Lunar anorthosites and the magma-ocean plagioclase-flotation hypothesis: Importance of FeO enrichment in the parent magma

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Abstract

The crust of the Earth's Moon comprises an anorthosite complex compared to which even the Bushveld and Dufek igneous complexes appear tiny. The origin of the lunar crust remains controversial, with current models ranging from crystallization of a "magma ocean" in the aftermath of whole-Moon melting, to purely "serial" magmatism, confined to numerous relatively small magmatic events. Excluding the volumetrically minor (and exceptionally young) mare basalts, the few texturally unaltered lunar rocks generally appear to be anorthositic, noritic, or troctolitic cumulates. Compositionally "pristine" rocks (i.e., those not mixed by impact brecciation) exhibit a striking geochemical bimodality. The bimodality is most clearly manifested by a tendency for one group (roughly half of the samples) to have far lower Mg/(Mg + Fe) ratios than rocks from the other group having comparably primitive Na/Ca, Eu/Al, etc. Rocks of the former group have traditionally been termed ferroan anorthosites, whereas rocks of the latter type are designated as Mg-rich, or Mg-suite, rocks. This geochemical bimodality appears to imply the crystallization of two or more unrelated layered series, derived from magmas intruded at two or more distinct phases of lunar evolution. The only way to reconcile this great geochemical diversity with the near-canonical magma-ocean model for earliest lunar evolution is to assume that one but not both of the two groups formed by plagioclase flotation over the magma ocean. If so, the obvious choice for the magma-ocean flotation product is the ferroan anorthositic suite, mainly because the Mg-suite rocks are generally not anorthositic enough to be plausible flotation cumulates.

A detailed assessment of the average mode of the relatively large (> 10 g) pristine ferroan rocks indicates that they contain 8.1 ± 8.6 wt% mafic silicates. Evaluation of mass balance to account for the relatively well determined bulk composition of the (upper) crust suggests an overall composition for the ferroan suite that is roughly one-sigma more mafic than the average mode. Models of crustal genesis by flotation of plagioclase over an appropriately FeO-enriched (and hence, extraordinarily dense) magma ocean, assuming that enough mafic silicates are rafted to equalize the bulk-crust density with the density of the underlying magma, similarly indicate that the overall ferroan suite, if it truly formed by this mechanism, should contain roughly 19 wt% mafic silicate. Thus, results from these three approaches to estimating the overall composition of the ferroan anorthositic suite appear consistent with the magma-ocean hypothesis, as modified to form the remainder of the crust (i.e., the Mg suite) within numerous, more localized layered intrusions, emplaced into the plagioclase-flotation crust (ferroan anorthositic suite) shortly after the magma ocean completed its crystallization. Acquisition of additional large lunar rocks will probably be necessary before we can learn with any confidence whether the overall ferroan suite is truly ferroan anorthosite, in the sense of having ≥90% plagioclase.

Introduction: Lunar samples and models of lunar-crustal genesis

The crust of the Earth's Moon, roughly 60 km thick (Taylor, 1982), comprises by far the largest known anorthosite complex in the solar system. The lunar surface area of 3.8×10^7 km² is equivalent to 25% of the entire land area of the Earth. Although ~17% of the lunar surface is covered by thin patches of a plagioclase-poor type of volcanic rock loosely described as basalt (mare basalt), nearly all of the rest (the lunar highlands) appears to be highly anorthositic, with Al_2O_3 concentration generally in the range 24–27 wt%. The Moon was once likened to an "albite-depleted giant Skaergaard" (Saxena et al., 1977), but after 20 years of lunar sample studies, we now know that the Moon is actually at least as complex as the Bushveld. Petrologists continue to debate the hypothesis (Wood et al., 1970) that some or all of the lunar crust was produced by plagioclase flotation over a global "ocean" of magma.

TABLE 1. Compositional data for lunar highlands regolith samples

	Y82/86032 Wtd. mean	ALHA81005 Wtd. mean	Y791197 Wtd. mean	Luna 20 Avg. soil	Apollo 16 Avg. soil	Apollo 17 Al-rich soil	Apollo 14 Avg. soil
Na_2O (wt%)	0.42	0.30	0.33	0.39	0.48	0.42	0.72
MgO (wt%)	5.3	8.2	6.1	9.3	6.0	9.7	9.3
Al_2O_3 (wt%)	27.7	25.6	26.1	23.1	27.2	21.3	17.7
SiO_2 (wt%)	44.6	45.7	43.9	45.4	44.9	45.2	48.1
CaO (wt%)	15.8	15.0	15.4	14.2	15.4	13.0	10.8
TiO_2 (wt%)	0.23	0.25	0.34	0.47	0.54	1.22	1.78
Cr_2O_3 (wt%)	0.109	0.129	0.130	0.185	0.106	0.216	0.205
MnO (wt%)	0.065	0.075	0.082	0.110	0.066	0.108	0.135
FeO (wt%)	4.6	5.4	6.4	7.2	5.0	8.0	10.3
Th (μg/g)	0.20	0.29	0.32	1.06	2.03	2.40	13.3
U (μg/g)	0.050	0.098	0.113	0.35	0.54	0.70	3.3
Molar Mg/(Mg + Fe)	0.673	0.728	0.629	0.697	0.678	0.684	0.617
FeO/MnO	70.6	72.3	78.5	65.9	76.0	73.9	75.8

Note: Samples are arranged in order of Th and U concentrations. Sources: For lunar meteorites, see Koeberl et al. (1989) or Warren et al. (1989). The Apollo 17 composition is a literature average for 73141, the most Al-rich soil from that site.

Lunar meteorites

Although estimates of the average composition of the lunar highlands surface have not changed greatly since the mid-1970s (Taylor, 1975), the *uncertainty* associated with these estimates has been considerably reduced by recent discoveries of lunar meteorites. As reviewed by Warren and Kallemeyn (1988), meteorites from at least two, and more likely three, separate lunar source craters have now been studied. Previous sampling of the Moon covered only a relatively minuscule area of the central nearside: a polyhedron can be drawn around the six Apollo and three Luna sites covering a mere 4.7% of the total lunar surface (Warren et al., 1989). The exact provenance of the lunar meteorites may never be known, but their extremely low concentrations of incompatible elements (e.g., Th, U) confirm the statistical probability that their sources are not coincidentally within the same 4.7% sampled by the Apollo and Luna programs. Orbital spectrometry results (e.g., Metzger et al., 1977) indicate that the small area sampled by Apollo and Luna is atypically rich in KREEP (i.e., REE, Th, U, etc.). The value of the lunar meteorites as constraints on the average composition of the crust is enhanced by the polymict nature of their rock types. Two of the three studied lunar meteorites (ALHA81005 and Y791197) are regolith breccias; the third (a suite of "paired" meteorites: Y82192, Y82193, and Y86032, hereafter designated by the most massive of the group, Y86032) is patently polymict and has at least some of the characteristics of an "immature" regolith breccia.

A regolith breccia is a rock produced by lithification (through meteoritic-impact shock) of a former surface mass of fine-grained, loose debris or "soil." Lunar soil is produced by extremely thorough mixing of the underlying crust. Except near mare-highland interfaces, the lunar regolith shows compositional variation barely detectable at the centimeter-meter scale and still relatively mild at the scale of the Apollo lunar traverses—e.g., Apollo 16, 8.3 km, Al_2O_3 range 23–30 wt%; Apollo 14, 1.6 km, Al_2O_3

range 17–19 wt%. Thus, as constraints on the gross composition of the Moon's crust, each separate lunar meteorite is tantamount to an extra Apollo or Luna mission, albeit to a random site. The compositions of the studied lunar meteorites are shown in Table 1. Aside from their relatively low concentrations of incompatible elements, and some scatter in their molar Mg/(Mg + Fe) ratios [all allusions to Mg/(Mg + Fe) in this paper refer to the molar ratio], these compositions closely resemble the average composition of soil from the one "true highlands" Apollo site (Apollo 16), as well as a typical mid-1970s estimation (Taylor, 1975) for the bulk composition of the lunar crust.

