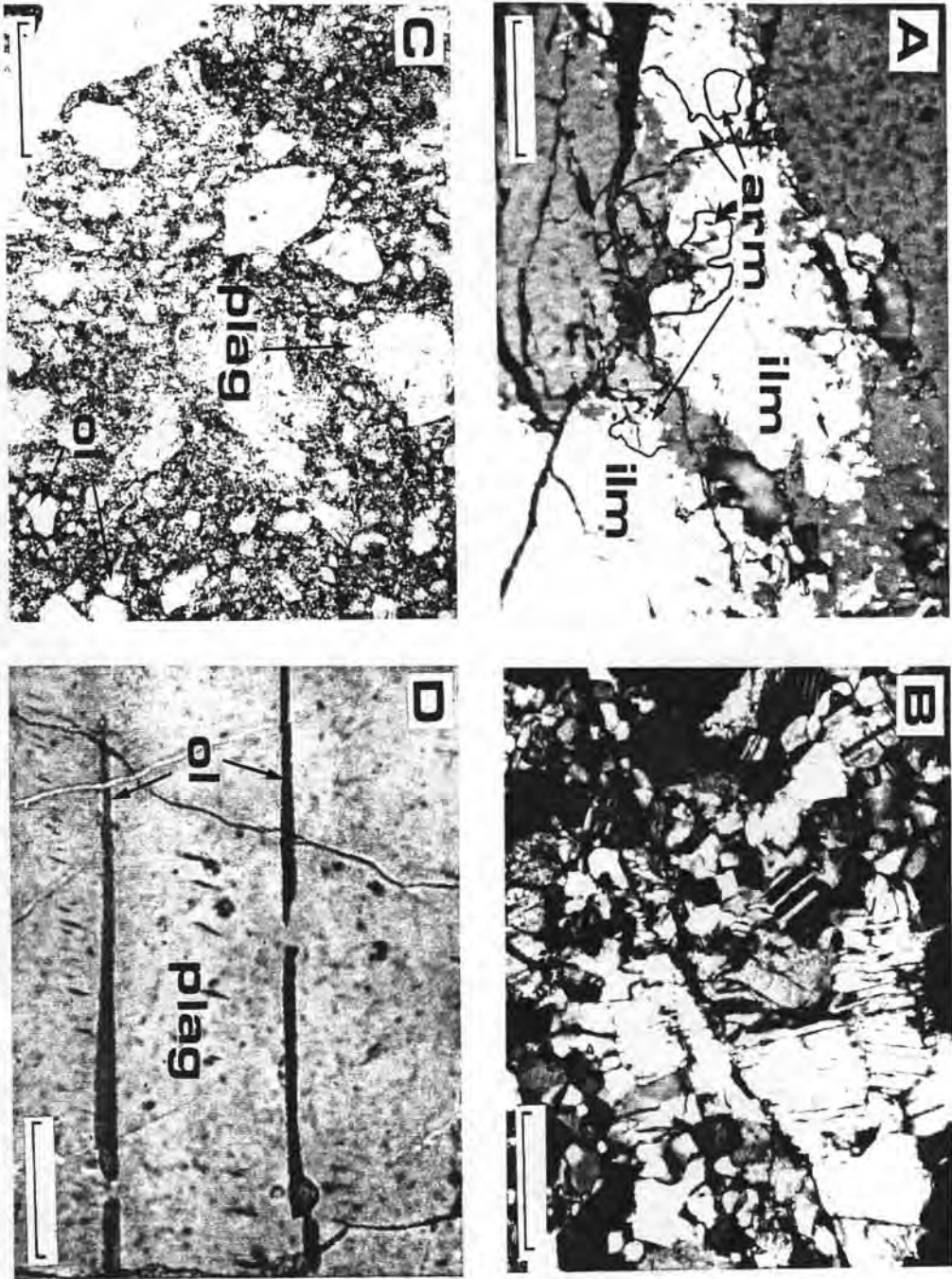


Figure 2. Representative photomicrographs of the Apollo 14 magnesian suite rocks. (a) Armalcolite rimmed with ilmenite in norite 14305,489(.489). Scale bar represents 60  $\mu$ m. (b) Plagioclase exhibiting deformed twin lamellae from norite 14304,153(.154). Scale bar represents 0.25 mm. (c) Cataclasis in troctolite 14304,171(.170). Scale bar represents 0.25 mm. (d) Olivine lamellae in plagioclase from anorthosite 14303,323(.302). Scale bar represents 0.25 mm.

Table 3. Whole Rock Compositions of Magnesian-Suite Clasts From Breccias 14303, 14304, and 14305

	14303	14303	14303	14304	14304*	14305	14305	14305*	14305*	14305	14305	14305	14305	14305	14305	14305	14305	14305	
INAA <sup>b</sup>	261	306	308	160	154	443	450	489	537	1331	302	166	172	451	453	460	539		
PM <sup>c</sup>	260	305	307	161	153	9023	450	489	538	1379	323	165	173	453	453	9025			
Rock <sup>d</sup>	D	D	D	D	N	N	N	N	T	T	T	A	A	A	T	T	A	A	
Mass, mg	13.4	27.3	27.5	46.3	52.2	57.7	13.9	11.9	12.0	14.8	105.9	94.3	56.9	51.8	11.5	24.1	22.3		
SiO <sub>2</sub>	41	40	40	40	48.1	47.4	52	55	44.2	43	43.9	44.2	45	44	46	44	44.4		
TiO <sub>2</sub>	0.09	0.15	0.09	0.06	0.26	0.16	0.28	0.16	0.55	0.26	0.26	0.29	0.14				0.06		
Al <sub>2</sub> O <sub>3</sub>	0.47	0.63	1.97	0.72	23.9	24.8	20.2	10.7	22.3	26.0	29.0	31.9	34.4	35.0	29.7	29.7	34.9		
Cr <sub>2</sub> O <sub>3</sub>	0.24	0.17	0.09	0.09	0.20	0.17	0.07	0.18	0.03	0.08	0.04	0.04	0.03	0.14	0.03	0.01	0.02		
FeO	16.5	14.6	11.3	12.1	4.45	3.97	7.17	6.01	9.79	5.17	3.33	2.44	1.67	0.75	2.21	2.58	0.24		
MnO	0.16	0.16	0.14	0.12	0.10	0.06	0.06	0.06	0.06	0.06	0.02	0.03	0.01	0.002	0.01	0.02	0.01		
MgO	40.8	43.4	45.2	46.4	9.21	9.38	8.00	22.7	13.8	13.3	7.76	3.39	1.02	0.68	5.60	7.30	0.37		
CaO	0.45	0.50	1.10	0.52	13.4	13.6	11.6	5.70	3.60	13.1	13.3	16.0	17.9	18.9	20.0	15.6	19.3		
Na <sub>2</sub> O	0.01	0.02	0.02	0.01	0.32	0.32	0.70	0.13	0.14	0.41	0.33	0.45	0.41	0.22	0.31	0.28	0.52		
K <sub>2</sub> O	0.01	0.01	0.01	0.01	0.08	0.07	0.60	0.03	0.12	0.11	0.10	0.12	0.04	<0.15	<0.15	<0.15	0.12		
P <sub>2</sub> O <sub>5</sub>	0.14	0.07	0.05	0.07	0.03	0.04	0.04	0.03	0.19	0.06	0.06	0.05	0.02	0.05	0.02	0.05	0.03		
Mg#	81.5	84.2	87.7	87.3	78.7	80.8	66.5	87.1	82.6	87.7	85.0	70.3	70.8	85.2	88.8	83.5	73.3		
Sc, ppm	2.8	3.4	6.8	5.4	9.8	7.6	13.1	16.2	5.4	7.8	2.0	2.3	3.3	1.2	0.52	1.8	1.7	0.72	
Cr	1680	1110	563	553	1549	1328	1000	2630	511	570	268	172	276	237	210	377	114	91	
Co	37.5	54.2	58.7	60.0	18.9	18.5	16.1	25.7	46.6	22.3	11.8	8.2	7.7	2.3	1.8	8.6	8.8	1.08	
Ni	170	142	221	143	<29	116	50	30	90	109	21	56	66	21	12	37	23	18	
Ba	28	28	34	11	233	240	740	72	100	429	248	451	301	272	70	54	365	374	
La	3.18	6.79	5.47	1.26	10.5	7.31	22.5	16.6	9.75	33.6	8.04	30.0	16.5	10.5	5.01	3.96	8.04	11.5	
Ce	10.5	17.3	12.9	3.43	24.5	18.8	56.5	39.0	25.3	85	17.0	77.2	40.5	24.8	12.0	10.0	20.0	22.5	
Sm	1.40	3.07	1.66	0.52	3.64	2.31	8.33	3.78	4.42	14.2	1.66	12.2	5.93	3.49	1.52	1.28	2.34	1.88	
Eu	0.091	0.138	0.217	0.06	2.00	1.71	1.72	0.49	0.69	1.83	1.94	3.09	2.21	2.49	1.25	1.06	3.18	3.03	
Tb	0.268	0.57	0.275	0.118	0.80	0.423	1.76	0.86	0.81	2.68	0.213	2.25	1.02	0.573	0.224	0.18	0.30	0.269	
Yb	1.45	2.04	2.31	2.44	4.04	2.93	10.5	5.79	2.83	9.69	1.13	6.68	3.65	1.69	0.71	0.887	0.96	0.973	
Lu	0.22	0.29	0.463	0.512	0.63	0.431	1.67	0.945	0.50	1.41	0.16	0.88	0.519	0.215	0.073	0.114	0.142	0.125	
Zr	53	72	<40	660	100	109	410	150	215	340	<20	280	104	32	<20	<30	34		
Hf	1.00	1.82	1.12	15.9	2.74	3.17	11.3	3.99	4.54	9.98	0.197	6.57	3.18	1.18	0.285	0.12	0.38	0.979	
Ta	0.148	0.207	0.141	0.049	0.318	0.226	0.83	0.42	0.43	1.11	0.062	0.916	0.456	0.156	0.026	<0.03	0.045	0.073	
Th	0.70	1.05	0.57	0.36	1.19	0.67	5.13	1.35	1.50	5.38	0.064	5.49	1.80	1.06	0.123	0.08	0.168	0.319	
U	0.28	0.25	0.11	0.24	0.22	0.30	1.65	0.70	0.52	1.66	0.03	2.10	0.57	0.20	0.049	0.069	0.18		
Ir, ppb	<2.1	<1.1	<3.2	<1.2	nd <sup>4</sup>	nd	nd	nd	1.6	nd	nd	0.5	0.8	<1	nd	nd	nd	nd	
Au	<1	nd	7.1	0.9	<1.2	1.8	<3	<4	1.6	<2.5	<3	<1	1.1	<1	<1	<3	<1.2		

All trace element data were performed by M.M. Lindstrom and L.A. Haskin using INAA at Washington University, St. Louis, and NASA/JSC, Houston, Tex., 1985-1987 [see Lindstrom and Korotev, 1982; Neal *et al.*, 1991]. Major elements were analyzed using broad-beam electron microprobe methods outlined in the text.

nd, not detected.

Footnotes: <sup>a</sup> Probably not pristine; <sup>b</sup> Sample number used for INAA; <sup>c</sup> Sample used for electron microprobe mineral-chemical analyses;

<sup>d</sup> Rock type: D, dunite; N, norite; T, troctolite; A, anorthosite.

elements analyzed by INA (Table 3). The calculated mode for this sample is 56% plagioclase ( $An_{88-92}$ ), 30.5% low-Ca pyroxene ( $Wo_{3-4}, En_{64-66}, Mg\# 66-68$ ); 3% high-Ca pyroxene ( $Wo_{37-43}, En_{43-45}, Mg\# 72-76$ ); 1.5% olivine ( $Fo_{66-67}$ ); 6% silica; and 3% orthoclase ( $Or_{90-91}$ ; Figure 1). The more evolved phases account for the high silica and  $K_2O$  contents of the whole rock analysis (Table 3). Sample 14305,450(,450) is composed of 71% low-Ca pyroxene ( $Wo_{1-3}, En_{86-89}, Mg\# 86-89$ ), 28% plagioclase ( $An_{96-98}$ ), and 1% olivine ( $Fo_{88-90}$ ).

Sample 14304,153(,154) is composed of annealed plagioclase (71%;  $An_{90-92}$ ), olivine (3%;  $Fo_{68-70}$ ), low-Ca pyroxene (25.9%;  $Wo_{2-3}, En_{77-78}$ ;  $Mg\# 77-80$ ), and chromite (0.1%;  $Mg\# 28-29$ ;  $Cr/(Cr+Al) \sim 79-80$ ) (Tables 1 and 3). The recrystallized plagioclase contains relict twin lamellae which are kinked (Figure 2b). The overall texture is granulitic exhibiting some recrystallization to granoblastic.

Thin section 14304,175 is made up of three  $1.5 \times 1.4$  mm chips, two of which are maskelynitized plagioclase containing  $\sim 20$   $\mu m$  wide "stringers" of olivine and low-Ca pyroxene. Generally, olivine is found as small ( $\sim 10$   $\mu m$ ) inclusions in pyroxene, although discrete olivines are also present. These "stringers" vary in length from 0.6 mm to  $<0.1$  mm. The third chip contains cataclastized and granulated plagioclase again with these olivine stringers, but also with a larger ( $\sim 0.1$  mm) low-Ca pyroxene grain. The larger grain contains a pyroxene of slightly different composition ( $Wo_{1-2}, En_{85-86}$ ;  $Mg\# 87-88$ ) than the pyroxene in the stringers ( $Wo_{2-3}, En_{77-80}$ ;  $Mg\# 78-81$ ). Olivine exhibits little intragrain or intergrain variation ( $Fo_{67-70}$ ) and plagioclase is similarly restricted in composition ( $An_{94-95}$ ; Figure 1). The calculated modes demonstrate that 14304,175(,174) is composed of 68% maskelynitized plagioclase, 1% olivine, and 31% low-Ca pyroxene.