Pristine rocks

Polymict samples, and especially regolith samples, are ideal for evaluating the bulk composition of the lunar crust, but models for the genesis of the crust require constraints on the compositional and petrologic diversity of its unmixed or "pristine" lithologic components. The ancient nonmare, or highlands, portion of the lunar crust has been so thoroughly brecciated (and locally melted) by impacts that only a relatively tiny proportion of the rocks collected are compositionally unmixed, and fewer still are texturally pristine. In fact, the term "pristine" in lunar parlance generally refers only to a sample's composition. Roughly 100 pristine nonmare rocks (or rock clasts) large enough to permit geochemical analysis have been discovered. Pristine lunar rocks are recognizable mainly from their relict igneous (or in a few cases coarse-granulitic) textures, and their generally low concentrations of siderophile (i.e., meteoritic-debris-associated) trace elements (Warren and Wasson, 1977; Warner and Bickel, 1978).

The petrology of pristine nonmare rocks has been reviewed by, e.g., Warren and Wasson (1977, 1980a), Norman and Ryder (1979, 1980), James (1980), Warren (1985), and Shervais and Taylor (1986). Most nonmare rocks with relatively pristine textures appear to be cumulates, in many cases poikilitic (e.g., Fig. 1), although KREEP-composition pristine rocks tend to be ophitic basalts. As might be expected for a suite of unmixed ma-
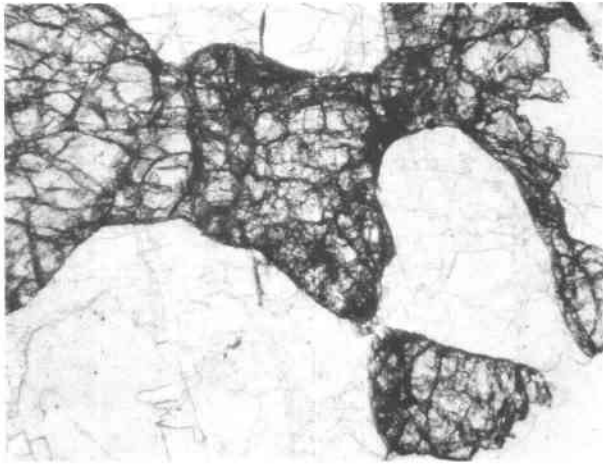
Fig. 1.   Photomicrograph of small, geochemically "ferroan" norite 66035c (thin section 66035,22); view is 2.1×1.6 mm, through partially crossed nicols. Note poikilitic texture, with anhedral pyroxene oikocryst surrounding plagioclase chadacrysts.

Fig. 2.   Bulk-rock Mg/(Mg + Fe) vs. Eu/Al for pristine nonmare rocks (× and + symbols) and highlands regoliths (▲). This diagram, even better than the more widely utilized Mg/(Mg + Fe) vs. Na/(Na + Ca) diagram, evinces the geochemical bimodality of the pristine rocks. Data from Table 1 (for regolith samples) and sources cited by Warren et al. (1989), most importantly the compilations of Ryder and Norman (1979) and Ryder (1985).

terials from a layered complex, these rocks exhibit great compositional diversity. They even manifest a striking geochemical bimodality, when plotted on variation diagrams for molar Mg/(Mg + Fe) ratio vs. a "plagiophile" ratio such as Na/Ca, Ga/Al, or Eu/Al (Fig. 2). One of the two main classes of pristine rocks has come to be known as ferroan anorthosite, because its members have distinctly lower Mg/(Mg + Fe) than all other pristine nonmare rocks with similarly "primitive" (low) Na/Ca, Ga/Al, Eu/Al, etc., and because most large samples with this "ferroan" pattern of geochemical ratios are modally anorthosite. The remaining pristine nonmare intrusive rocks are known by the broad classification "Mg-rich rocks" or "Mg-suite" rocks. Various subdivisions of the Mg suite have been proposed (e.g., James and Flohr, 1983; Lindstrom et al., 1989). James et al. (1989) have also proposed subdivisions of the relatively uniform ferroan anorthositic suite.

Models of lunar-crustal genesis

A common interpretation of the ferroan vs. Mg-rich bimodality among the pristine rocks is that the ferroan anorthositic rocks formed as the earliest substantial lunar crust, by flotation of plagioclase over a primordial Moonwide magma "ocean," and that shortly afterward this ferroan anorthositic crust was intruded by Mg-rich magmas, which formed the Mg suite within a number of compositionally diverse layered intrusions. Early objections to the magma-ocean model (e.g., Wetherill, 1981) centered around the difficulty for any heat source to engender simultaneous melting throughout a major fraction of the lunar volume. However, the recently popular giant impact model of lunar origin (Stevenson, 1987) predicts an extremely hot early Moon. The first age for a ferroan rock based on either Nd or Sr (in this case Nd) isotopes was reported for Apollo 16 anorthosite 60025 by Carlson and
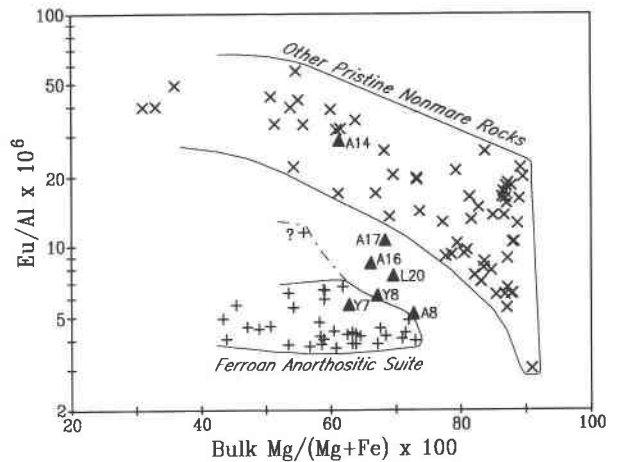
Lugmair (1988). The reported age, 4.44 ± 0.02 Ga, is the second oldest Nd-based age yet determined for a lunar rock. Several ostensibly older Sr-based ages have been reported for Mg-suite rocks (see compilation in Carlson and Lugmair, 1988). The only older Nd-based age, 4.51 ± 0.06 Ga, is a preliminary result for an Mg-suite anorthositic norite clast from breccia 15455 (C.-Y. Shih, personal communication, 1989).

A key constraint for this "magma ocean" (or more specifically, magma ocean plus later Mg-rich magmatism) scenario of lunar-crustal genesis is the consistently anorthositic nature of the geochemically ferroan rocks. The plagioclase-flotation aspect of the model obviously predicts a high average modal feldspar content for the ferroan suite, although so far remarkably little work had been done toward quantifying that prediction. Also, assuming that the ultimate parent magma of the ferroan suite was roughly basaltic (phase equilibria militate against production of anorthositic melt, albeit anorthositic crystal mushes might be derived from basalt by concentration of feldspar and residual melt apart from early mafic silicates), its crystallization should have yielded a comparable mass of cumulate that is "ferroan" geochemically, but relatively mafic. Apparently all of these mafic ferroan materials are buried at least several tens of kilometers deep (i.e., too deep to have been sampled by even the largest lunar craters). The scale of this separation process suggests that the genesis of the ferroan suite involved a deep, global-scale magma. A roughly analogous situation is found for the anorthositic suite of the Duluth Complex, which Weiblen and Miller (1988) have interpreted as a result of buoyant detachment of anorthositic crystal mushes apart from mafic cumulates, generated in deep-

crustal or subcrustal magma chambers associated with the dynamic environment of the Midcontinent rift.

Longhi and Ashwal (1985) have argued that the lunar ferroan suite might have formed by detachment and rise of anorthositic crystal mushes from a "stack" of many overlapping layered intrusions, produced by accretionary heating as the Moon grew. Longhi and Ashwal (1985) did not propose a detailed scenario for the origin of the Mg suite. They suggested that the Mg suite formed from high-degree partial melts that were intruded "beneath the layer of accumulated [ferroan] anorthositic diapirs." This model scarcely predicts the geochemical bimodality separating ferroan suite rocks from Mg-suite rocks (Fig. 2), however. The magma-ocean (plus later Mg-rich magmatism) model unambiguously predicts the strangely "ferroan" Mg/(Mg + Fe) ratios of the main variety of lunar anorthosite, because the fractional crystallization of an ultra-high-degree partial melt (the magma ocean) does not engender abundant plagioclase cumulate until the melt Mg/(Mg + Fe) ratio is greatly diminished by olivine and pyroxene fractionation. Thus, calculations indicate that for any given set of assumptions about the bulk-Moon contents of Al_2O_3, MgO, and FeO, the higher the degree of melting involved in the genesis of the magma ocean, the lower the Mg/(Mg + Fe) ratio that results for the magma-ocean plagioclase-flotation crust (Warren, 1986b). In contrast, the many separate "stacked" intrusions of the Longhi and Ashwal (1985) model would probably produce anorthosites with diverse Mg/(Mg + Fe) ratios, depending mainly on the Al_2O_3 concentration (and thus, the extent of before-plagioclase fractional crystallization) of the initial melt. According to the magma-ocean model, the Mg suite formed by subsequent crystallization of magmas intruded into (and thus prone to assimilate) a highly anorthositic crust. Consequently, plagioclase cumulates formed while the melt Mg/(Mg + Fe) was still relatively high, and the Mg-suite rocks seldom have low Mg/(Mg + Fe) unless they also have high Na/Ca, Eu/Al, etc. (Warren, 1986b).

Noting that our sampling of the pristine nonmare crust is still limited to a few tens of rocks from a mere handful of central nearside sites, Korotev and Haskin (1988) have argued that mass balance for the better-sampled polymict portion of the crust requires that ferroan anorthosite must be only a minor crustal component. Korotev and Haskin (1988) based this conclusion primarily on an inferred "excess" of Eu among polymict samples. The issue of bulk-crustal Eu is beyond the scope of this paper, but is discussed by Warren et al. (1990) [who have argued that the "excess" disappears if the results from lunar meteorites are factored into estimations of the mean composition of the crust; see also Warren and Kallemeyn (1988)]. However, Korotev and Haskin (1988) also suggested that modeling typical lunar polymict samples as mixtures of ferroan anorthosite with plausible combinations of pristine Mg-rich rocks cannot work because of mass-balance problems involving the elements Mg, Fe, and Al. Basically, they argued that the polymict samples have Mg/(Mg + Fe) ratios too ferroan to be derived by mixing Mg-

rich rocks with ferroan anorthosite, because the ferroan anorthosite (virtually pure plagioclase) contributes negligible Mg + Fe. Korotev (1983) had earlier employed similar reasoning to conclude that "ferroan anorthositic norite" (a rock type dominated by plagioclase and low-Ca pyroxene, with plagioclase between 60 and 77.5 vol%—Stöffler et al., 1980) may be an important "primary" lunar rock type, despite its virtual absence among large pristine rocks. For similar reasons, ferroan anorthositic norite has also been consistently advocated as an important "precursor" rock type by the Lindstroms (e.g., Lindstrom, 1983; Lindstrom and Lindstrom, 1986). A problem with the models of Korotev (1983) and Korotev and Haskin (1988), to which we shall return later, is that the composition assumed for ferroan anorthosite (35.5 wt% Al_2O_3) is unrealistically mafic poor.

The three aims of this paper are (1) to review the available evidence regarding the average modal feldspar contents of both the ferroan and the Mg-rich suites of pristine lunar rocks; (2) to compare the results with constraints from mass balance (as done by Korotev and Haskin, 1988) for polymict samples; and (3) to compare the results from both of these approaches with the average mode predicted for the ferroan suite by a model of flotation of plagioclase (along with "rafted" mafic silicates) atop a magma with Mg/(Mg + Fe) low enough to make a suitable parent for the ferroan suite.

Average plagioclase content of the ferroan suite, or, how high is high?

Remarkably, no one has ever before attempted to derive a statistical average for the modes of the two basic classes of pristine nonmare rocks. Perhaps the nearest precedent is the histogram of modes for pristine rocks >2 g in mass shown by Warren and Wasson (1980a). Of course, cumulates are notoriously heterogeneous in mode. Terrestrial cumulates often feature modal banding, such that monomineralic anorthosite is found a few centimeters apart from harzburgite or dunite. An estimate for the overall mode of the ferroan suite must allow in some fashion for the great disparity in mass among the samples for which modes have been published. Clearly, the mode of the tiny 5-mm² ferroan gabbro (35% plagioclase) clast found in a thin section of the ALHA81005 lunar meteorite (Warren et al., 1983a) should not be equated in significance with the mode of the 4.6-kg ferroan anorthosite 60015. In fact, a bias might be introduced into the database if all available modes are used, because modes for very small samples are probably seldom measured and published unless they appear extraordinary.

Table 2 shows a compilation of modal feldspar and whole-rock alumina contents for all pristine ferroan-suite rocks with well-determined masses of ≥ 10 g. All of these rocks are unambiguously ferroan in their geochemistry, and in most cases they are certainly pristine. Two of the tabulated rocks (15418 and 67215) have somewhat ambiguous textures (Nord et al., 1977; Ryder, 1985; Ryder and Norman, 1980; Taylor, 1982; Warren et al., 1983b;

TABLE 2. Simplified modes and Al$_2$O$_3$ concentrations for pristine lunar ferroan-suite rocks with masses >10 g

Sample	15415	60015	60025	60055	60135	60215	60515	61016	62236	
Mass (g)	269	4600	1836	35.5	120	300	16.7	200	57.3	
Plagioclase (vol%)	99.5	99	90	98	77	97	95	99.5	86	
Mafic phases (vol%)	0.5	1	10	2	23	3	5	0.5	14	
Mafic phases (wt%)	0.6	1.3	12.6	2.6	28.0	3.9	6.4	0.6	17.5	
Al$_2$O$_3$ (wt%)	35.5	35.5	31.9	34.0	32.7	35.5	35.3	34.6	30.1	
Al$_2$O$_3$ implied by mode	35.7	35.4	31.4	35.0	26.0	34.5	33.6	35.7	29.7	
References*	1	2, 3	4, 5	3	4, 6	2, 7	4, 6	2, 3	2, 3, 4, 8, 9	

Sample	62237	62255	62275	64435	65315	65325	67075	67915	15418†	67215†
Mass (g)	62.4	800	443	100	285	65	50	50	1141	276
Plagioclase (vol%)	85	99.5	93	97	98.5	98.5	95	85	70	77
Mafic phases (vol%)	15	0.5	7	3	1.5	1.5	5	15	30	23
Mafic phases (wt%)	18.7	0.6	8.9	3.9	1.9	1.9	6.4	18.7	35.8	28.0
Al$_2$O$_3$ (wt%)	31.2	35.3	34.2	35.5	34.9	35.1	32.1	28.9	26.4	25.9
Al$_2$O$_3$ implied by mode	29.3	35.7	32.7	34.5	35.2	35.2	33.6	29.3	23.3	26.0
References*	3, 10	2, 3	6, 11	2, 12	3	2	2	7, 13	1	14, 15, 16

	Surely pristine rocks only				Surely pristine rocks + 15418				Including 15418 and 67215			
	Mean	Std. dev.	Mw mean	Srw mean	Mean	Std. dev.	Mw mean	Srw mean	Mean	Std. dev.	Mw mean	Srw mean
Plagioclase (vol%)	93.7	6.7	96.3	95.1	92.4	8.6	93.5	92.5	91.6	9.1	93.0	91.7
Mafic phases (vol%)	6.3	6.7	3.7	4.9	7.6	8.6	6.5	7.5	8.5	9.1	7.0	8.3
Mafic phases (wt%)	8.1	8.6	4.7	6.3	9.7	10.9	8.3	9.5	10.7	11.5	8.9	10.5
Al$_2$O$_3$ (wt%)	33.7	2.1	34.5	34.1	33.3	2.7	33.6	34.1	32.9	3.1	33.4	33.0
Al$_2$O$_3$ implied by mode	33.0	2.9	34.2	33.7	32.5	3.6	32.9	32.5	32.1	3.8	32.8	32.2

Note: Abbreviations: Std. dev. = standard deviation, Mw = mass-weighted, Srw = square-root mass-weighted (see text).