**Troctolites: 14303,323(,302); 14304,171(,170); 14305,453(,453); 14305,538(,537); 14305,9025(,460); and 14321,1379(,1331)**

All plagioclase has been converted to maskelynite. Sample 14304,171(,170) has experienced extensive cataclasis, which has produced a highly variable grain size ( $<0.1-0.3$  mm; Figure 2c). The olivine forms the largest grains and contains small ( $<0.1$  mm) chromite-ulvospinel inclusions. Mineral compositions are restricted (plagioclase  $An_{95-97}$ ; olivine  $Fo_{62-66}$ ; Cr-ulvospinel:  $Mg\# 22-28$ ,  $Cr/(Cr+Al) = 51-54$ ; Table 1 and Figure 1). Calculated modes demonstrate that this sample is composed of 72% maskelynitized plagioclase, 27.9% olivine, and 0.1% Cr-ulvospinel (Table 1). Sample 14303,323(,302), contains "lamellar" inclusions or "stringers" of olivine, and one sample [14304,453(,453)] contains lamellar inclusions of low-Ca pyroxene. There was no section available for 14305,538(,537) and only a few grains were retained after the INA analysis.

Two thin sections have been made for one "large" troctolite anorthosite from 14303 [323 and 324] which was  $2 \times 2$  cm (prior to plucking) and weighed over 2 g. A split of the remaining sample was also allocated (14303,347) and used for Nd and Sr isotopic analyses. Sample 14303,323(,302) is composed of disaggregated maskelynite containing lamellae of olivine (up to 20  $\mu m$  wide). Several olivine grains (up to 2 mm) are present in 323. The lamellar olivines exhibit a restricted composition of  $Fo_{87}$ , whereas the discrete grains exhibit core-to-rim zoning from  $Fo_{87}$  to  $Fo_{81}$  (Figure 1). Plagioclase exhibits a restricted composition of  $An_{93-96}$  (Table 1 and Figure 1). The calculated mode of 14303,323(,302) is 83% maskelynitized plagioclase and 17% olivine (Table 1).

Sample 14305,453(,453) consists of discrete grains (up to 1 mm) of maskelynite and pyroxene. One small ( $<0.1$  mm) ilmenite was

the only inclusion observed and this was found in the plagioclase. No probe analyses could be carried out on this inclusion, and no other ilmenites were observed. Mineral compositions demonstrate the presence of low-Ca ( $Wo_{1-2}, En_{89-90}$ ;  $Mg\# 90-91$ ) and high-Ca ( $Wo_{47-48}, En_{48-49}$ ;  $Mg\# 93-94$ ) pyroxene (Table 1). Calculated modes demonstrate that this rock is composed of 81% maskelynitized plagioclase ( $An_{94-97}$ ), 16% low-Ca pyroxene, and 3% high-Ca pyroxene.

Sample 14305,9025(,460) is made up of bi-mineralic chips (up to 0.7 mm) of maskelynite containing euhedral inclusions of olivine (up to 0.15 mm). Mineral compositions are restricted ( $An_{92-94}$  and  $Fo_{83-85}$ ; Table 1). The calculated mode for this sample is 84% maskelynitized plagioclase and 16% olivine.

Troctolite 14321,1379(,1331) has a granulitic texture with recrystallized plagioclase (containing unstrained twin lamellae) up to 0.15 mm in longest dimension and partially recrystallized olivine. Where recrystallized, the olivine is of a similar size to the plagioclase, but where it has escaped recrystallization, it reaches up to 0.4 mm. The high-Ca pyroxene is usually associated with olivine. The constituent minerals are of restricted composition: olivine  $Fo_{86-88}$ ; plagioclase  $An_{94-96}$ ; and high-Ca pyroxene  $Wo_{45-47}, En_{49-51}$  and  $Mg\# 92-93$  (Table 3). The calculated mode of 14321,1379(,1331) is 72.5% plagioclase, 27% olivine, and 0.5% high-Ca pyroxene (Table 1).

**Anorthosites: 14304,165(,166); 14304,173(,171)**

The new anorthosites all contain maskelynite and have been extensively cataclastized. Anorthosite 14304,165(,166) is made up of mono- and bi-mineralic chips of maskelynitized plagioclase and olivine  $\sim 1 \times 1$  mm. As in 14303,323(,302), plagioclase contains stringers of olivine  $\sim 20$   $\mu m$  wide; the largest stringer observed was  $\sim 200$   $\mu m$  in length (Figure 2d). The discrete olivine grains also contain evidence of cataclasis from meteorite impact. Mineral compositions are restricted ( $An_{95-96}$  and  $Fo_{87-89}$ ) and the calculated modes yield abundances of 92% plagioclase and 8% olivine.

Sample 14304,173(,171) is made up of chips  $\sim 1-2$  mm in diameter. Commonly, maskelynitized plagioclase contains stringers of olivine up to 50  $\mu m$  wide and 300  $\mu m$  in length. Adhering breccia matrix was noted on several chips. Clast 14304,173(,172) contains "lamellar" inclusions or "stringers" of olivine. The individual olivine stringers are of uniform composition, but intragrain variation causes the range in olivine compositions ( $Fo_{77-84}$ ; Table 1). Plagioclase is more restricted in composition ( $An_{93-96}$ ; Table 1 and Figure 1). The calculated mode demonstrates that 14304,173(,172) is composed of 97% maskelynitized plagioclase and 3% olivine (Table 1).

## Whole Rock Chemistry

Whole rock chemistry has been determined on 18 magnesian-suite samples, and the data are reported in Table 3. However, before all of these data are presented and modeled, we must evaluate the chemical pristinity of these clasts through a critical analysis of siderophile element distributions.

### Pristinity of Highlands Rocks and Sample Size Conventions

"Many pristine rocks are brecciated and few unequivocal vestiges of igneous texture remain." *Ryder et al.* [1980].

"Certainly, siderophile elements should not be considered proof for, or against, pristinity, in isolation from all other evidence." *Warren* [1993].

The pristinity of lunar rocks is a concept which is critical in establishing lunar chronology and any petrogenetic interpretation of returned Moon rocks. The effects of meteorite bombardment will have a tendency to obscure primary lunar igneous textures, contaminate samples with "extra-lunar" components, as well as mix unrelated lithologies. Meteorite bombardment may also simply metamorphose an existing lunar lithology, neither mixing nor adding other components. In order to recognize these different effects, *Warren and Wasson* [1977] presented criteria for establishing the pristine nature of nonmare (i.e., highland) rocks: "(1) low siderophile-element contents ( $<3 \times 10^{-4}$  times CI chondrites); (2) low incompatible-element contents ( $<5 \times 10^{-3}$  times KREEP levels); (3) coarse-grained texture (grains  $> 3$  mm); (4) antiquity ( $> 4.2$  Ga); (5) homogeneity of composition between similar grains; (6) low initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios ( $< 0.6992$ ); (7) samples which exhibit a general cumulate texture." *Warren and Wasson* [1977, 1978, 1979a, 1980] and *Warren* [1993] have emphasized the importance of siderophile-element concentrations in determining the pristinity of lunar highlands lithologies. However, these authors have cautioned against using this criterion alone, as endogenous lunar igneous processes may have the potential to concentrate the siderophile elements [e.g., *Warren et al.*, 1987], not unlike those processes which lead to the formation of terrestrial ore deposits. In fact, *Warren et al.* [1987] classified several crustal rocks with Ir contents of 1.27, 1.30, 3.40, 4.30, and 13.0 ppb as pristine and stated that "these results underscore the huge uncertainty associated with any attempt to estimate the overall siderophile element contents of the moon's crust." Finally, the term "pristine" has been defined variously as lunar rocks which are "endogenously igneous, created by igneous events within the Moon and subsequently have retained their bulk chemistry without contamination by other lunar or non-lunar materials" and "compositionally endogenous (igneous or thermally metamorphosed igneous) or monomict brecciated, unadulterated by other materials" [*Warren and Wasson*, 1977, 1978, 1979a, b, 1980; *Warren et al.*, 1981, 1983a, b, c; *Norman and Ryder*, 1979; *Ryder et al.*, 1980].

If, indeed, we wish to utilize these "new" rocks to determine the petrogenesis of the magnesian suite, we must demonstrate that they are monomict. Most of the samples are brecciated, so the possibility exists that two different lithologies could have been mechanically mixed together by the meteoritic impact which likely caused brecciation. If this is the case, then a broad range of mineral compositions should be observed, unless the sample has undergone total reequilibration akin to granulite facies metamorphism [*Lindstrom and Lindstrom*, 1986]. Such variation in mineral compositions is observed in two samples: dunite 14305,305,(306) and norite 14305,489,(489) (Table 1). The variation witnessed in olivine ( $\text{Fo}_{76-88}$ ) in the dunite is core-to-rim, but that in the norite (plagioclase  $\text{An}_{85-95}$ ; olivine  $\text{Fo}_{64-88}$ ) is between grains. Norite 14305,489,(489) also contains anomalously low Sc and Cr and anomalously high Co when compared to other norites (Table 3). Furthermore, the major-element whole rock composition of this sample could not be reconstructed from measured mineral analyses, suggesting that some other "nonnoritic" component was included in the sample analyzed by INA. Five other clasts exhibit bulk-rock Mg# which are inconsistent with observed mafic mineral chemistry (i.e., the Mg# of the whole rock analyses are vastly different than the Mg# of either low-Ca pyroxene or olivine): norites 14304,153,(154) and 14304,175,(174); troctolite 14305,538,(537); and anorthosites 14304,165,(166) and 14304,173,(172). All other clasts exhibit little variation in mineral chemistry, and all whole rock major-element compositions can be reconstructed satisfactorily

using observed mineral compositions [*Neal et al.*, 1991]. On the basis of this evidence, we conclude that norites 14304,153,(154), 14304,175,(174), 14305,489,(489); troctolite 14305,538,(537); and anorthosites 14304,165,(166) and 14304,173,(172) are polymict, and thus, not pristine. Mineral chemistry could not be determined on anorthosites 14305,453 and 14305,539; however, the relatively low Ni, Co, Ir (below detection limit) and Au (below detection limit) suggest that these samples are probably pristine (Table 3). Out of the original 21 magnesian-suite clasts analyzed by *Neal et al.* [1991], 3 were determined to be nonpristine. We have further outlined criteria to show that six more of those samples are polymict, whereas the remaining 12 are probably to possibly pristine (a la *Warren*), monomict/brecciated samples.

None of the samples described within are texturally pristine and all indicate evidence of metamorphism due to meteorite impact. However, there is a significant lack of textural evidence (i.e., triple junctions) that would suggest thermal reequilibration within these samples. Therefore, the lack of variation in mineral chemistry likely is inherited from a monomict rock which has been brecciated. Sr isotopic studies of one of the anorthosites, 14303,302,(323; 347) which is extensively maskelynitized, suggest that this sample exhibits a primitive initial ratio [ $^{87}\text{Sr}/^{86}\text{Sr}$  (at 4.2 Ga) = 0.69915], consistent with a "pristine" rock. Furthermore, Sm, Nd, and Rb abundances in this rock are not consistent with the late addition of a KREEP component (see below). Thus, texture is not particularly helpful in determining the likelihood of chemical pristinity for these samples. Most of the samples described within have Au and Ir concentrations below detection limits; however, the limitations of the INA technique do not permit Au and Ir to be detected precisely at the sub-ppb level (Table 3). Therefore, the fact that many samples exhibit Au and Ir below the detection limit is a necessary prerequisite to the designation as pristine, but is not sufficient to positively determine their pristinity. We will classify such samples as "probably pristine". Other samples exhibit resolvable, albeit very low, Ir abundances of  $0.9 \pm 0.6$  ppb. Like Ir, Au is either not detected or values are reported as upper limits; where resolvable, Au is determined at  $1.4 \pm 0.6$  ppb levels. These samples are considered to be "possibly pristine", although this designation must be made with caution and awaits further evaluation (see discussion below). However, one sample, dunite 14303,307,(308) contains  $7.1 \pm 0.5$  ppb Au. This immediately suggests meteoritic contamination, but its Ir abundance is below detection ( $< 3.2$  ppb) and, whereas the CI chondrites exhibit Ir in greater abundance than Au (Ir/Au = 3-5), it is not likely that these abundances are due to meteoritic contamination. Furthermore, the presence of siderophiles in pristine highlands samples is not unprecedented. *Morgan and Wandless* [1988] repeatedly analyzed dunite 74217 for siderophile elements and demonstrated higher than "accepted" pristine values for this pristine rock (Ir =  $0.64 \pm 1.14$  ppb; Au =  $2.33 \pm 1.66$  ppb). Recall that *Warren et al.* [1990] reported 7.4 ppb Ir and 12.4 ppb Au for harzburgitic sample 12033,503b. These authors classified this sample as pristine despite elevated siderophile abundances because the lithology was "obviously monomict." With this in mind, we suggest that 14303,307,(308) is also "possibly pristine".