* References: 1, Ryder (1985); 2, Ryder and Norman (1979); 3, Ryder and Norman (1980); 4, Butler et al. (1972); 5, Warren et al. (1987); 6, Warren et al. (1983c); 7, Rose et al. (1975); 8, Nord and Wandless (1983); 9, Warren and Wasson (1978); 10, Haskin et al. (1981); 11, Prinz et al. (1973); 12, James et al. (1989); 13, Marti et al. (1983); 14, McGee (1988); 15, Lindstrom and Lindstrom (1986); 16, Warren et al. (1990).

† Samples 15418 and especially 67215 are dubiously pristine.

McGee, 1988), but appear to be contaminated with meteoritic siderophile elements (Ganapathy et al., 1973; Lindstrom and Lindstrom, 1986; Warren et al., 1990). McGee (1988) has inferred that 67215 is polymict, albeit dominated to an unusual degree by a single parent rock type, ferroan noritic anorthosite. It seems likely that 15418 is of similar derivation. Table 2 shows only a total for all of the mafic phases in each sample. The modes of ferroan-suite rocks include very little of anything other than plagioclase, pyroxene (mostly "inverted" pigeonite), and olivine. However, the "mafic phases" category of Table 2 includes traces of Cr- and Fe-bearing spinel, ilmenite, FeNi metal, and troilite. The pyroxene/(pyroxene + olivine) ratio varies widely, from 0 in 67915c and ~0.06 in 62237, to 1 in many instances [in general, the more nearly monomineralic anorthosites have the highest pyroxene/(pyroxene + olivine) ratios, but a few relatively mafic samples such as 60135 also lack olivine].

Table 2 shows a mean and standard deviation for the plagioclase, total mafic phases, and Al$_2$O$_3$ contents of the ferroan suite, with and without the dubiously pristine 15418 and 67215. Mass-weighted means are also shown. However, much of the diversity among the samples probably reflects real meter- and even kilometer-scale heterogeneity in the crust, and a single 4.6-kg sample (60015) accounts for 50% of the total mass of the suite, so the simple mean is probably a better gage of the true crustal average. Means based on weighting in proportion to the square root of the mass are also shown. Modes are mea-

sured in volume percent, but Table 2 also shows modal content of mafic phases, translated into weight percent. For purposes of this calculation, the mafic fraction was modeled as a mixture of equal proportions of low-Ca pyroxene and olivine, with Mg/(Mg + Fe) ratios of 0.66 and 0.63, respectively, implying a mean density of 3.56 g·cm^{-3} (Deer et al., 1965). At these Mg/(Mg + Fe) ratios, the densities of low-Ca pyroxene and olivine are similar, so the pyroxene/(pyroxene + olivine) ratio has little effect on the density of the mafic fraction. The average composition of the plagioclase was assumed to be An$_{96}$, resulting in an average density of 2.74 g·cm^{-3} (Campbell et al., 1978). The weight fraction of mafic phases, w_m, was thus calculated as

$$w_m = 3.56v_m/(3.56v_m + 2.74v_p), \qquad (1)$$

where v_m is the volume fraction of mafic phases, and v_p is the volume fraction of plagioclase (= $1 - v_m$). This method probably leads to a slight underestimate of the "true" w_m, because traces of the material assigned to the mafic category in the simplified modes of Table 2 are actually chromite, ilmenite, FeNi metal, and troilite, all of which are denser than the mafic silicates.

For purposes of comparison between modes and analyses, Table 2 also shows estimates for the bulk-rock concentration of Al$_2$O$_3$ implied by the mode. These were calculated by simply assuming that all plagioclase has an Al$_2$O$_3$ content of 35.9 wt% and all mafic material has an average Al$_2$O$_3$ content of 0.57 wt%. In some cases, these

TABLE 3. Simplified modes and Al_2O_3, MgO, and FeO concentrations for pristine Mg-suite and KREEP rocks with masses >5 g

Sample	1538X	15445c	15455c	67667	72255	72275c	7241X	76255c	76335	76535	76536	77035c	77215+	782XX
Mass (g)	10.7	10	200	7.9	10	5	58.7	300	360	155	10.3	100	840	395
Plagioclase (vol %)	42	62.5	72	22	40	40	4	50.6	77	58	60	60	55	46.8
Al_2O_3 (wt%)	15.7	22.5	27.0	7.6	15.5	14.4	1.3	15.8	29.4	20.7	26.3	19.1	15.1	20.8
MgO (wt%)	8.6	9.9	6.9	26.0	16.0	8.4	45.0	11.0	9.0	19.0	14.0	12.0	13.0	11.0
FeO (wt%)	9.7	3.9	2.9	17.0	7.4	14.0	12.0	9.7	2.3	5.0	3.6	5.8	9.9	5.0
References*	1	2	2, 3, 4	5, 6	7	1	1, 7	8, 9	10	7	5, 11	5	1, 12	5, 7, 11, 13

	Intrusive rocks only				Intrusive rocks + 1538X				Intrusive rocks + 1538X + 72275c			
	Mean	Std. dev.	Mw mean	Srw mean	Mean	Std. dev.	Mw mean	Srw mean	Mean	Std. dev.	Mw mean	Srw mean
Plagioclase (vol%)	50.7	20.6	56.8	55.1	50.0	19.9	56.7	52.6	49.3	19.3	56.7	54.6
Al_2O_3 (wt%)	18.4	8.1	19.4	19.4	18.2	7.8	19.4	18.8	17.9	7.6	19.4	19.2
MgO (wt%)	16.0	10.4	12.4	13.8	15.5	10.2	12.4	13.6	15.0	10.0	12.4	13.6
FeO (wt%)	7.0	4.4	6.9	6.7	7.2	4.2	6.9	6.7	7.7	4.5	6.9	6.8

Note: Abbreviations: 1538X = combination of 15382 and 15386; 7241X = combination of 72415, 72416, 72417, and 72418; 77215+ = combination of 77075, 77076, 77077, and 77215; 782XX = combination of 78235, 78236, 78238, and 78255; Std. dev. = standard deviation, Mw = mass-weighted, Srw = square-root mass-weighted (see text).

* References: 1, Warren (1990); 2, Ryder (1985); 3, Warren and Wasson (1980b); 4, Butler et al. (1971); 5, Warren and Wasson (1979b); 6, Ryder and Norman (1980); 7, Ryder and Norman (1979); 8, Phinney (1981); 9, Warner et al. (1976); 10, Warren and Wasson (1978) and unpublished UCLA data; 11, Butler et al. (1973); 12, Chao et al. (1976); 13, Warren et al. (1986).

results show only rough agreement with the Al_2O_3 concentrations of the actual analyses of the rocks. However, in the case where agreement is poorest (60135), the chemical analysis is based on a relatively tiny sample (645 mg) of an exceptionally coarse-grained (4–5 mm) rock (Warren et al., 1983), whereas the mode is based (in part) on binocular-microscopic observations of a surface exposure of many square centimeters (Butler et al., 1972).