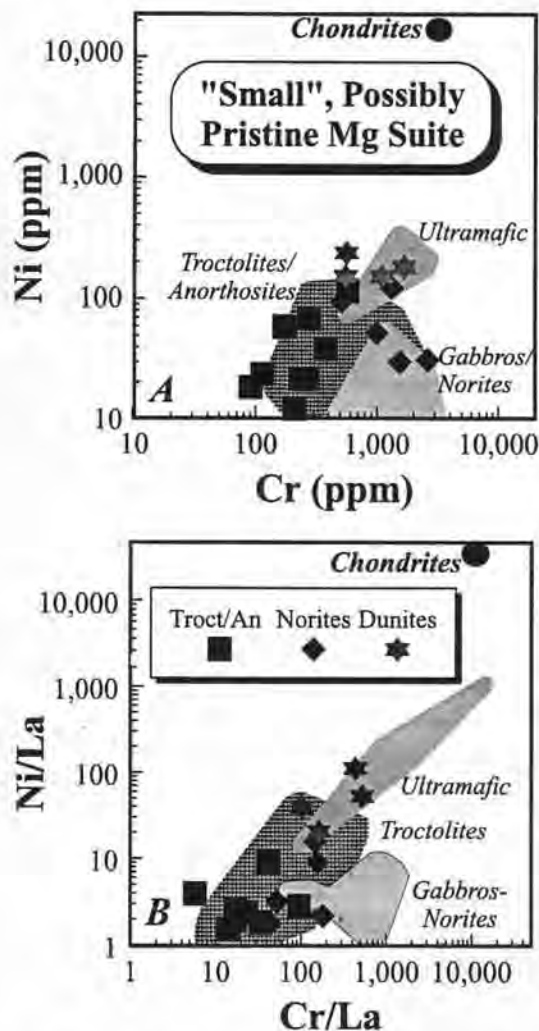
Prior to this study, a total of 88 samples had been studied that are considered members of the magnesian suite, and which could possibly be determined pristine [*Warren*, 1993]. However, many of these samples are extremely small (down to 0.3 mg in mass) and others are less likely pristine. In order to evaluate these 19 new samples in reference to previously analyzed magnesian suite rocks, it was necessary to determine which of the previous rocks could most plausibly be considered pristine. As per the compilation of



Warren [1993], we have carefully considered mass, pristinity (on a scale of 1 to 10 with 10 being undoubtedly pristine), and homogeneity of mineral-chemical compositions, and have concluded that only those original samples which were larger than 100 mg in mass and have a pristinity value  $> 6$  should be considered in our data compilations. This will allow us to compare our 18 "new" samples with those which are most likely both representative of actual rocks and pristine. This is not to say that very small samples cannot be and are not pristine (many of the samples within this present study are, indeed, very small). These size and pristinity boundaries are necessarily arbitrary, however, it is interesting to note that, of the original samples which were  $< 100$  mg in mass,

only six have been assigned by Warren [1993] a pristinity value  $> 6$ .

Using this screen, we then arrive at 55 samples which comprise the highlands magnesian-suite. Of these "more likely" pristine samples, 2 are from the Apollo 12 collection, 19 from the Apollo 14 collection, 11 from the Apollo 15 collection, 5 from the Apollo 16 collection, and 18 from the Apollo 17 collection. These include samples which were classified by Warren [1993] and which we have slightly modified into troctolites/anorthosites (28), gabbros/norites (23), ultramafic cumulates (2), and dunites (2). In the following discussions, only data from these 55 samples (minus seven which either do not have analyses or the analyses were considered "impure(?)" by Warren [1993]) will be included in data compilations for the magnesian suite. Note that the abundances of moderately siderophile elements, such as Ni, Co, Cr, for our "new" magnesian suite clasts do not exceed the values of these elements in the compiled "large, probably pristine" rocks (Figure 3a). Furthermore, siderophile/lithophile ratios of our samples also fall within the ranges of the "large, probably pristine" rocks (Figure 3b).



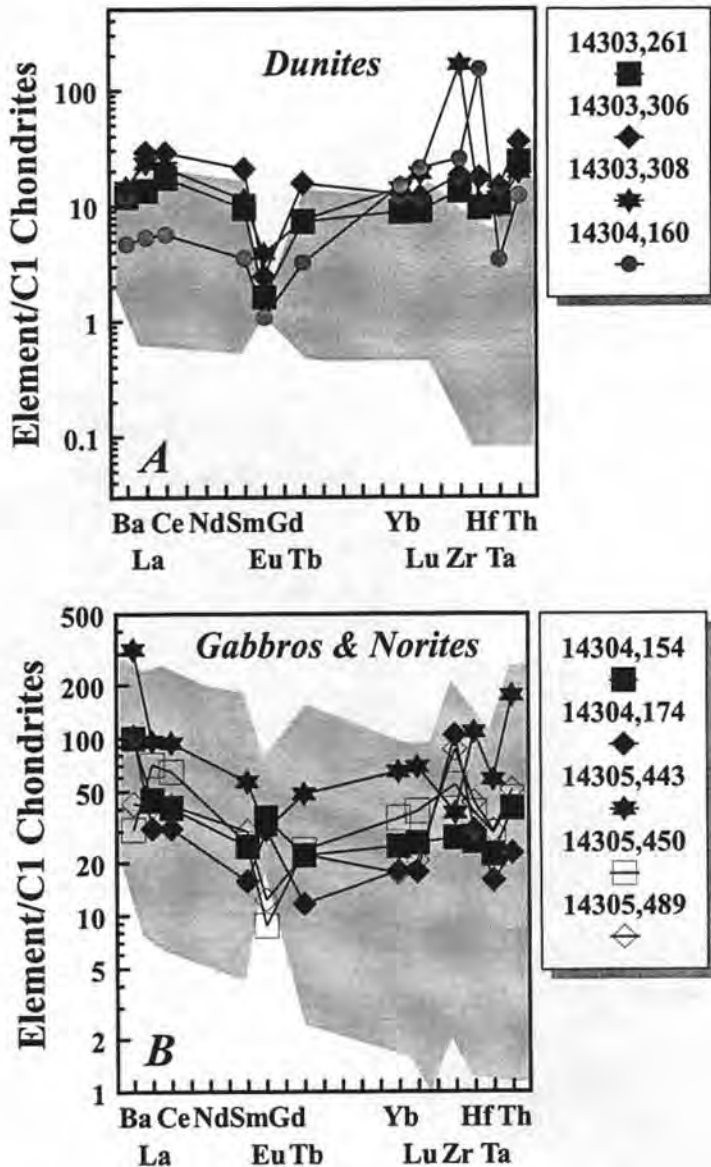
**Figure 3.** Log-log plots of (a) the moderately siderophile elements Cr (ppm) versus Ni (ppm) for our "new" magnesian-suite rocks, and (b) moderately siderophile/lithophile element ratios Cr/La versus Ni/La. Also shown are fields for all large ( $> 100$  mg), probably pristine gabbros/norites, troctolites/anorthosites, ultramafic rocks, and chondritic meteorites.

### Chemistry of Petrologic Groups

**Dunites.** The major element compositions of the four dunites are predictably similar to olivine with Mg# from 81.5 to 87.7. The four dunites all contain negative Eu anomalies ( $\text{Eu}/\text{Eu}^* = 0.16\text{--}0.41$ ) and three exhibit a positive slope in the heavy rare earth elements (HREE) (Figure 4a). One such sample is 14304,161(,160) and contains an overall light rare earth element (LREE) depletion relative to the HREE [ $(\text{La}/\text{Lu})_n = \text{La}/\text{Lu}$  normalized to C1 chondrites = 0.25]. Neal *et al.* (1991) have suggested that this LREE depletion, combined with enrichment in Hf (Table 3), may be due to the inclusion of zircon in the analyzed sample. The other three dunites are LREE enriched [ $(\text{La}/\text{Lu})_n = 1.23\text{--}2.34$ ]. All four dunites are grossly similar to the two previously analyzed dunites from the magnesian suite (Figure 4a). An important feature of this diagram is the relatively high REE abundances of the dunites; La ranges from 4-20x that in chondrites. This is anomalously high for a rock dominated by olivine. The compatible elements reflect the dominance of olivine in the assemblage (i.e., high Co abundances of 38-60 ppm and low Sc abundances of 2-7 ppm; Table 3).

**Gabbros and norites.** The whole rock Mg# of the norites range from 66.5 to 87.7 (Table 3). All of the norites from this study are LREE enriched [ $(\text{La}/\text{Lu})_n = 1.39\text{--}2.27$ ], although two samples, 14305,450(,450) and 14305,9023(,443), exhibit positive HREE slopes (Figure 4b). Three of the norites contain negative Eu anomalies ( $\text{Eu}/\text{Eu}^* = 0.32\text{--}0.90$ ), whereas non-pristine clasts 14304,153(,154) and 14304,175(,174) contain positive Eu anomalies ( $\text{Eu}/\text{Eu}^* = 1.53$  and 2.17, respectively). The REE abundances in the norites are generally higher than in the dunites, with La ranging from approximately 29 to 70 times chondrites. Elemental compositions are also within the ranges of previously analyzed magnesian gabbros and norites (Figure 4b). The compatible elements reflect the predominance of pyroxene as the mafic mineral with Sc  $> 7$  ppm and Co  $< 26$  ppm. The one exception to this is the non-pristine clast 14305,489(,489) which contains lower abundances of Sc (5.4 ppm) and higher abundances of Co (46.6 ppm). The significance of this will be discussed below. The high field strength elements (HFSE; large, multiply charged cations which tend to be incompatible in common igneous minerals, except ilmenite; i.e., Zr, Hf, Ta, Nb, Th, and U) are present in similar abundances as in the dunites.

**Troctolites and anorthosites.** Mg# for the troctolites range from 80.8 to 87.7 and those for anorthosites range from 70.8 to 88.8. Although only four troctolites were included in this study,



**Figure 4.** Chondrite-normalized plots of trace-elements for magnesian suite rocks: (a) dunites; (b) norites; (c) troctolites and anorthosites. Also shown are shaded fields for all other "large," pristine, magnesian-suite rocks of that lithology (Tables 4 and 6).

they exhibit a wide range of REE abundances (Figure 4c). They are all LREE enriched [ $(La/Lu)_n = 2.47$  and  $5.18$  for 14305,538,(537) and 14321,1379,(1331), respectively], but 14305,1379,(1331) exhibits a positive trend in the HREE. Sample 14321,1379,(1331) also contains a positive Eu anomaly ( $Eu/Eu^* = 3.91$ ). The compatible elements are generally depleted relative to the other rock types, whereas the HFSE have similar abundances.

The anorthosites also exhibit a broad range in REE abundances (Figure 4c). For example, the range in La abundances (relative to chondrites) is from approximately 20 to 90. All are LREE enriched [ $(La/Lu)_n = 3.31$ - $9.96$ ], although one sample, 14305,453,(453), exhibits a positive HREE slope (Figure 4c). Positive Eu anomalies are present in most samples ( $Eu/Eu^* = 1.10$ - $4.93$ ), but diminish with increasing REE abundances. The most REE-rich sample

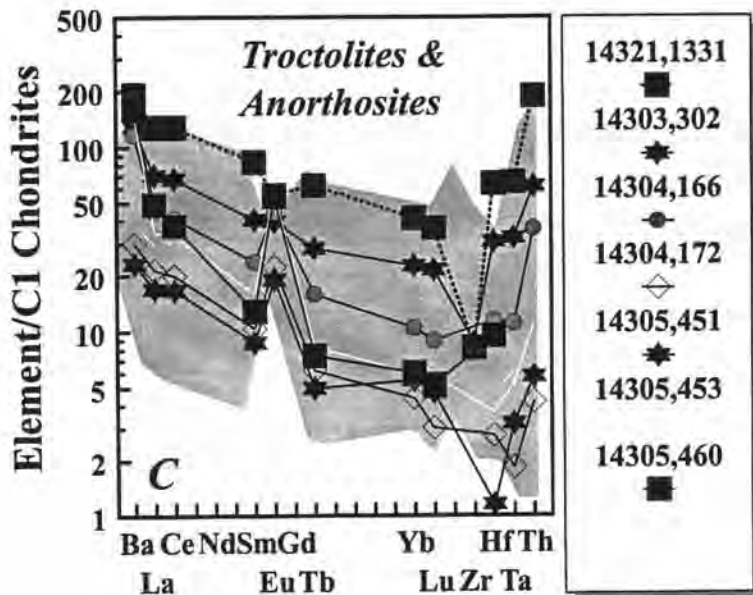


Figure 4. (continued)

[14303,323,(302)] actually contains a slight negative Eu anomaly ( $\text{Eu}/\text{Eu}^* = 0.75$ ). This is anomalous for a plagioclase cumulate. The compatible-element abundances are below those of the norites and dunites, demonstrating the paucity of mafic phases in these rocks. The HFSE abundances span the ranges of the other rock types.

A total of 25 large magnesian anorthosites and troctolites have been analyzed for trace and major elements (Table 4). Eight more troctolites/anorthosites are reported in this paper; however, only 14303,323,(302) was large enough to be analyzed for both its Nd and Sr isotopic composition (Table 2; radiogenic isotope subsample number ,347). Rb and Sr abundances in magnesian anorthosite 14303,347 are comparable to those in ferroan anorthosites [e.g., *Nyquist*, 1977]. Rb in 14303,347 (0.494 ppm) is an order of magnitude, or more, less abundant than that in alkali anorthosites (3.09 to 24.0 ppm; [*Snyder et al.*, 1995]). The measured  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio for anorthosite 14303,347 (0.69949) is slightly higher than the highest values for ferroan anorthosites [*Nyquist*, 1977], but demonstrably lower than that for alkali anorthosites (0.70031 to 0.72278; [*Snyder et al.*, 1995]) and much lower than KREEP (Figure 5; we have assumed an age of 4.2 Ga as explained later in the text). Sm and Nd abundances (12.3 and 44.0 ppm) are similar to values at the low end of the range of values for alkali anorthosites (7.01 to 41.6 and 26.1 to 149 ppm, respectively [*Snyder et al.*, 1992, 1994a, b, 1995]). The measured  $^{147}\text{Sm}/^{144}\text{Nd}$  ratio (0.168) for anorthosite 14303,347 is similar to that measured for lunar KREEP (0.168-0.169 [e.g., *Nyquist and Shih*, 1992]).

## Discussion

The crux of the debate about the origin of magnesian-suite rocks concerns the seeming discrepancy between their primitive major element and mineral-chemical compositions and their relatively evolved REE compositions. Recent advances in analytical methods have allowed the direct determination of REE abundances of lunar magnesian-suite orthopyroxenes [*Papike et al.*, 1994]. Back calculation of liquids which could have precipitated these orthopyroxenes are distinctly REE enriched and have been interpreted as picritic melts which have somehow assimilated a significant KREEP component [*Papike et al.*, 1994]. *Snyder et al.* [1995] also have shown that alkali-suite lunar highlands rocks likely are cumulates from evolved pristine KREEP basalt magmas. Further, they hypothesized that magnesian-suite cumulates also could have precipitated from pristine KREEP basalts, prior to the crystallization of alkali-suite rocks [*Snyder et al.*, 1995].

### Cumulates From Primitive KREEP Basalt Crystallization

To test this hypothesis, compositions of relatively primitive pristine KREEP basalts were compiled as per *Snyder et al.* [1995]. The formation of pristine KREEP basalt is not within the scope of this paper; however, it is speculated that these basaltic melts are the products of deep melting in the lunar interior and subsequent assimilation of primitive KREEPy residual liquids during passage of the basalt to the lunar surface. These assimilated KREEPy liquids were residual from the original magma ocean and could have been trapped within upper level mantle cumulates. Those basalts which

Table 4. Whole Rock Chemistry of "Large," Probably Pristine, Magnesian Troctolites/Anorthosites

	Parent												
	12071	14172	14179	14303	14304	14305	14305	14305	14305	14305	14321	14321	14321
Clast	c10	c11	c6	c194	c95/a	c264	c279	c317/322	c358	c361	c1020	c1024	c1037
PM									w6	w7			
Mass, g	1.3	0.67	0.67	2	0.89	0.23	0.23	0.34	0.49	0.14	9.2	0.67	0.11
Pristinity <sup>a</sup>	7	6	6	6	7	7	8	7	7	6	7	7	6
SiO <sub>2</sub>	45.6	43.0	42.6	43.4	[43.3]	43.6	43.4	[43.6]	[43.7]	[44.9]	43.0	41.9	43.0
TiO <sub>2</sub>	0.22	0.06	0.28	0.03	0.17	0.04	0.05	0.13	<0.5	<0.2	0.06	0.68	0.09
Al <sub>2</sub> O <sub>3</sub>	24.9	24.2	22.9	27.0	16.6	28.0	28.5	21.9	21.5	32.3	28.7	26.4	26.1
FeO	5.02	4.08	4.44	3.16	5.53	2.83	2.25	4.2	3.8	0.74	2.59	5.06	3.71
MnO	0.077	0.046	0.051	0.034	0.075	0.031	0.026	0.049	0.049	0.010	0.025	0.045	0.038
MgO	10.8	15.4	16.1	11.77	25.2	11.44	8.29	17.6	17.7	2.3	9.45	12.4	11.8
CaO	14.3	12.9	12.5	14.4	8.68	14.3	15.9	11.7	12.0	19.1	15.11	13.9	13.9
Na <sub>2</sub> O	0.325	0.377	0.381	0.406	0.28	0.434	0.469	0.44	0.46	0.35	0.379	0.318	0.367
K <sub>2</sub> O	0.020	0.053	0.081	0.073	0.15	0.073	0.069	0.063	0.056	0.051	0.075	0.054	0.073
Mg#	79	87	87	87	89	88	87	88	89	85	87	81	85
Sc, ppm	7.8	2.99	4.23	3.88	6.4	1.78	2.6	3.9	3.2	1.4	1.69	4.5	2.4
V	<66				29			70	50	10			
Cr	710	251	303	261	1200	140	201	2400	1330	250	1000	643	730
Co	17.0	28.8	25.1	23.4	27	19.0	12.6	17.8	15.0	3.2	9.2	14.8	16.9
Ni	37	28	66	46	70	19.0	10	20	<50	<30	32	24	72
Rb					1.9								
Sr	164				106		211	120					0.21
Cs	<0.29				0.067								
Ba	27	450	260	430	270	630	450	140	120	100	250	280	320
La	1.61	7.9	33.2	31.7	7.7	5.6	6.6	6.6	6.15	6.0	25.0	22.9	24.2
Ce	4.9	18.6	82	77	17.7	12.2	14.8	15.0	15.0	15.0	61	52	67
Nd	2.2	12	52	49	8.3	7.1	8.0	9.4	8.0	9.0	38	23	41
Sm	0.69	2.61	13.8	12.0	2.24	1.56	2.16	2.00	2.00	2.20	10.5	5.9	10.8
Eu	0.89	2.34	2.1	2.32	1.42	2.72	2.6	1.45	1.35	1.35	2.07	2.3	1.88
Tb	0.145	0.52	2.6	2.31	0.48	0.20	0.40	0.40	0.35	0.45	1.9	1.27	1.99
Dy	1.10		15.7		3.6		2.33	2.6	2.4	3.0	10.4	8.5	13.2
Yb	0.63	1.58	8.3	8.4	3.0	0.83	1.44	1.7	1.60	1.45	4.2	9.6	4.9
Lu	0.095	0.25	1.19	1.15	0.47	0.13	0.19	0.25	0.23	0.19	0.56	1.56	0.73
Zr	<90	350	310	260	53	120	<50	<30			38	350	560
Hf	0.45	0.75	5.9	4.5	1.27	0.25	0.37	0.88	0.92	0.92	0.15	8.8	10.3
Ta	0.07	0.18	0.57	0.45	0.131	<0.3	0.071	0.050	0.063	0.13	0.037	1.77	0.18
Th	0.16	0.75	3.7	3.7	0.79	0.22	0.46	0.35	0.4	0.85	2.0	2.6	2.27
U	<0.24	0.13	0.78	0.55	0.17	0.17	0.15	<0.1	<0.2	0.11	<0.4	1.6	0.27
Ir, ppb	0.080	0.07	0.14	0.13	<1.1	<0.05	0.009				0.053	0.031	0.58
Au	0.038	0.030	0.22			0.016	0.190				0.17	0.031	0.058
Reference	1	2	3	2	4	2	5	6	7	7	3	3	8

did not reach the surface could have crystallized within the upper lunar crust.

Using the program MAGFOX (J. Longhi, personal communication, 1992), fractional crystallization was modeled to its end point using an average, pristine, KREEP magma under 1 bar of pressure [Snyder *et al.*, 1995]. This fractionation sequence was then used to calculate mineral compositions and trace element compositions of successive liquids, using the Rayleigh fractionation equation

$$C_L = C_0 F^{(D-1)} \quad (1)$$

(where,  $C_L$  is concentration of a trace element in the liquid,  $C_0$  is concentration of a trace element in the original liquid,  $F$  is fraction of liquid remaining, and  $D$  is the bulk minerals/melt distribution

coefficient for the particular trace element of interest). The trace-element contents of associated cumulates were calculated using the simple, equilibrium partition-coefficient relation

$$C_S = DC_L \quad (2)$$

(where  $C_S$  is concentration of the trace-element in the solid, cumulate).

The calculated cumulate evolution path is indicated in Figure 6, where the An content of plagioclase is plotted relative to the Mg# of either olivine or orthopyroxene (this shaded path is that for accumulative compositions). As can be seen, the first cumulates to crystallize from a KREEP basalt magma are enriched in Mg (for mafic minerals) and Ca (for plagioclase) and, thus, quite primitive. In fact, the crystallization of a primitive KREEP basalt from the

Table 4. (continued)

	Parent											
	14321	14321	15426	15455	65785	72705	73146	76335	76504	76504	76535	76536
Clast	c1140	c1211	c137	c106	c4	c1			.12	.15		
PM												
Mass, g	0.10	0.10	0.10	3.0	0.34	0.49	3.0	465	0.27	0.13	155	10.3
Pristinity <sup>a</sup>	6	6	6	7	8	8	7	8	8	8	9	7
SiO <sub>2</sub>	[36.4]	[42.6]	[42.1]	44.3	41.1	42.1	43.0	43.4	44.3	44.1	42.9	42.4
TiO <sub>2</sub>	0.06		0.31	0.08	0.07			0.07	0.05	0.07	0.05	
Al <sub>2</sub> O <sub>3</sub>	15.0	35.2	12.7	21.9	29.9	23.8	30.0	29.4	29.3	28.3	20.7	26.3
FeO	8.55	0.430	7.47	5.79	3.7	3.99	2.32	2.61	2.06	2.44	4.99	3.60
MnO	0.068	0.008		0.072	0.03	0.040	0.022	0.032	0.030	0.035	0.07	0.040
MgO	30.5	2.08	31.3	16.1	9.6	16.7	7.63	9.62	8.29	9.78	19.1	13.6
CaO	9.2	19.1	5.8	11.6	14.8	12.6	16.2	15.9	16.4	15.4	11.4	13.3
Na <sub>2</sub> O	0.198	0.50	0.177	0.235	0.29	0.305	0.340	0.315	0.336	0.346	0.2	0.288
K <sub>2</sub> O			<0.14	0.045	0.04	0.064	0.059	0.036	0.039	0.045	0.03	0.041
Mg#	86	90	88	83	82	88	85	87	88	88	87	87
Sc, ppm	3.46	1.42	5.41	4.1		2.6	1.01	1.53	1.59	2.25		1.8
V									26	18		
Cr	397	71	630	970	1200	920	270	382	490	680		530
Co	22.2	1.22	38.9	25		25	8.7	14.4	13.1			20
Ni	44		110	26		18	100	20.4	24	7		5.0
Rb			<10							1.8	0.24	
Sr	127	240	80						158	164	114	
Cs			0.05						0.330	0.122		
Ba	152	460	180	77		170	58	51	40	34	33	49
La	10.2	231	17.1	3.2		8.8	2.6	2.30	1.68	2.24	1.51	1.9
Ce	24.8	620	38.3	8.1		19.3	6.2	6.0	3.5	5.3	3.8	4.1
Nd	14.8	410	15.7	4.4		11	3.3		2.25	3.1	2.3	2.5
Sm	3.80	110	4.10	1.23		2.62	0.83	0.75	0.59	0.81	0.61	0.65
Eu	1.31	3.73	1.23	0.82		1.32	0.99	0.97	0.97	1.05	0.73	0.78
Tb	0.76	23.6	0.80	0.25		0.36	0.13	0.13	0.095	0.159		0.13
Dy									0.61	1.11	0.80	
Yb	2.20	55.3	5.69	1.2		1.7	0.46	0.56	0.47	0.68	0.56	0.44
Lu	0.356	7.84	1.00	0.17		0.25	0.051	0.078	0.055	0.079	0.079	0.062
Zr	<40		1030		74	190			13.1	45	23.6	
Hf	0.198	0.68	22.9	0.86		1.6	0.27	0.43	0.25	0.42	0.52	0.36
Ta	0.055	0.10	2.73	0.14		0.062			0.033			0.031
Th	0.75	30.0	2.4	0.58		0.32	0.17	0.16	0.037	0.58		0.20
U	0.21	2.8	3.64	0.18		0.12		0.10				
Ir, ppb			<2	0.024		0.016	0.130	0.07	<0.050	<0.020		0.051
Au			<1	1.9		0.170	0.690	0.051	0.018	0.089		0.020
Reference	9	9	10	11	12	11	11	13	14	15	16	11

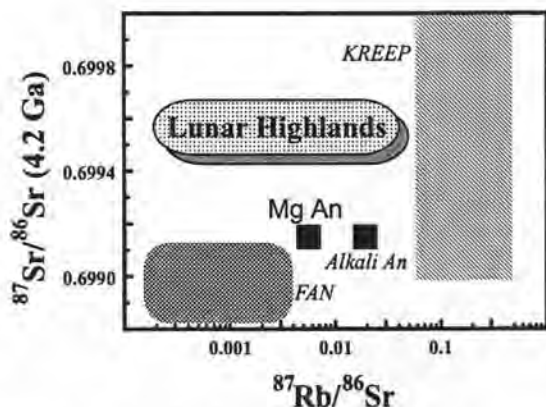
<sup>a</sup> Pristinity index (1-10 from least to most likely pristine) from Warren [1993].

References: 1, Warren et al. [1990]; 2, Warren and Wasson [1980]; 3, Warren et al. [1981]; 4, Goodrich et al. [1986]; 5, Warren et al. [1983b]; 6, Shervais et al. [1983]; 7, Shervais et al. [1984]; 8, Warren et al. [1983a]; 9, Lindstrom et al. [1984]; 10, Lindstrom et al. [1988]; 11, Warren and Wasson [1979a]; 12, Dowty et al. [1974]; 13, Warren and Wasson [1978]; 14, Warren et al. [1987]; 15, Warren et al. [1986]; 16, Ryder and Norman [1979].

outset to about 55% crystallization yields cumulates which would define the whole field of magnesian-suite rocks. The next 45% of crystallization precipitates minerals with compositions which would fall in the range of alkali suite rocks (see Figure 6 and Snyder et al. [1995]). Addition of residual KREEP basalt liquid to a crystallizing mineral assemblage will draw compositions off of this array and result in Fe-enriched mafic minerals and more Na-rich plagioclase (see small arrows on Figure 6). Thus, through simple fractional crystallization and addition of trapped residual liquid, the pyroxene,

olivine, and plagioclase compositions of magnesian-suite cumulates can be reproduced.