Two important conclusions can be drawn from Table 2. First, although the ferroan suite is frequently regarded as a collection of rocks that are virtually monomineralic plagioclase (e.g., Korotev and Haskin, 1988), in fact their statistical mean plagioclase content is 92 wt% (or 90 wt% if 15418 is included). Second, even if we ignore the large uncertainties still associated with some individual modes (e.g., 60025, 60135), the uncertainty in the statistical mean of the modes is very large. In terms of weight percent mafic phases, the standard deviation is slightly greater (8.6 wt%) than the mean (8.1 wt%). It should be noted, however, that the standard deviation for a "closed" data set, such as a set of modes (which are constrained to sum to 100%), is inherently more ambiguous than the standard deviation for a perfectly random "open" data set (Butler, 1979).

MIXING-DECONVOLUTION CONSTRAINTS ON THE AVERAGE FERROAN-SUITE COMPOSITION

As was adduced (somewhat indirectly) by Korotev and Haskin (1988), the average plagioclase/mafic ratio of the ferroan suite can also be usefully constrained by examining mass-balance models for Mg, Fe, and Al in the average lunar crust. The average composition of the upper crust can be inferred with relative precision by averaging the compositions of the diverse highlands regolith samples, including the lunar meteorites (Table 1). More difficult to determine, however, is the average composition of the material that was mixed with the ferroan suite to form the total nonmare upper crust. Simplified modes and concentrations of Al_2O_3, MgO, and FeO for relatively large Mg-suite and KREEP rocks (i.e., all nonmare, nonferroan rocks ≥5 g in mass) are shown in Table 3. Samples designated with "c" appended to their NASA numbers (e.g., 15445c) were found as clasts within much larger polymict breccias. Statistics analogous to those in Table 2 are also shown. Separate statistics are shown for the data set stripped of one or both of the extrusive (KREEP) rock types, but for purposes of modeling bulk-crustal mass balance, the full data set is presumably the most appropriate.

For Al_2O_3, the standard deviation of the mean in Table 3 is large, but fortunately an additional constraint is available besides the direct rock analyses. Even among smaller pristine rocks, dunites are relatively scarce compared to troctolitic rocks. Thus, it appears that the parental magmas of the overall assemblage were generally at least nearly saturated with plagioclase even as they began crystallization. Phase-equilibria constraints (e.g., Fig. 4 of Longhi and Pan, 1988) imply that moderately magnesian magmas prone to crystallize minor dunite before commencing to form troctolites (and later norites) begin with roughly 18 wt% Al_2O_3—which agrees with the imprecise mean of the measured rocks (Table 3). Note that the mean modal plagioclase (49 vol%, or roughly 44 wt%) also implies roughly 16–18 wt% Al_2O_3.

For FeO and especially MgO, the standard deviations (Table 3) are extremely large. Therefore, it seems best to constrain these two elements by an indirect method, as follows. The SiO_2 concentration of the nonferroan mélange is known to within 1–2 wt%, simply because the crust is extremely homogeneous for SiO_2 (Table 1). The
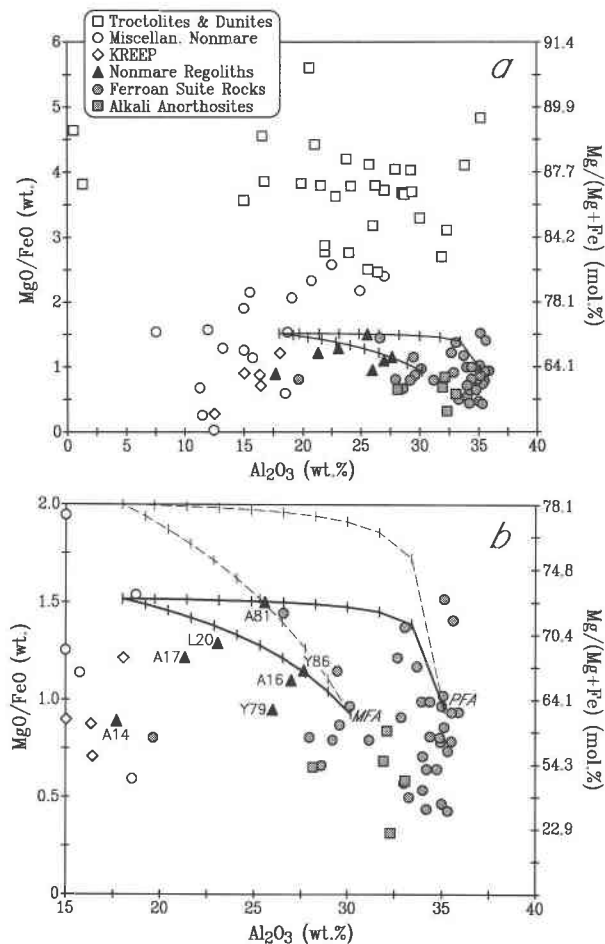
Fig. 3. (a) Bulk-rock Al_2O_3 vs. MgO/FeO for pristine nonmare rocks and highlands regoliths (▲). Also shown are results from mixing models involving an estimate for the bulk nonferroan portion of the crust ($Al_2O_3 = 18$ wt%) mixing with two different estimations for the bulk ferroan portion: PFA = "pure" ferroan anorthosite, MFA = "mafic" ferroan anorthosite (see Table 4). Note that among nonferroan pristine rocks (other than the volumetrically minor alkali anorthosites), the MgO/FeO ratio tends to be positively correlated with Al_2O_3 and thus inversely correlated with the total content of MgO + FeO. The vertical tick marks along the mixing lines represent 10 wt% increments in the mixing proportions. Regarding data sources, see Fig. 2. (b) Expanded view of the lower right portion of (a); with an additional set of mixing models shown (dashed curves), based on a MgO/FeO ratio of 2.0 for the assumed composition of the nonferroan portion of the nonmare crust (see text).

CaO/Al_2O_3 ratio also appears relatively homogeneous throughout the sampled crust. The average CaO/Al_2O_3 ratio from Table 1 (0.59) is only marginally higher than the typical ratio for ferroan anorthosites, ~0.55 (Ryder and Norman, 1979). Hence, if Al_2O_3 is assumed to be 18 wt%, then CaO must be 10–11 wt%. Thus, the sum for SiO_2 + CaO + Al_2O_3 must be close to 74 wt%. Other components such as Na_2O and TiO_2 combine to contribute about 1 wt%, which leaves 25 wt% for the total of MgO + FeO. Finally, we need to infer the MgO/FeO ratio. The MgO/FeO ratio implied by the average compositions of large pristine Mg-suite and KREEP rocks (Table 3) is roughly 2.0, but this result carries an enormous uncertainty. However, we can establish a relatively secure upper limit by noting that very few lunar olivines are more magnesian than Fo_{90}, and virtually none are more magnesian than Fo_{92} (Warren, 1986a, cf. Fig. 2). At lunar-crustal pressure, the K_D for olivine-melt Fe/Mg exchange is ~0.30 (Warren, 1986a), and thus magmas that failed to form olivines more magnesian than Fo_{90} probably never had melt Mg/(Mg + Fe) higher than ~0.73 [i.e., MgO/FeO higher than 1.52]; and the lack of $Fo_{>92}$ olivines implies that melt Mg/(Mg + Fe) was virtually never higher than 0.78 (i.e., MgO/FeO higher than 1.99). Of course, if the Mg suite truly formed in many different layered intrusions, then presumably not all of their parent magmas were as magnesian as the most extreme ones. The KREEP basalts (our only clearly extrusive pristine nonmare rocks) obviously formed from magmas with Mg/(Mg + Fe) ≪0.73. Thus, the average MgO/FeO for the nonferroan portion of the nonmare crust is implied to be ≪2.0, and most likely slightly <1.5. On balance (considering also the more direct evidence from the compositions of the large rocks; Table 3), 1.5 appears to be a good estimate for the average MgO/FeO ratio of the nonferroan portion of the nonmare crust. Thus, the concentrations of MgO and FeO are inferred to be 15 and 10 wt%, respectively. The inferred bulk composition of the nonferroan crust is shown in Table 4.

Figure 3 shows Al_2O_3 vs. MgO/FeO for pristine nonmare rocks of all categories, along with results from mixing (mass-balance) models, based on two different assumptions regarding the overall composition of the ferroan suite. The PFA composition (Table 4) is almost (not quite) as devoid of mafic phases as the FA composition (35.5 wt% Al_2O_3) discussed by Korotev and Haskin (1988). The MFA composition is almost precisely one-sigma more mafic than the mean ferroan-suite composition based on Table 2. Figure 3b includes a pair of mixing models (dashed curves) with the MgO/FeO ratio of the nonferroan crust assumed to be 2.0, instead of the more likely value of 1.5. Figure 3 shows that the fit to the highlands regoliths is clearly more generally satisfactory using the MFA end-member than using the PFA end-member, and the proportion of the ferroan-suite material is generally in the range 45–75 wt%. Even the MFA-based model results in a poor fit for Apollo 14 regolith (and both MFA-based and PFA-based models are inadequate for all but the ALHA81005 regolith if the MgO/FeO ratio of the nonferroan crust assumed to be 2.0). However, the extraordinarily U-rich Apollo 14 composition is clearly dominated by KREEP and thus is not expected to contain a large proportion of ferroan-suite material. [It is also worth noting that in addition to the components mod-

TABLE 4. Compositions of end-members used for mixing models and (for comparison with PFA and MFA) compositions predicted for primordial crust by four models of plagioclase flotation over a ferroan magma ocean

	Na₂O (wt%)	MgO (wt%)	Al₂O₃ (wt%)	SiO₂ (wt%)	K₂O (wt%)	CaO (wt%)	TiO₂ (wt%)	Cr₂O₃ (wt%)	MnO (wt%)	FeO (wt%)	Mg/ (Mg + Fe) (mol%)
Est. avg. nonferroan crust*		15	18	45–46		10–11				10	72.8
PFA ("pure" FA)**	0.38	0.31	35.1	44.4	0.014	19.1	0.017	0.004	0.008	0.33	63.0
MFA ("mafic" FA)†	0.27	4.15	30.2	43.7	0.010	17.4	0.054	0.075	0.068	4.48	62.3
Model flotation crusts‡											
A	0.37§	3.84	29.4	44.9	0.016§	17.1	0.068	0.059	0.048	3.99	63.2
B	0.32§	3.87	29.1	45.1	0.014§	16.9	0.063	0.037	0.061	4.19	62.2
C	0.30§	4.10	28.3	44.4	0.013§	16.1	0.051	0.046	0.099	6.49	52.9
D	0.40§	4.23	29.0	44.6	0.018§	16.6	0.065	0.134	0.073	4.59	62.2

* Regarding estimated average nonferroan crust, see text.
** "Pure" ferroan anorthosite: average of 15415, 60015, 60025 "pure" portions, and 65315 (see Table 2).
† "Mafic" ferroan anorthosite: average of 60025,696 (a "mafic" portion), 60135, 62236, and 62237 (see Table 2).
‡ Model flotation crusts are derived from Warren's (1989) model of magma-ocean crystal fractionation, with various initial compositions—(A) mixture of ⅔ peridotite KLB-1 (Takahashi, 1986) and ⅓ Warren's (1986a) "SI," (B) pure KLB-1, (C) bulk-Moon composition proposed by Dreibus et al. (1977), and (D) bulk-Moon composition proposed by Taylor (1982)—and arbitrarily chosen initial depths of (A) 500, (B) 1200, (C) 1500, and (D) 1500 km.
§ In model flotation-crust compositions, initial magma concentrations of Na₂O and K₂O are essentially free parameters (Na and K are depleted in the whole Moon as volatile elements). The Na₂O and K₂O concentrations for KLB-1 have been arbitrarily adjusted to 0.05 and 0.005 wt%, respectively.

eled, highlands regoliths contain traces of mare-basaltic and meteoritic debris, both of which tend to slightly diminish the overall Mg/(Mg + Fe) ratio and Al₂O₃ content. In Table 1 and Figure 3, all Fe is treated as FeO, as is customary among lunar petrologists, but actually a small proportion of the Fe in any lunar soil is FeNi metal, mainly of meteoritic derivation. Among the regoliths included in Figure 3, the proportion of mare material is probably highest for Apollo 14 and 17.] Thus, mass-balance considerations hardly imply that ferroan anorthosite must be a minor component of the lunar crust (Korotev and Haskin, 1988), but they do tend to suggest that the overall composition of the ferroan suite is roughly MFA-like, i.e., about one-sigma more mafic than the composition implied by simple averaging of massive ferroan-suite samples.

The MFA composition implies a modal plagioclase content of ~87 vol%. The name recommended by the Lunar and Planetary Sample Team (LAPST) Nomenclature Committee (Stöffler et al., 1980) for such a rock type is noritic anorthosite (or troctolitic anorthosite, if the pyroxene/olivine ratio is less than 1). In IUGS nomenclature, it is a leucogabbronorite (the only IUGS rock type with anorthosite as part of its name has ≥90% plagioclase). Considering the ambiguity of the present lunar database regarding the average composition of the ferroan suite, the customary tendency to refer to the entire suite as "ferroan anorthosite" should probably be eschewed, or at least the term should be used advisedly. A less prejudicial, but similarly descriptive, term might be "ferroan anorthositic suite" (which might as an option be abbreviated as FAS, or even converted into a new rock name with the traditional "ite" ending, e.g., ferranite), defined on the basis of a high, but not necessarily extreme, plagioclase content, coupled with low Mg/(Mg + Fe) and Na/Ca ratios.

PLAGIOCLASE-FLOTATION MODELS OF FERROAN-SUITE ORIGIN

Many models have been developed to simulate the cooling and fractional crystallization of a primordial lunar magma ocean (e.g., Longhi, 1977; Warren and Wasson, 1979a; Shirley, 1983). Although the floated-plagioclase crust would obviously have "rafted" a small proportion of mafic material, surprisingly little work has been done toward quantitatively constraining this effect. The proportion of rafted mafic matter must be closely related to the density, and thus indirectly to the composition, of the parent melt. The density of a silicate melt of any given composition can be inferred, to within ~1% relative, by calculation of the mean molar volumes of its oxide constituents (Bottinga and Weill, 1970; Mo et al., 1982). Calculations using the method of Bottinga and Weill (1970) have previously been shown to predict flotation of plagioclase if the parent melt is relatively Fe-rich (Warren and Wasson, 1979a), but not if the parent melt is relatively magnesian (Roeder et al., 1977). In this section, the implications of the close relationship between melt composition and density will be more fully explored, by (1) evaluating petrologic constraints on the composition of the magma ocean as it putatively crystallized the FAS and (2) translating the densities implied by those compositional constraints into predictions for the plagioclase to mafic silicate ratio of the FAS.

Some models (e.g., Delano and Ringwood, 1978) have assumed that the mafic component of the lunar crust originated as the intercumulus "trapped melt" fraction of the primordial flotation crust (with the proportion of trapped melt chosen to fit the bulk composition of the crust, assuming all of the cumulus matter is plagioclase and the parent melt was basaltic). Such models are grossly inconsistent with the complex, bimodal geochemistry of the

pristine rocks (Fig. 2). Moreover, incompatible-element systematics and textural evidence (Warren and Wasson, 1980a; Ryder, 1982; James et al., 1989) imply that most of the mafic material of the ferroan-anorthositic suite originated not as trapped melt, but rather as cumulus olivine and pyroxene crystals and compositionally equivalent adcumulus (and poikilitic "heteradcumulus") matter. (For an alternative interpretation, see Haskin et al., 1981.) For modeling purposes, I will assume that none of the mafic matter originated as trapped melt. The final proportion of mafic silicates calculated in this manner might be a slight underestimate, because the density of the melt is much lower than the density of the mafic silicates. The underestimation is probably not large, however, because even the mafic silicates produced by the trapped melt must have crystallized contemporaneously with, if not faster than, the underlying magma ocean.