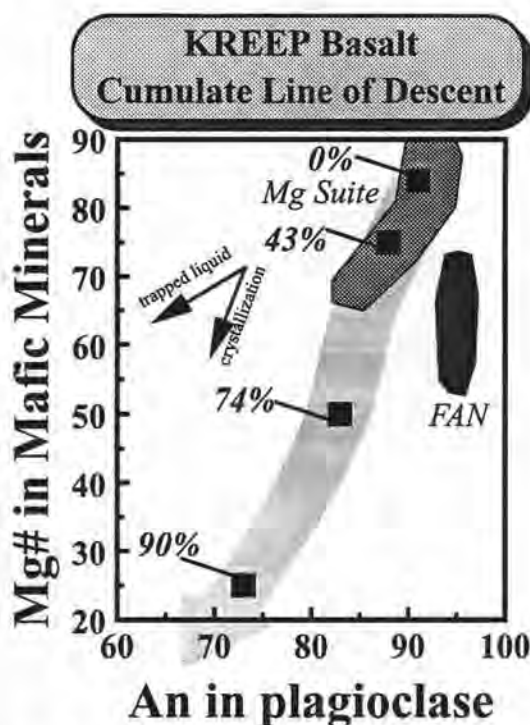
The trace element composition of these successive KREEP basalt residual liquids after 0% (low) and 55% (high) crystallization are shown in Table 5 and compared to equilibrium liquids backcalculated from trace-element data on magnesian-suite orthopyroxenes [Papike et al., 1994]. Note that our calculated KREEP basalt residual liquids are within the range of the parental liquids calculated using the data from Papike et al. [1994]. Although the bulk  $D_{Sr}$  appears to be near unity for both sets of



**Figure 5.** Plot of  $^{87}\text{Rb}/^{86}\text{Sr}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$  (at 4.2 Ga reference age) for magnesian anorthosite clast 14303,347 relative to fields for the ferroan anorthosites (FAN; [Nyquist, 1977]), KREEP [Nyquist, 1977; Nyquist and Shih, 1992], and an alkali anorthosite [Snyder et al., 1995]. Notice that the scale on the abscissa is in log units. KREEP may also extend to higher  $^{87}\text{Sr}/^{86}\text{Sr}$  values.

calculated liquids (if those derived from Papike et al. [1994] are successive, related, liquids), the absolute values are different by a factor of 2. Furthermore, the mathematically constructed KREEP values of Warren [1989] also lie approximately at the midpoint of our KREEP basalt differentiation path (i.e., halfway between the S(low) and S(high) values). Thus, the average KREEP values determined by Warren [1989] may represent an average parent magma for the magnesian-suite cumulates. Note that the magnesian-suite cumulates studied by Jolliff et al. [1993] have trace element compositions which are comparable to our calculated liquids, and are even higher than KREEP. Thus, liquids calculated to be in equilibrium with the supposedly magnesian-suite cumulates in Jolliff et al. [1993] must have had even higher trace element abundances (probably a factor of 2 to 4 higher based on the fractionation assemblages used by Jolliff et al. [1993]). In fact, even the impact melt used by Jolliff et al. [1993] to model equilibrium crystallization is 3x higher than KREEP and 1.6 to 2.2x higher than the most evolved parent liquids calculated for magnesian-suite cumulates by Snyder et al. [1995] and Papike et al. [1994]. Thus, parent liquids for the clasts studied by Jolliff et al. [1993] are demonstrably more evolved in their trace element compositions than either KREEP or an evolved KREEP basalt. Furthermore, trace element abundances of these few magnesian-suite clasts are a factor of 5-10 higher than any of the large, probably pristine magnesian-suite rocks compiled in Tables 4 and 6, making them of questionable significance in understanding this suite. The most cogent of their arguments, that the presence of whitlockite in a few samples is due to equilibrium crystallization and "local melt-pocket equilibrium," will be discussed below.

Although the fragments and clasts of mineral aggregates presented within often are called rocks, they are typically of very small dimensions. For instance, note that the masses of the samples in Table 3 are mostly less than 100 mg. Therefore, the modal mineralogy of these samples may not be representative of a larger rock body, but could indicate a segregation of a particular mineral assemblage. Many others have noted this insidious problem in the



**Figure 6.** Plot of An content in plagioclase versus Mg# in orthopyroxene and/or olivine for cumulates from an average Apollo 15 KREEP basalt (as per Snyder et al. [1995]). Squares indicate the percentages of crystallization of the KREEP basalt parent. Arrows are for crystallization and addition of a trapped, residual, KREEP-basalt liquid.

**Table 5.** Calculated Trace Element Compositions (in ppm) of Parental Melts for Magnesian Suite Cumulates Compared to KREEP and Whitlockite-Bearing Cumulates

	P(low)	S(low)	P(high)	S(high)	KREEP	J(whit)
La	52	69	200	148	110	325
Ce	144	180	360	387	280	836
Nd	81	105	167	226	178	547
Sm	31	30.3	78	65	48	150
Eu	0.74	2.65	1.32	3.1	3.3	4.3
Tb		6.2		13.4	10.0	28
Dy	44	38	164	82	65	181
Er	26	25	102	53	40	112
Yb	20	20.5	87	43	36	101
Lu		2.95		6.2	5.0	14
Zr	861	900	1580	1930	1400	
Sr	97	200	98	187	200	

Sources for calculated "data" and cumulate data: P (low and high) from Papike et al. [1994]; S (low and high) from Snyder et al. [1995]; KREEP, Warren [1989]; J (whit), whitlockite-bearing "cumulate" 14161,7233 used by Jolliff et al. [1993] as a starting liquid composition for equilibrium crystallization.

**Table 6.** Whole Rock Chemistry of "Large", Probably Pristine, Magnesian Gabbros, Norites, and Ultramafic Cumulates

	Parent									
	14318	14318	15360	15361	15445	15455	15565	61224	67667	67915
Clast	c146	c150	.11		c17 "b"	c228	c113	,6		c163
PM										
Rock	N	N	N	N	N	N	N	GN	GN	GN
Mass, g	1.2	0.49	0.67	0.90	10.0	200	0.34	0.34	7.9	0.23
Pristinity <sup>a</sup>	6	6	9	8	8	9	8	8	7	6
SiO <sub>2</sub>	49.4	46.6	47.7	49.8	48.7	44.4	48.1	50.7	42.4	56.4
TiO <sub>2</sub>	0.62	1.07	0.117	0.182	0.15	0.1	0.10	0.40	1.03	5.2
Al <sub>2</sub> O <sub>3</sub>	18.7	23.2	24.7	18.1	23.8	26.2	24.9	13.2	7.56	11.5
FeO	7.98	5.57	4.37	5.40	3.84	4.2	4.05	9.91	17.1	12.4
MnO	0.106	0.076	0.079	0.080	0.08	0.048	0.074	0.159	0.201	0.160
MgO	12.3	8.95	8.46	15.6	9.82	10.9	8.79	12.8	26.4	3.2
CaO	10.1	13.2	13.3	9.5	12.9	14.3	13.3	11.6	5.32	9.3
Na <sub>2</sub> O	0.662	1.07	0.372	0.324	0.33	0.36	0.425	0.91	0.158	1.29
K <sub>2</sub> O	0.228	0.380	0.061	0.034	0.059	0.07	0.060	0.017	0.023	0.51
Mg#	73	74	78	84	82	81	79	70	73	32
Sc, ppm	12.8	11.7	8.2	9.4		5.33	6.7	20.8	24.4	32
V				03		16				3
Cr	1080	875	1820	2300	1561	440	1510	1990	2590	200
Co	20.3	15.4	13.3	23.1		10	20.0	23.6	26	6.6
Ni	52	21	1.25	28		12	8.8	8	4.4	
Rb	6.1	7.9		<5	1.13	1.1				
Sr		261	124	107	138	270				260
Cs	0.29	0.21	<0.13	0.14		0.13				
Ba	470	690	62	41	59.7	42	68	30	51	340
La	15.6	58	3.23	2.96	5.11	3.0	4.25	1.47	3.6	22.7
Ce	39	160	8.1	8.0	12.6	6.7	9.5	4.3	9.6	54
Nd	21.1	93	2.8	3.5	7.79	3.73	5.4	<9	7.6	37
Sm	5.7	26.7	0.90	1.15	2.13	0.88	1.38	0.87	2.09	11.0
Eu	1.96	4.2	1.19	0.78	1.07	1.67	1.36	1.43	0.78	2.42
Tb	1.40	5.4	0.213	0.25		0.14	0.24	0.22	0.45	2.2
Dy	9.9	34.5	1.5	1.9	2.69	0.84	1.9			15
Yb	8.9	15.3	1.13	1.27	1.48	0.36	1.35	1.06	2.2	8.5
Lu	1.49	2.31	0.160	0.19	0.213	0.06	0.19	0.16	0.32	1.27
Zr	370	780	<65	42	115	11	<260	170		300
Hf	8.0	19.2	0.43	0.80	4.1	0.17	0.69	0.55	1.4	8.6
Ta	0.74	1.81	<0.10	0.13		0.14		0.16	0.20	2.2
Th	3.8	7.60	<0.11	0.43		0.23	0.43	0.19	0.46	3.9
U	1.25	2.1		0.11	0.54	0.05	0.31	<0.6	0.15	1.0
Ir, ppb	0.200	0.096	0.042	0.008	0.072	0.020	<0.03	0.148	0.013	
Au	0.270		0.065	0.25	0.022	0.023	0.069	0.079	0.029	0.044
Reference	1	2	3	3	4,5	6	7	8	9	10

interpretation of lunar "rocks" which are truly clasts from breccias and fragments from soils [e.g., *Wilshire and Jackson, 1972*]. Therefore, we have taken successive residual liquids calculated using the Rayleigh fractionation equation and calculated dunitic, anorthositic, and noritic adcumulates from these liquids, using equation (2), at 0% crystallized and 55% crystallized (the point at which the mineral compositions of the most evolved magnesian suite rocks plot; Figure 6). Finally, we have calculated orthocumulates with variable proportions of trapped residual liquid at 0% and 55% crystallized. These plots are shown in Figure 7 for dunitic (100% olivine), anorthositic (90% plagioclase + 10% pigeonite), and norites (70% plagioclase + 30% orthopyroxene).

Compare these modeled plots with the data of our "new" magnesian-suite clasts (Figures 4a, b, c) and all other large, pristine magnesian suite rocks (Figures 8a, b, c). It is evident that nearly all magnesian-suite rocks can be explained by simple fractional crystallization of a primitive KREEP basalt and addition of  $\leq 40\%$  trapped instantaneous residual liquid. However, one element which is poorly modeled by the crystallization-trapped liquid scenario is Zr. This could be due to several factors, including the precipitation of small amounts of zircon and/or ilmenite, which are not included in the KREEP basalt crystallization assemblage until after 74% crystallization [*Snyder et al., 1995*]. However, assuming that KREEP basalt is a suitable parent magma, this modeling does set an

Table 6. (continued)

	Parent										
	72255	73255	76034	77035	77075/ 77215	78234	78235/ 78255	12033	14304	14321	72415- 72418
Clast PM	c42	c27,45	,9	c130		,1		,503	c121 "d"	c1141	
Rock	N	GN	N	N	N	N	N	HZ	UM	D	D
Mass, g	10.0	0.9	0.16	100	840	0.37	395	0.10	0.14	0.10	55.2
Pristinity <sup>a</sup>	8	7	8	7	8	8	8	8	6	6	9
SiO <sub>2</sub>	52.0		44.7	50.3	51.2	50.9	48.5	[49.0]	[35.3]	[33.9]	40.6
TiO <sub>2</sub>	0.3		0.050	0.200	0.33	0.25	0.117	<0.63	<0.8	0.080	0.03
Al <sub>2</sub> O <sub>3</sub>	15.5		33.8	19.1	15.0	14.4	23.3	0.45	<1.3		1.3
FeO	7.4		0.347	5.79	10.7	7.33	4.34	8.10	9.78	11.55	11.9
MnO	0.122		0.006	0.096	0.170	0.127	0.071	0.110	0.107	0.123	0.11
MgO	15.9		1.43	11.9	13.0	16.4	9.62	41.5	51.4	53.7	45.4
CaO	9.1		19.0	11.8	8.8	9.23	12.9	0.210	<1.3		1.1
Na <sub>2</sub> O	0.33		0.357	0.044	0.382	0.253	0.388	0.024	0.043	0.023	0.02
K <sub>2</sub> O	0.08		0.052	0.088	0.18	0.055	0.066	0.008	0.016		
Mg#	79		88	79	68	80	80	90	90	89	87
Sc, ppm	13.2		0.66	10.9	16.6	13.0	7.7	3.2	3.0	5.02	4.3
V			<15			94		48	3.7		50
Cr	1100		140	2220	2650	2770	1600	2550	1470	522	2300
Co	29		2.0	22	33.0	29.3	26.8	54	63	61.3	55
Ni			7.5	9.5	6.1	11.5	21.9	194	315	70	160
Rb	1.27		[4]						<13		
Sr			183			107			<129	<30	8.2
Cs	0.058		0.057			0.162					
Ba			50	96	160	53	76	<35	<84	24	4.1
La	16		1.94	5.5	7.2	3.3	3.4	1.19	2.02	5.14	0.15
Ce	46		4.0	13	22	8.6	7.9	3.6	5.3	13.5	0.37
Nd			2.3	8.6	8.5	4.5	4.5	<4	3.1	8.4	
Sm	7.6		0.64	2.19	3.0	1.49	1.28	0.49	0.88	2.55	0.080
Eu	1.75		1.13	0.093	0.98	0.70	1.02	0.061	0.080	0.065	0.061
Tb	1.9		0.087	0.49	0.74	0.38	0.27	<0.19	0.164	0.503	0.017
Dy			0.56			2.73			1.4		0.11
Yb	6.6		0.25	2.2	3.9	2.33	1.34	0.38	0.66	1.98	0.074
Lu	1.01		0.023	0.32	0.59	0.35	0.20	0.054	0.096	0.391	0.012
Zr			7.9		210	29			<200	<70	
Hf	5.5		0.152	1.9	3.5	1.66	0.84	0.27	0.57	0.93	0.010
Ta			<0.06	0.20	0.34	0.25	0.093	<0.16	<0.2	0.10	
Th			0.033	1.1	1.57	0.62	0.49	0.18	0.28	0.71	
U	0.240			0.31	0.5	0.22	0.17	<0.38	0.074	0.090	
Ir, ppb	0.0040		0.024	0.050	0.25	<0.015	0.570	<0.063	<6		
Au	0.008		0.056	0.026	0.026	0.015	0.140	0.019	<3.8		
Reference	5		10	9	5	10	9	3	10	11	3

<sup>a</sup> Pristinity index (1-10 from least to most likely pristine) from Warren [1993].

References: 1, Warren et al. [1983a]; 2, Warren et al. [1986]; 3, Warren et al. [1990]; 4, Shih et al. [1993];

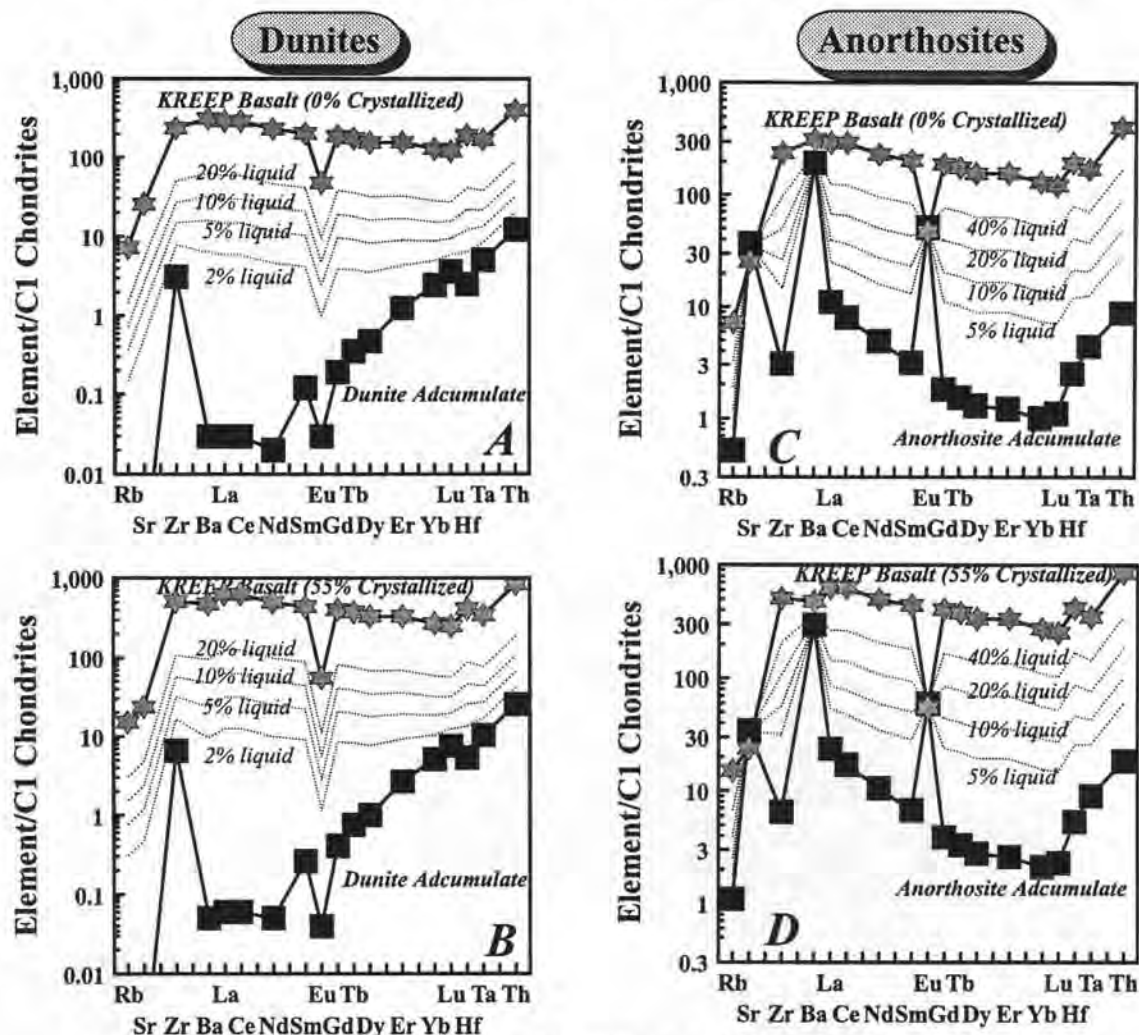
5, Ryder and Norman [1979]; 6, Ryder [1985]; 7, Warren et al. [1981]; 8, Marvin and Warren [1980]; 9, Warren and Wasson [1979a]; 10, Marti et al. [1983]; 11, Lindstrom et al. [1984].

upper limit on the trace element abundances of magnesian-suite cumulates.

The most primitive magnesian-suite cumulates exhibit mafic minerals with Mg# in excess of 90. Liquids in equilibrium with these mafic minerals must have Mg# of approximately 70-72. KREEP basalts are the most evolved, yet primitive, samples on the Moon which are considered to be near-liquid compositions. However, even the most primitive and pristine KREEP basalts do not exceed an Mg# of 66 (15303.51 [Simon et al., 1988]) and must

have undergone a small amount of crystallization prior to, or subsequent to, extrusion. Also, if a rock is to be considered a cumulate, by definition it must contain less than 50% trapped liquid. Therefore, the most evolved magnesian-suite cumulate which can be generated would have precipitated after 55% fractional crystallization (see above) combined with 50% trapped residual liquid. Such cumulates cannot exceed approximately 200x and 100x C1 chondrites for the LREE and HREE, respectively (see Figures 7a-7f). Such is indeed the case for all magnesian-suite





**Figure 7.** Semilog plots of trace elements, relative to C1 chondrites, for residual KREEP basalt liquids (at 0% and 55% crystallized), a model equilibrium accumulate, and orthocumulates with 5%, 10%, 20%, and 40% trapped liquid: mixtures of residual KREEP basalt (a) after 0% crystallization with a dunite accumulate, (b) after 55% crystallization with a dunite accumulate; (c) after 0% crystallization with an anorthosite accumulate, (d) after 55% crystallization with an anorthosite accumulate; (e) after 0% crystallization with a noritic accumulate, and (f) 55% crystallization with a noritic accumulate.

rocks and clasts presented within. In fact, most magnesian suite rocks can be interpreted as precipitates with less than 20% trapped residual liquid.

#### The Presence of Whitlockite: Metasomatism Versus Equilibrium Crystallization

One of the major petrogenetic problems concerning the western magnesian suite is its large range in incompatible trace element abundances relative to the "eastern" sites. There are small clasts that are magnesian in character which exhibit much higher abundances of the REE. Most notable in this regard is the gabbronorite (14161,7044) presented by Jolliff *et al.* [1993] as evidence against

metasomatism of the lunar highlands. Another sample, an Apollo 14 anorthosite from the magnesian suite contains 231 ppm La [Lindstrom *et al.*, 1984] due to the presence of a REE-rich (evolved) phosphate in this primitive lithology [Lindstrom *et al.*, 1985]. Whether these phases were included or omitted during whole rock chemical analyses due to sampling problems cannot be unequivocally confirmed or denied. Neal and Taylor [1990, 1991a] suggested that phosphates in lunar highlands rocks are not products of intercumulus melt crystallization and are exotic.

It has been demonstrated that silicate liquid immiscibility (SLI) can occur in an igneous system after 85-95% "Fenner trend" crystallization [e.g., Hess *et al.*, 1975; Rutherford *et al.*, 1976]. This

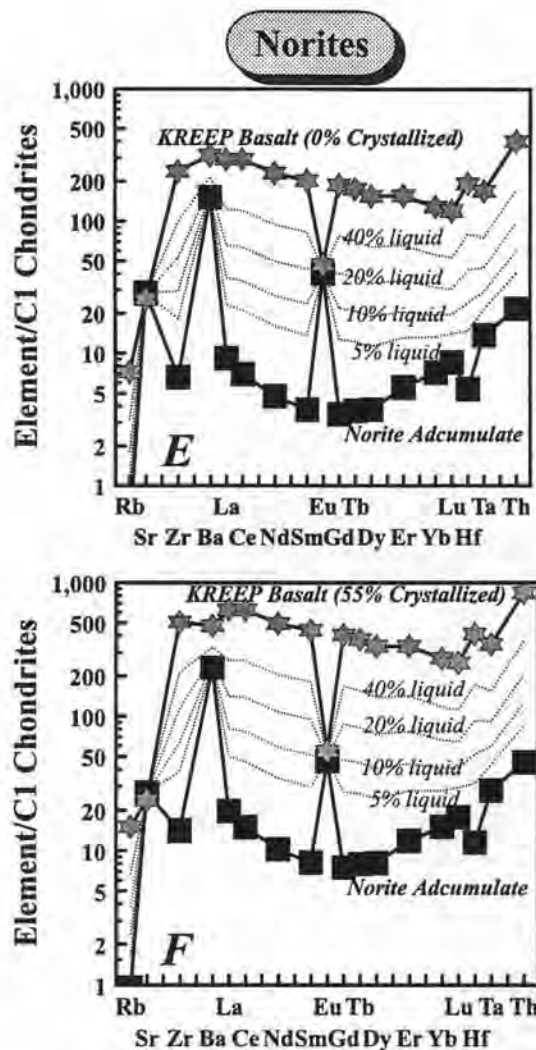


Figure 7. (continued)

process produces granitic (high SiO<sub>2</sub>) and ferrobaltic (high Fe) immiscible melts. Neal and Taylor [1989a, b] have also proposed that some urKREEP underwent SLI as it presumably was the product of extreme fractionation of the lunar magma ocean (LMO). This process could have split the KREEP signature into a K fraction (granitic) and a REEP fraction (ferrobaltic). The viscosity differences between these two (30,000 versus 11 poise, respectively; [Neal and Taylor, 1989b]) facilitates separation, perhaps by grain boundary percolation of the REEP-fraction. Neal and Taylor [1989b] postulated that post-SLI fractionation of olivine and ilmenite would allow the REEP-fraction to rise and metasomatize the lunar highlands. Therefore, we might see a REEPy signature, defined below, in some highlands rocks and not a KREEPy signature.

The REEP fraction is enriched in the incompatible elements, including the HFSE, REE, and P [Watson, 1976; Ryerson and Hess, 1980]. However, as it is a product of extreme fractional crystallization, especially of plagioclase, it contains a large negative Eu anomaly [Neal and Taylor, 1989b]. Such fluid will either cryptically (e.g., elevate REE contents, except Eu, of the original minerals) or patently (addition of a mineral phase, such as phosphate) metasomatize the surrounding rocks. Observations of metasomatized terrestrial rocks demonstrate that if only mild patent metasomatism occurs, usually only one metasomatic mineral is deposited [e.g., Menzies and Hawkesworth, 1987; Neal, 1988]. Within the Earth's mantle, this metasomatic mineral is usually hydrous, either amphibole or phlogopite. However, Menzies and Wass [1983] documented the occurrence of mantle xenoliths which had been patently metasomatized with apatite. These authors postulated that degassing of mantle CO<sub>2</sub> was responsible for causing apatite to crystallize and metasomatize these xenoliths. Furthermore, if allowed to equilibrate, the metasomatic and primary minerals will exhibit no chemical zonation, nor rimming of the primary minerals by those of metasomatic origin [e.g., Neal, 1988]. Rather, the metasomatic mineral(s) appear as part of the primary assemblage.

Lunar metasomatism would be unlike terrestrial examples, as water is absent. Rather, we envisage a low-viscosity silicate melt as the metasomatic fluid. Dickinson et al. [1989] proposed lunar metasomatism of mare basalt source regions by a halogen-rich silicate melt to explain the Ge abundances in these basalts. Neal and Taylor [1991a] have modeled lunar whitlockites as the crystallization of REEP fraction after SLI and have stated that

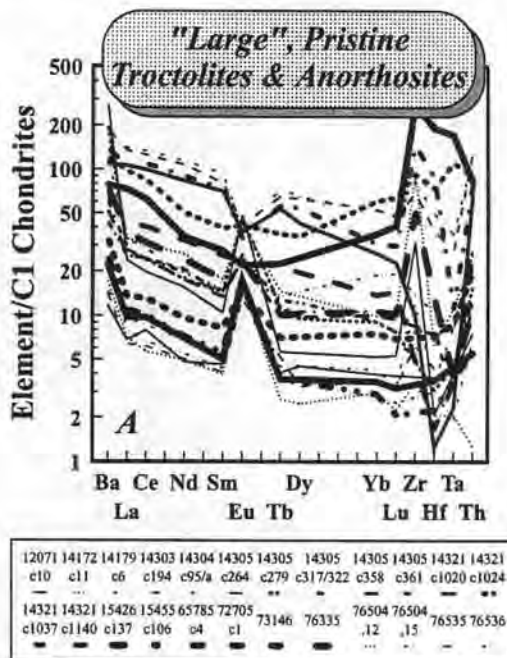


Figure 8. Semilog plots of selected trace elements, relative to C1 chondrites, for all "large," probably pristine, magnesian-suite highlands rocks (Tables 4 and 6): (a) troctolites and anorthosites, (b) gabbros and norites, and (c) ultramafic rocks.

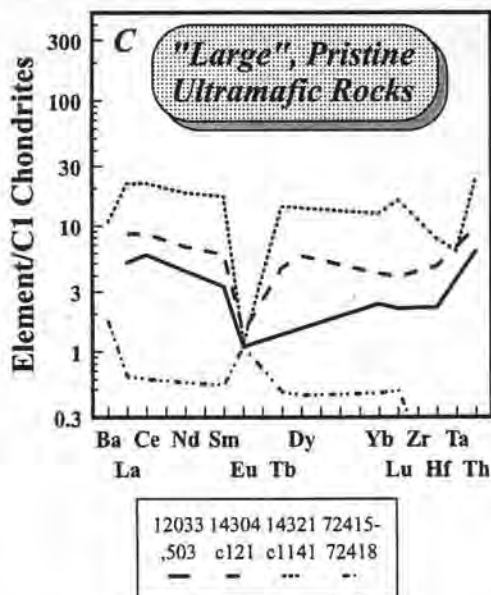
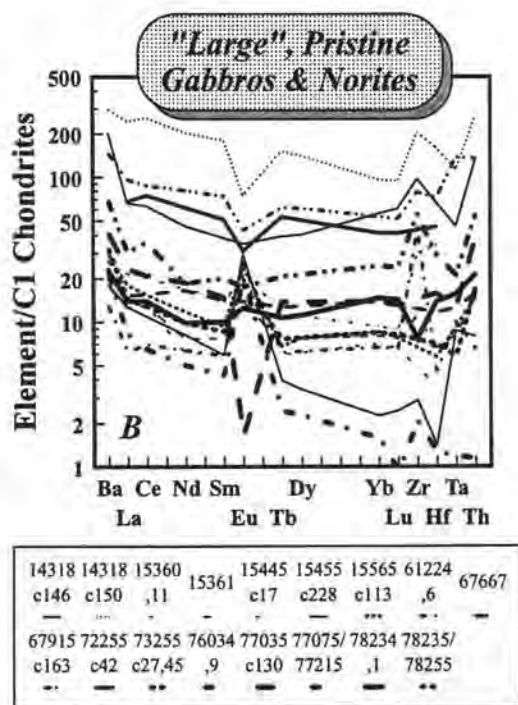


Figure 8. (continued)

whitlockites could not have been produced by crystallization of the mesostasis in primitive highlands rocks. Jolliff *et al.* [1993] have presented evidence that suggests that lunar phosphates could be produced by equilibrium crystallization of the assemblages within which they are found. However, their interpretation does not explain subtle aspects of whitlockite mineralogy, geochemistry, and

occurrence (such as REE zoning in whitlockite [Snyder *et al.*, 1992]).