The magma-ocean crystal-fractionation model used here has been described in detail by Warren (1989). It is the latest in a series than has been continually refined since first developed by Warren and Wasson (1979a). The entire family of models originated by adaptation of, and continues to resemble, a series of models by Longhi (e.g., 1977, 1980). Briefly described, these models use a finite-difference technique to simulate fractional crystallization of olivine, pyroxene, plagioclase, etc.; in sequences, proportions and compositions constrained by mathematical representations of phase diagrams (most notably for the system olivine-plagioclase-silica) and crystal/liquid distribution coefficients, based on pertinent results from the experimental-petrology literature. For lunar purposes, the most important difference between the present model and its forebears lies in its extended modeling of density effects during the crustal-genesis process.

In the model, the density of the nascent crust is constrained to match that of the underlying magma ocean, or more specifically, the melt fraction of the magma ocean (presumably, at least for most of its life, a finite proportion of the "ocean" comprised loose, yet-to-accumulate crystals). Arguably, the effective stability of the crust may have been enhanced by the strengthening effect of its domical shape (J. T. Wasson, personal communication, 1978), and its buoyancy may have been enhanced by porosity produced by brecciation in the upper few kilometers. Again, these effects would lead to an underestimation for the final proportion of mafic silicates; but probably not a large underestimation. Cratering would have continually undermined the crust's strength, and heat-sintering would restrict the zone of substantial porosity to within a kilometer or two of the surface. Factors unrelated to buoyancy might have tended to enhance the plagioclase to mafic silicate ratio of the floated crust. For example, Morse (1982) noted that the stability of plagioclase relative to mafic silicates increases with decreasing pressure. (However, in the H_2O-poor lunar environment, pressure also strongly affects the relative stabilities of pyroxene and olivine, and yet pyroxene and olivine occur in roughly equal proportions among relatively mafic ferroan-suite

rocks.) On balance, it seems best to simply assume that the density of the flotation crust roughly matched that of the underlying magma ocean.

Given the assumption that ρ_{melt} and ρ_{crust} (the average densities of the melt and crust) are equivalent, one can write

$$\rho_{melt} = \rho_{plag} v_{plag} + \rho_{maf}(1 - v_{plag}), \qquad (2)$$

where ρ_{plag} is the density of the plagioclase, v_{plag} is the volume fraction of plagioclase in the FAS crust, and ρ_{maf} is the average density of the mafic component of the FAS crust. Solving for v_{plag} yields

$$v_{plag} = (\rho_{maf} - \rho_{melt})/(\rho_{maf} - \rho_{plag}). \qquad (3)$$

Of the quantities on the right side of Equation 3, the best known is ρ_{plag}. Ferroan-suite plagioclase is uniformly $An_{96\pm1}$, implying a near-solidus (roughly 1000 °C) density of 2.70 ± 0.02 g/cm^3 (Campbell et al., 1978). The value of ρ_{maf} can likewise be constrained by noting that the mafic silicates of the FAS are also relatively uniform in their $Mg/(Mg + Fe)$ ratio and thus in their densities. As discussed above, the mafic component of the FAS comprises roughly equal proportions of low-Ca pyroxene and olivine (with minor high-Ca pyroxene), and the $Mg/(Mg + Fe)$ ratio of the low-Ca pyroxene averages ~0.66, whereas the olivine averages ~Fo_{63}, implying a mean near-solidus density of ~3.49 ($\pm$ ~0.10) g cm^{-3} (Deer et al., 1965). Constraining the remaining unknown, ρ_{melt}, will be more difficult. However, ρ_{melt} is also constrained (indirectly) by the conveniently uniform $Mg/(Mg + Fe)$ ratios of the FAS mafic silicates. For both olivine/melt and low-Ca pyroxene/melt equilibria, and K_D for Fe/Mg exchange (at low pressure) is constrained by a large number of experimental studies to be ~0.30 ($\pm$ ~0.02) (Warren, 1986a). Consequently, the observation (Fig. 3b) that FAS rocks with significant mafic components (i.e., having $Al_2O_3 \ll$ 35 wt%) consistently have MgO/FeO between 0.6 and 1.4 [$Mg/(Mg + Fe)$ between 0.52 and 0.72] implies that their parent melt(s) had $Mg/(Mg + Fe)$ confined to the range 0.25–0.44.

Among the results of the magma-ocean crystal-fractionation model of Warren (1989), regardless of initial melt-zone composition (within reason) and depth, the melt FeO content shows a steady increase, and its $Mg/(Mg + Fe)$ ratio shows a steady decrease, throughout the fractionation sequence (except for the final few percent of fractionation, where ilmenite begins to fractionate). This result is by no means unique to Warren's (1989) model (cf. Binder, 1976; Drake, 1976; Longhi, 1977, 1980; Gromet et al., 1981) and is a natural consequence of the uniformly low K_D for Mg/Fe exchange involving (Mg,Fe)-rich primitive lunar minerals (i.e., olivine and few varieties of pyroxene). Compared to the other constituents of silicate melts, FeO has an extremely high density; e.g., at 1200 °C, the ratio M_i/V_i (gram formula weight over molar volume) is 5.35 for FeO, 3.70 for Fe_2O_3, 3.57 for CaO, 3.56 for MgO, 2.22 for SiO_2, and 2.82 for Al_2O_3 (Mo et al., 1982). Consequently, melt density in the anhydrous,
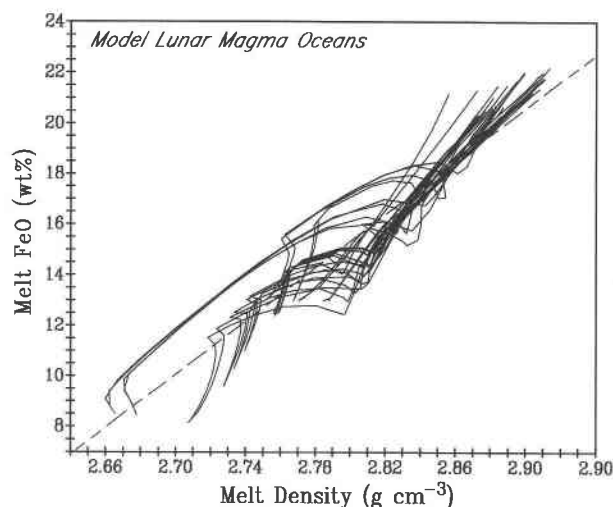
Fig. 4. Melt density vs. FeO concentration for magma oceans of various initial compositions and depths (permutations of the values listed in Table 4, plus a few models starting with the "H-IAB-RKUM 1:1:1 hybrid" composition of Warren and Wasson, 1979a), evolving according to the crystal-fractionation model of Warren (1989). Melt density is calculated with the use of Bottinga and Weill's (1970) model and updated partial molar volumes from Mo et al. (1982).
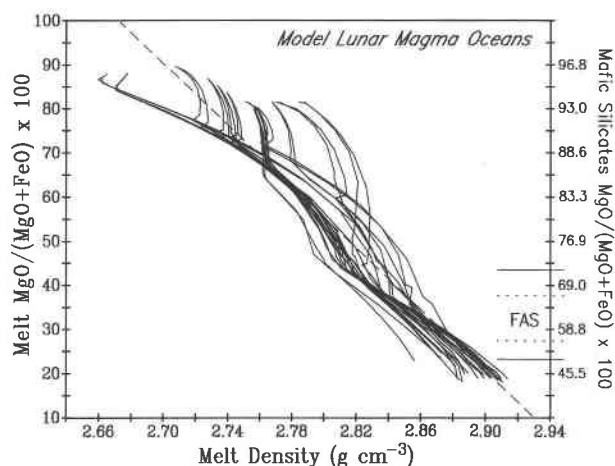
Fig. 5. Melt density vs. Mg/(Mg + Fe) ratio for magma oceans of various initial compositions and depths, evolving according to the crystal-fractionation model of Warren (1989); analogous to Fig. 4. Mafic silicate MgO/(MgO + FeO) ratios shown are calculated assuming $K_D = 0.30$ (see text).