Assuming that lunar granite is indeed the product of SLI, it should be possible to estimate the trace element compositions of the REEP fraction. This estimation can be made in two ways. First, one can use the simple immiscible liquid/liquid distribution equation [Watson, 1976]

$$D_{b/a} = C_{\text{basic liquid}} / C_{\text{acidic liquid}} \quad (3)$$

(where  $C_{\text{basic liquid}}$  and  $C_{\text{acidic liquid}}$  are the concentrations of the element in the basic complement and acidic complement, respectively) and assume that an average lunar granite is equivalent to the acidic liquid complement after SLI. The second method is to perform a simple mass balance, assuming that the weight ratio of basic complement to acidic complement is approximately 4 [Hess *et al.*, 1975]. Then, the mass-balance equation becomes

$$C_{\text{urKREEP}} = 0.8(C_{\text{REEP-frac}}) + 0.2(C_{\text{K-frac}}) \quad (4)$$

(where  $C_{\text{urKREEP}}$ ,  $C_{\text{REEP-frac}}$ , and  $C_{\text{K-frac}}$  are the concentrations of the element in urKREEP, the basic complement of SLI (the so-called "REEP-frac"), and the acidic complement of SLI (the so-called "K-frac"), respectively). Chambers *et al.* [1995] performed these calculations and a discussion of their results is presented below.

Jolliff *et al.* [1993] state that their modeling of whitlockites and apatites does not require "recourse to unusual REE concentrations in the system and metasomatic or metamorphic processes." However, the "systems" which are their starting points are themselves already REE-enriched by a factor of 2-10 (e.g., Table 5) over any typical lunar lithology. Their starting compositions (they are not parental liquids, *per se*, but the bulk composition of the sample) are so

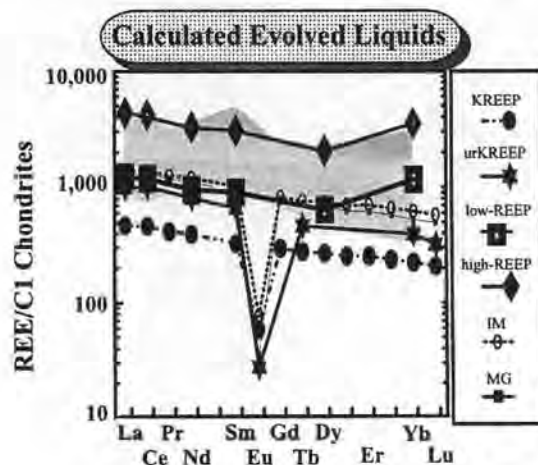


Figure 9. REE plot of calculated samples relative to C1 chondrites. KREEP, calculated by Warren [1988]; urKREEP, calculated by Neal and Taylor [1989a, b, c]; low-REEP is REEP-frac as calculated with low-P liquid-liquid  $k_D$ 's; high-REEP is using high-P liquid-liquid  $k_D$ 's; starting compositions of Jolliff *et al.* [1993]: IM, 14161,7233 impact melt; MG, monzogabbro 14161,7264; and a field for calculated liquids in equilibrium with whitlockites from Neal and Taylor [1991].

elevated in the REE, and yet, in two cases, still exhibit relatively unevolved Mg#, as to rule out their derivation as evolved late-stage liquids. In fact, their starting compositions have similar REE abundances to the calculated whitlockite equilibrium liquids of *Neal and Taylor* [1991a, b] and to the calculated composition of the REEP-fraction of SLI (Figure 9). Therefore, all that truly is proven by such modeling is that evolved liquids, which have compositions similar to the REEP fraction of SLI, are likely to precipitate whitlockite as a consequence of equilibrium processes. Therefore, it is still possible, if not likely, that whitlockite is precipitated by metasomatic fluids percolating through a consolidated/semiconsolidated magnesian-suite intrusion. It is the mere presence of these exotic phases which is intriguing.

#### Assimilation of Anorthositic Lunar Crust or urKREEP

Due to their high  $Al_2O_3$  contents, *Warren* [1986] and *Hess* [1994] have indicated that magnesian-suite troctolites cannot be precipitates of melts of the deep, primitive, lunar mantle. They postulated that magnesian-suite troctolites could be the products of primitive melts of the lunar interior which have assimilated plagioclase-rich lunar crust. *Hess* [1994] went on to prove that bulk assimilation was less favorable energetically than selective dissolution of plagioclase. Finally, he concluded that, regardless of the mechanism, only a few percent assimilation would be expected to occur. He found all other mechanisms, including convective overturn of the lunar mantle and foundering of early, bulk-Moon composition crust into a deep, primitive source region to be wanting in several aspects.

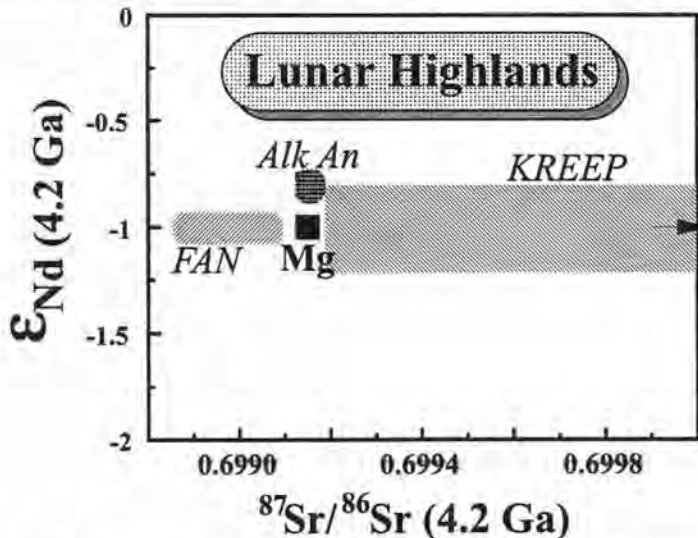
Assimilation of urKREEP has been proposed to explain the elevated REE signatures of some members of the magnesian suite at Apollo 14 [*Warren et al.*, 1981, 1983a,b; *Lindstrom et al.*, 1984]. *Warren* [1988] also presented convincing arguments that Mg-rich

melts of the lunar interior could have assimilated an urKREEP residuum from the original LMO during their upward migration from the mantle into the crust. Derivative melts are similar to KREEP basalts and retain the relatively high Mg# and high Ni coupled with the high REE indicative of this basalt type. These melts could then have precipitated cumulates to form the highlands magnesian and alkali suites. However, this alone cannot account for crystallization of magnesian-suite rocks of similar mineral compositions, but different trace element signatures. To explain this phenomenon, we favor the addition of variable proportions of trapped instantaneous residual liquid as a mechanism to enrich the incompatible elements in most magnesian-suite rocks (see above). Only those few magnesian-suite rocks with demonstrably elevated REE and which contain whitlockite show evidence of a late metasomatic addition.

#### The Diversity of Lunar Anorthosites

Based on mineral chemistry alone (Figure 2), three distinct types of anorthosites have been determined from the Moon: (1) ferroan anorthosites, uncommon in the western highlands (see *Shervais and Taylor* [1986] for a review); (2) magnesian anorthosites, of which there are only eight recognized from the western highlands [*Lindstrom et al.*, 1984, 1989; this work]; and (3) alkali anorthosites, the most common type in the western highlands (19 have been recognized; see *Snyder et al.* [1995] for compilation and references).

Based primarily on their complementary relationship to mare basalts, as well as primitive plagioclase compositions ( $An_{92-96}$ ), evolved mafic-mineral compositions, and ages of  $4.44 \pm 0.02$  Ga [*Carlson and Lugmair*, 1988] and  $4.56 \pm 0.07$  Ga [*Alibert et al.*, 1994], ferroan anorthosites are thought to be flotation cumulates from the LMO. Because of their mineral compositions and trace



**Figure 10.** Plot of  $^{87}Sr/^{86}Sr$  versus  $\epsilon_{Nd}$  (at 4.2 Ga reference age) for magnesian anorthosite 14303,347 relative to fields for ferroan anorthosites (FAN [Nyquist, 1977; Carlson and Lugmair, 1988]), KREEP [Nyquist, 1977; Nyquist and Shih, 1992; Lugmair and Carlson, 1978], and an alkali anorthosite [Snyder et al., 1994a, b, 1995]. Note that all lunar anorthosites plot within a restricted range of  $\epsilon_{Nd}$  values (-0.8 to -1.2) that is indistinguishable from lunar KREEP basalts.

element contents, alkali anorthosites are considered to be cumulates from an evolved KREEP basalt liquid [Snyder *et al.*, 1995]. A Sm-Nd age of  $4108 \pm 53$  Ma for noritic alkali anorthosite 14304,267 [Snyder *et al.*, 1995] precludes a direct relationship to the LMO and is consistent with the parent being a mixture of a primitive mafic magma and remobilized urKREEP.

Magnesian anorthosites have remained an enigmatic and much-debated lunar rock type. As stated earlier, they contain mineral-chemical compositions consistent with their derivation from a primitive source, have REE contents consistent with an evolved parentage, but are relatively depleted in the alkali metals (e.g., K and Rb). REE contents vary widely, but in all cases, the REE are at least an order of magnitude higher than those in ferroan anorthosites [Lindstrom *et al.*, 1984]. The enriched Nd isotopic signature for troctolitic anorthosite 14303,347 ( $\epsilon_{Nd} = -1.0$  at 4.2 Ga) obviates a derivation from depleted mantle cumulates alone (which would have exhibited positive  $\epsilon_{Nd}$  values at 4.2 Ga) and suggests the possible presence of some KREEP-like component, from which its Nd isotopic signature ( $\epsilon_{Nd} = -1.0$ ) is indistinguishable at this time (Figure 10). Therefore, the magnesian suite could represent remelted portions of the Moon's deep interior which have been mixed with a component which is enriched in the REE. Note that this  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio at 4.2 Ga (0.69915) is indistinguishable from that of alkali anorthosites and suggests a similar source for these two types of anorthosites (Figure 10).

#### Ages and Regional Distribution of Magnesian-Suite Rocks

Nyquist and Shih [1992] have compiled age information from the lunar highlands magnesian suite and indicate that these ages range from  $4.61 \pm 0.07$  to  $4.17 \pm 0.02$  Ga. However, ages (with the exception of one Rb-Sr determination) for the most plagioclase-rich clasts (troctolites) fall in a range from 4.27 to 4.16 Ga. A compilation of U-Pb zircon ages for Apollo 14 magnesian suite rocks (14066,47: gabbroanorthite  $4141 \pm 5$  Ma; 14305,91: norite  $4211 \pm 5$  Ma; 14306,150: troctolite  $4245 \pm 75$  Ma; 14306,60: gabbroanorthite  $4200 \pm 30$  Ma; [Meyer *et al.*, 1989]) yields a possible range of 4136 to 4320 Ma and a weighted average age of  $4191 \pm 4$  Ma. Thus, an age of  $\sim 4.2$  Ga is considered reasonable for Apollo 14 magnesian-suite rocks and is used in calculating initial isotopic ratios.

Shih *et al.* [1993] recently have determined precise Sm-Nd ages for two magnesian suite norites (15445,17:  $4.46 \pm 0.07$  Ga; 15445,247:  $4.28 \pm 0.03$  Ga) and an Rb-Sr age on one other (15455,228:  $4.55 \pm 0.13$  Ga) from the Apollo 15 landing site, which indicate that at least some magnesian-suite rocks are in excess of 4.4 b.y. old. Obviously, precise age information is needed for more magnesian-suite rocks. Current isotopic data indicates that magnesian suite rocks were produced by a variety of parental magmas over an extended period of early lunar history, although the bulk of magnesian-suite rocks are younger than the ferroan anorthosites.

Furthermore, as Figure 11 suggests, it is possible that Mg-suite magmatism was initiated at different times in different regions of the Moon. In fact, the pattern in Figure 11 could lead to the interpretation that magnesian-suite magmatism on the nearside occurred first in the northeast and then swept slowly to the southwest over a period of 300-400 m.y. In this regard, we note that the only Apollo 17 KREEP basalt for which an age has been successfully determined yielded an age in excess of those found at the Apollo 14 and 15 landing sites (to the west and southwest of Apollo 17). Granted, reliable age data for magnesian-suite rocks are sparse, especially at landing sites 14, 15, and 16. However, if this

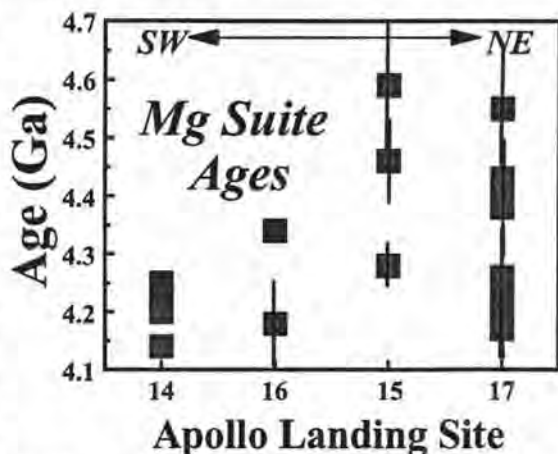


Figure 11. Plot of weighted average ages for magnesian-suite rocks relative to the landing site from which the sample was collected (data from Nyquist and Shih [1992]; and Shih *et al.* [1993]). The landing sites are placed in relative geographic order from left to right corresponding to their projection onto a southwest to northeast traverse (along a line connecting the Apollo 14 and 17 landing sites) on the nearside of the Moon.

age pattern is found to be real, we speculate that this regional progression could be due to the progressive freezing of the final residual liquids of the lunar magma ocean. This final liquid layer spread beneath the surface on the lunar near-side and was deepest where the largest concentration of urKREEP has been found -- at Apollo 14. As this final liquid layer crystallized, it did so from northeast to southwest, and culminated in the production of the most evolved KREEPy liquids beneath the Apollo 14 landing site. Radiogenic heat from the decay of isotopes of U, Th, and K then began building and could have expedited melting earliest in the northeast and progressively to the southwest on the lunar nearside. These KREEPy trapped liquids were then "melted-out" by ascending primitive melts from the deep lunar interior which subsequently fractionated and precipitated cumulates, first of the highlands magnesian suite and then of the highlands alkali suite.