Fe_2O_3-poor Moon is strongly correlated with melt FeO concentration. Thus, when during magma-ocean evolution the melt becomes progressively enriched in FeO, its density (i.e., ρ_{melt}) shows a commensurate increase (Fig. 4), and ρ_{melt} shows an equally strong inverse correlation with melt Mg/(Mg + Fe) (Fig. 5). From these results, it appears that for any given melt Mg/(Mg + Fe) in the vicinity of 0.30, the value of ρ_{melt} is constrained to within 1–2% relative.

Even including near-monomineralic anorthosites, the range in average mafic silicate Mg/(Mg + Fe) ratio among FAS rocks extends only from 0.50 (67455c, Minkin et al., 1977) to 0.72 [64435c, James et al., 1989; note: for the bulk-clast Mg/(Mg + Fe) ratio of 64435c, the INAA data of James et al. indicate 0.74, but their microprobe analyses of its constituent minerals indicate precisely 0.72; the value used for Figs. 2 and 3 is 0.72]. Applying the $K_D \approx 0.30$ constraint, the corresponding range for the Mg/(Mg + Fe) of the parent melt is 0.23–0.44 (Fig. 5). The average Mg/(Mg + Fe) of the FAS is almost certainly in the range 0.56–0.67 (dotted lines in Fig. 5), implying a parent melt Mg/(Mg + Fe) in the range 0.28–0.38. Thus, the magma-ocean modeling results (Fig. 5) imply that the density ρ_{melt} of the FAS parent melt was never far from 2.85 g cm^{-3}. Adding this result to the previously derived values for ρ_{maf} and ρ_{plag}, Equation 3 implies that v_{plag} should be ~0.81 (or, expressed as weight fraction of plagioclase in the FAS crust, ~0.77).

The application of Equation 3 has thus far been discussed in a generic fashion, taking only an average (based on all models tested) for the unknowns ρ_{maf} and, more important, for ρ_{melt}. However, the model of Warren (1989) applies Equation 3 to each individual magma-ocean scenario tested and calculates an average composition for the floated crust as a function of the crystals floated over the entire course of crystallization. The individual models are thus analogous to the generic model just described, except more precise. The models assume that the average composition of the mafic fraction floated (during any given interval) is the same as the average composition of all mafic crystals contemporaneously fractionated. (Although olivine is slightly denser than pyroxene of equivalent Mg/(Mg + Fe), this density effect may be roughly offset by the tendency for pressure to foster a higher olivine/pyroxene ratio among the crystals grown at the top of the magma than among those grown at its base.) Results for the "final" (crystallization 99 wt% complete) composition of the flotation crust, based on a range of different magma-ocean scenarios, are compared with the PFA and MFA compositions in Table 4. The results shown are merely representative of results that have been obtained from models starting with a great diversity of magma-ocean initial composition and depth combinations—models too numerous to describe here. However, all models that generate a flotation crust with an appropriately "ferroan" Mg/(Mg + Fe) ratio lead to an Al_2O_3 content for that crust within a few weight percent of 29. Expressed in terms of the volume fraction of plagioclase predicted for the FAS crust (v_{plag}), this result is ~85 vol% or, in terms of the weight fraction of plagioclase, ~81 wt%.

CONCLUSIONS

1. The average mafic content of the ferroan anorthositic suite (FAS) of pristine lunar rocks, as directly constrained by statistics for modes of relatively large sam-

ples, is 8.1 ± 8.6 wt%. This mean is only marginally consistent with the common tendency to refer to the overall suite as "ferroan anorthosite." The large standard deviation for the FAS implies that the true overall composition of the suite could easily be too mafic to justify the term "anorthosite," at least in its narrower sense (implying plagioclase ≥90 vol%). However, the mean mafic content for large pristine rocks with nonferroan geochemical affinities (based on Fig. 2 and similar diagrams) is much farther from anorthosite, and yet the overall composition of the crust (or at least the upper portion of the crust, which contributes material to highlands regoliths) is now well constrained, thanks to recent discoveries of lunar meteorites, to be relatively anorthositic, with ~24–27 wt% Al_2O_3.

2. An assessment of bulk-crustal mass balance for Al_2O_3, MgO, and FeO suggests (a) that the overall composition of the FAS is about one-sigma more mafic than the average mafic content of the large pristine FAS rocks (i.e., the overall mafic content of the FAS is roughly 17 wt%) and (b) that the proportion of FAS rocks in the crust is generally in the range 45–75 wt%; i.e., 60 ± 15 wt%. It should be noted here that according to Warren's (1986b) model for the subsequent origin of the Mg suite, roughly one fourth of the material now found as Mg-rich rocks originated by assimilation of FAS rock into the Mg-rich parental magmas; hence, the proportion of the (upper) crust ultimately derived from the magma-ocean plagioclase-flotation process is implied to be 70 ± 15 wt%. A high proportion of FAS material in the overall (upper) crust has also been inferred on the basis of extrapolation from Apollo orbital spectrometery results (Spudis and Davis, 1986).

3. Models of genesis of the FAS by flotation of plagioclase plus rafted mafic silicates atop a magma ocean with an appropriately "ferroan" Mg/(Mg + Fe) ratio also suggest that the overall composition of the suite should be about one-sigma more mafic than the average mafic content of the large pristine FAS rocks. This "high" mafic content is predicted for the flotation crust because the parental magma is inferred to have been uncommonly FeO-rich, and consequently, far more dense than pure plagioclase.

4. The same density-modeling approach tends to confirm earlier suggestions that the combined FAS and Mg-suite portion of the crust could not have formed entirely by plagioclase flotation over a magma ocean. Addition of the Mg suite to the overall flotation crust would imply a lower average FeO content, and thus a lower density, for the parental magma; which would in turn have engendered a higher overall plagioclase content (i.e., an Al_2O_3 content ≫29 wt%) for the flotation crust. However, even among samples derived primarily from the upper crust (Table 1), the overall Al_2O_3 concentration is clearly not ≫27 wt%.

5. The uncertainties in all three approaches (1–3 above) to estimating the overall mafic silicate content of the ferroan suite are such that the results appear to agree about

as well as could reasonably be expected. Unfortunately, there are probably few large pristine Apollo rocks that have not yet been examined for mode. The modes for many of the "known" large pristine rocks could be improved by further study, but in many cases a precise mode would require disaggregation of a major proportion of the total sample. Until we acquire more large rock samples from the lunar highlands, the average mafic content of the ferroan anorthositic suite may never be constrained better than to within a factor of two.

6. Should future studies indicate that the overall composition of the ferroan anorthositic suite is truly ferroan anorthosite (sensu stricto, ≥90 vol% plagioclase), the converses of conclusions 2 and 3 on crustal mass balance and magma-ocean crust flotation, respectively, would tend to suggest that the currently favored model of lunar-crustal genesis (magma-ocean plagioclase flotation followed by scattered Mg-rich intrusives) may be seriously flawed. The most likely alternative, from a petrologic perspective, would be to discard the magma-ocean hypothesis, in favor of purely "serial" magmatism (e.g., Walker, 1983). However, even the two-stage model for lunar petrogenesis without a magma ocean (Longhi and Ashwal, 1985), according to which the anorthosites originally accumulated earlier and deeper than the Mg-suite rocks, scarcely predicts the strong geochemical bimodality of the pristine rocks (Fig. 2). Moreover, the magma-ocean hypothesis has recently been strongly favored (in a reversal from the first 15 years of post-Apollo lunar science) by a consensus of geophysical models for the origin of the Moon (Stevenson, 1987).

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