#### Summary

That the Apollo 14 site is generally enriched in the incompatible elements has been well-documented through remote sensing [e.g., Wood and Head, 1975; Echevaray-Ramirez *et al.*, 1983] and the study of returned samples [e.g., McKay *et al.*, 1977, 1978; Shervais *et al.*, 1985; Dickinson *et al.*, 1985; Neal *et al.*, 1988; and Neal and Taylor, 1989b]. It is the process of enrichment of mineralogically primitive magnesian-suite lithologies which is addressed in this paper.

The LMO was probably formed soon after the Moon's birth (at  $\sim 4.5$  Ga) and persisted for 150-200 m.y.. After 70-80% of the LMO had crystallized, plagioclase became a liquidus phase and subsequently floated to form the ferroan anorthosites of the nascent lunar crust. During the latest stages of LMO crystallization, approximately at 4.36 Ga [Taylor *et al.*, 1993], the residual liquid was enriched in Fe, Si, P, K, and the REE and was trapped in the upper mantle and/or lower crust of the Moon, where it crystallized to form urKREEP. This evolved residual LMO liquid was then

"melted-out" by ascending primitive basalts of the deep lunar interior. These so-called KREEP basalts were then intruded into the upper lunar crust, where they precipitated (0-55% crystallization) cumulates of the highlands magnesian-suite. The trace element variations which are found within magnesian-suite cumulates may be the result of trapping varied amounts of residual, KREEP-basalt liquid. Later fractionation of the more evolved KREEP basalt residual liquids then formed cumulates of the highlands alkali suite.

A few highlands magnesian-suite samples are so enriched in incompatible elements as to indicate that secondary enrichment is likely. Evolved urKREEP is ripe for splitting into two immiscible silicate liquids and may have done so. Being enriched in radioactive heat-producing elements such as K, U, and Th, this urKREEP could also readily remelt at some later date and differentiate until it underwent silicate liquid immiscibility (SLI). The basic portion of SLI (which was REE- and P-rich, so-called "REEP-frac") was then available to metasomatize portions of the lunar crust.

**Acknowledgments.** This research would not have been possible without the expeditious help of the JSC curatorial staff, namely Kim Willis, Charles Galindo, and John Dietrich. Instrumental neutron activation analyses were performed by Marilyn M. Lindstrom and Lawrence A. Haskin at Washington University in St. Louis, and NASA Johnson Space Center in Houston. We thank her for recent critical analyses of her data generated during 1985-1987 and for guiding us, often not so gently, through the pristinity maze. We thank Randy Korotev and Brad Jolliff for reviewing the "beta" version of this manuscript, and Graham Ryder and Larry Nyquist for in-depth reviews of a later, abridged version of the manuscript. We also thank Paul Warren for his lengthy, often inciteful, sometimes insightful reviews; although his statements and approach are often diametrically opposed to ours. Clark Chapman and Bruce Jakosky provided timely editorial assistance as did the ever-patient Rachel Ginsburg. Though we are deeply indebted to our lunar colleagues for pushing and prodding us through helpful discussions and reviews, as always, any omissions or faults found in the published work are the sole responsibility of the authors. This research was supported by NASA grants NAG 9-62 and NAG 9-415 (to L.A.T.) and NSF grants 90-04133 and 91-04877 (to A.N.H.).

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(Received February 3, 1994; revised November 9, 1994; accepted February 14, 1995.)

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Paper number 95JE00575.  
0148-0227/95/95JE-00575\$05.00



## Correction to "Processes involved in the formation of magnesian-suite plutonic rocks from the highlands of the Earth's Moon" by Gregory A. Snyder, Clive R. Neal, and Lawrence A. Taylor

In the paper "Processes involved in the formation of magnesian-suite plutonic rocks from the highlands of the Earth's Moon" by Gregory A. Snyder, Clive R. Neal, and Lawrence A. Taylor (*Journal of Geophysical Research*, 100(E5), 9365-9388, 1995), the following figures did not reproduce well. The figures and their captions are reprinted below.

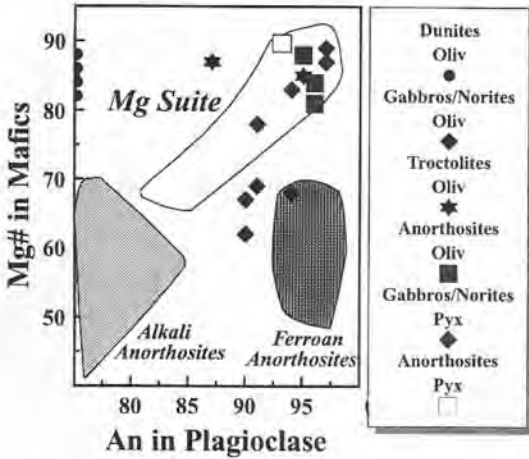


Figure 1. Plot of An content of plagioclase versus Mg# in mafic minerals for "new" magnesian suite clasts. Also shown are fields for lunar alkali anorthosites, ferroan anorthosites, and the magnesian-suite.

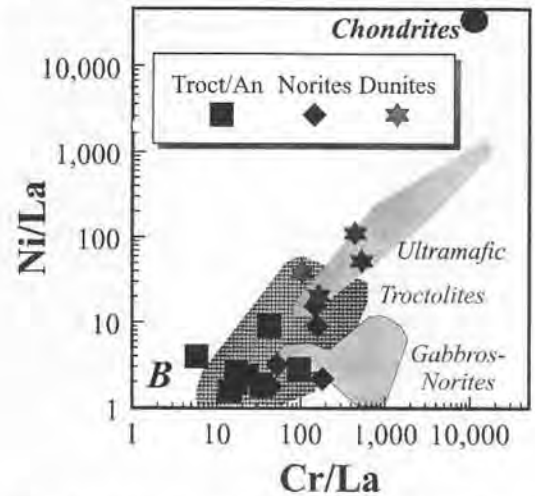
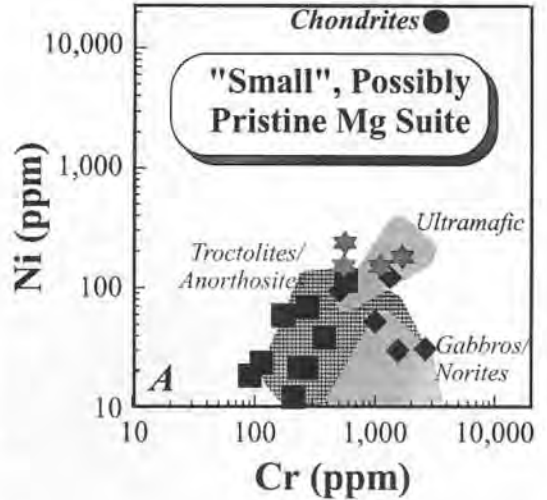
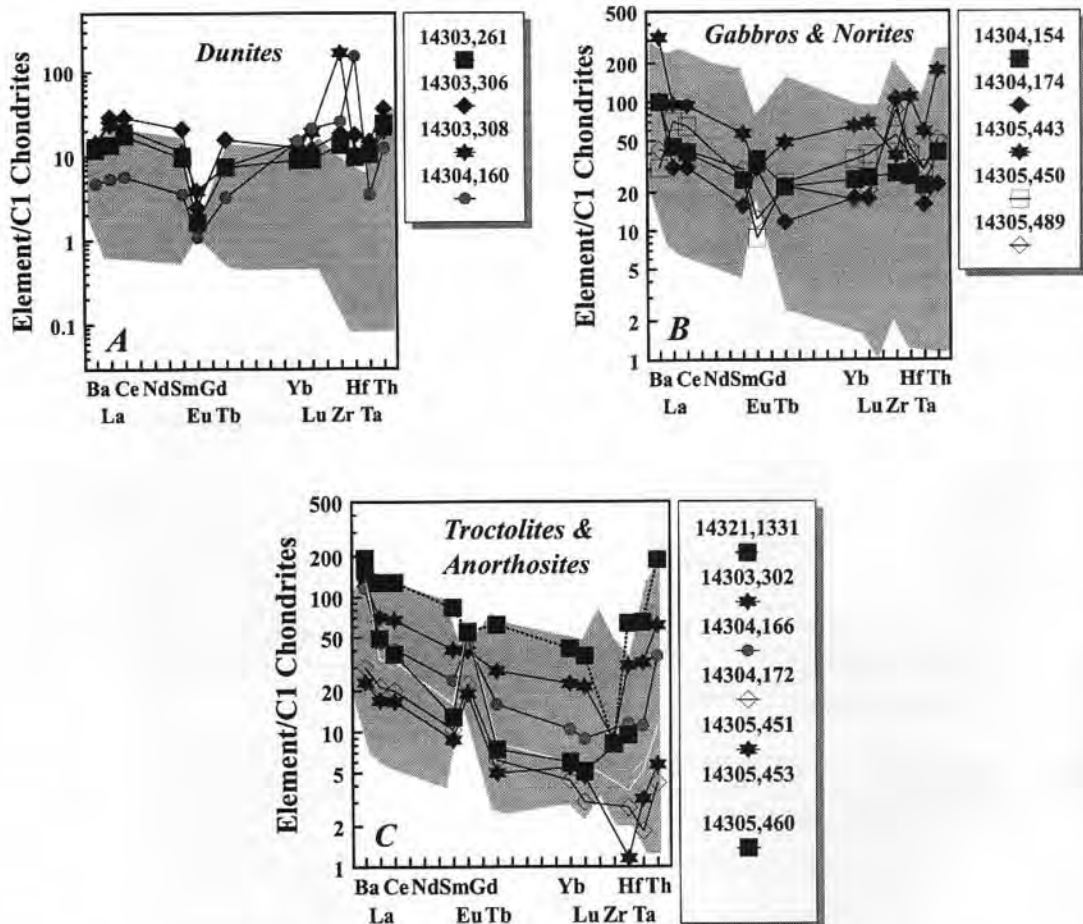


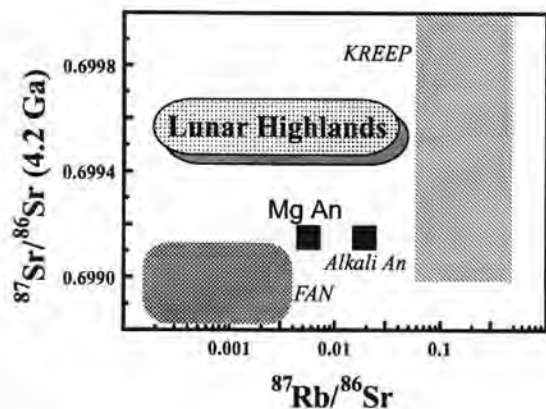
Figure 3. Log-log plots of (a) the moderately siderophile elements Cr (ppm) versus Ni (ppm) for our "new" magnesian-suite rocks, and (b) moderately siderophile/lithophile element ratios Cr/La versus Ni/La. Also shown are fields for all large (>100 mg), probably pristine gabbros/norites, troctolites/anorthosites, ultramafic rocks, and chondritic meteorites.

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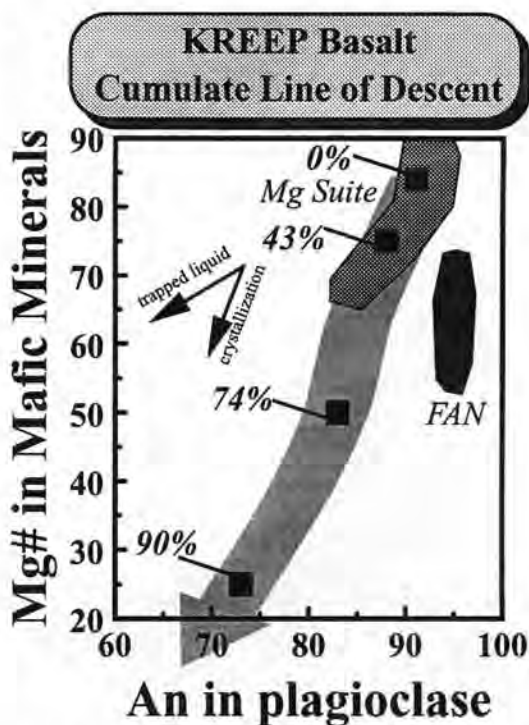
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**Figure 4.** Chondrite-normalized plots of trace-elements for magnesian suite rocks: (a) dunites; (b) norites; (c) troctolites and anorthosites. Also shown are shaded fields for all other "large," pristine, magnesian-suite rocks of that lithology (Tables 4 and 6).



**Figure 5.** Plot of  $^{87}\text{Rb}/^{86}\text{Sr}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$  (at 4.2 Ga reference age) for magnesian anorthosite clast 14303,347 relative to fields for the ferroan anorthosites (FAN; [Nyquist, 1977]), KREEP [Nyquist, 1977; Nyquist and Shih, 1992], and an alkali anorthosite [Snyder *et al.*, 1995]. Notice that the scale on the abscissa is in log units. KREEP may also extend to higher  $^{87}\text{Sr}/^{86}\text{Sr}$  values.



**Figure 6.** Plot of An content in plagioclase versus Mg# in orthopyroxene and/or olivine for cumulates from an average Apollo 15 KREEP basalt (as per Snyder *et al.* [1995]). Squares indicate the percentages of crystallization of the KREEP basalt parent. Arrows are for crystallization and addition of a trapped, residual, KREEP-basalt liquid.

(Received July 12, 1995)