

## A chemical model for generating the sources of mare basalts: Combined equilibrium and fractional crystallization of the lunar magmasphere

GREGORY A. SNYDER, LAWRENCE A. TAYLOR, and CLIVE R. NEAL\*

Department of Geological Sciences, University of Tennessee, Knoxville, TN 37996, USA

(Received October 8, 1991; accepted in revised form April 10, 1992)

**Abstract**—It is generally considered that mare basalts were generated by the melting of a cumulate mantle formed in an early Moon-wide magma ocean or magmasphere. However, the nature and chemistry of this cumulate mantle and the logistics of its origin have remained elusive. Extensive studies of terrestrial layered mafic intrusions over the past sixty years have emphasized the imperfection of fractional crystallization and attendant crystal-crystal and crystal-liquid separation in a convecting magma chamber. Crystal-liquid and crystal-crystal separations were similarly inefficient during evolution of the lunar magma ocean (LMO), allowing for the trapping of interstitial liquid and entrainment of a small proportion of less-dense plagioclase into the denser mafic cumulate mush. Indeed, petrography of lunar highlands samples demonstrates this for anorthosites (with 1–10% olivine). The residual liquid after 80–90% crystallization was very evolved (in fact KREEPy) and, even in small proportions (1–5%), would have a noticeable effect on the trace-element chemistry of melts generated from these cumulates. This trapped residual liquid would elevate total REE abundances in the cumulate pile, while synchronously deepening the already negative Eu anomaly. Essentially, this trapped liquid will make the cumulate more fertile for melting to generate both KREEP basalt and mare basalt magmas. Plagioclase entrained in the mafic cumulate pile adds an essential Al component to the high-Ti basalt source and will moderate the requisite negative Eu anomaly in the cumulate.

Early in the evolution of the lunar mantle, when the LMO still was largely liquid, it is likely that vigorous convection was an important factor in crystallization. Such convection would allow crystals to remain suspended and in equilibrium with the LMO liquid for relatively long periods of time. This extended period of equilibrium crystallization would then have been followed by fractional crystallization once plagioclase became a liquidus phase and began to float to form the lunar highlands crust.

Previous authors have proposed a three-component model for the evolution of high-Ti mare basalt source regions. This model includes KREEP, early (olivine-rich, high Mg#) cumulates, and late (ilmenite-rich, low Mg#) cumulates in various proportions. However, we propose a model for high-Ti basalt parent magmas which is in accord with studies of terrestrial layered intrusions. This model for the high-Ti source includes trapped instantaneous residual liquid (TIRL; 1–3%) and entrainment of a small (2–5%) proportion of plagioclase into the late-stage cumulate pile in order to account for both the observed Al compositions and trace-element characteristics of high-Ti mare basalts. Melting of this relatively shallow, ilmenite- and clinopyroxene-bearing, late-stage cumulate can generate high-Ti mare basalt magmas. Furthermore, we are in agreement with other workers that only through a process of nonmodal melting will the high Ti values for the parent magmas be realized. Large-scale convective overturn of the cumulate pile and mixing of KREEP with early- and late-stage cumulates is not required. However, localized overturn of the upper tenth of the cumulate pile is likely and, in fact, required to achieve an appropriate major-element balance for the high-Ti mare basalt source region.

### INTRODUCTION

THE MAGMA OCEAN CONCEPT has been an integral part of the lunar literature practically since the return of the first lunar samples (e.g., SMITH et al., 1970; WOOD et al., 1970; WARREN, 1985). Several authors have indicated problems and inconsistencies inherent in an early Moon-wide magma ocean and have challenged its validity (WALKER, 1983; SHIRLEY, 1983; LONGHI and ASHWAL, 1985). However, the bulk of petrologic, geochemical, and geophysical data for the highlands and mare basins of the Moon point to a common, ancient, global reservoir. The proceeding discussion will assume the existence of this early Moon-wide magma ocean.

Many variations of the lunar magma ocean (LMO) concept have been proposed (e.g., SHIH and SCHONFELD, 1976; HUGHES et al., 1988, 1989), but a few aspects appear to be common in currently favored models: (1) formation of a Moon-wide melt layer, followed by (2) crystallization and settling of mafic (olivine, pyroxenes, spinel) cumulates, and (3) flotation of plagioclase to form the lunar highlands. It is the mafic cumulate pile which is of greatest concern to those interested in modelling the source regions for mare basalts, and remelting of various layers of this mafic cumulate pile could have generated liquids which were parental to mare basalts (WALKER et al., 1975).

The body of studies on both lunar rocks and layered mafic intrusions which has accumulated over the years has consistently been refined and therefore produced more complicated models of igneous differentiation. Perfect adcumulates (i.e.,

\* Present address: Department of Civil Engineering and Geological Sciences, University of Notre Dame, Notre Dame, IN 45565, USA.

absolutely no trapped liquid) are rare, as are perfectly monomineralic rocks (i.e., dunites with exactly 100% olivine, and anorthosites with 100% plagioclase). This is neither Earth- (or should we say Moon-) shattering or unexpected. However, models for the evolution of the LMO generally have not reflected this fundamental observation. It is the intent of this paper to present a more realistic model for evolution and solidification of the LMO, which accounts for the inefficiency of crystal-crystal and crystal-liquid separations.

Several parameters (as follows) must be considered when developing a model for mare basalt source regions:

- 1) the bulk composition of the initial LMO;
- 2) the relevant phase equilibria at required pressures in order to map crystallization paths of the LMO;
- 3) the amount and composition of a trapped liquid component;
- 4) appropriate, lunar-relevant mineral/melt partition coefficients for the trace elements;
- 5) the possible extent of equilibrium crystallization and degree of perfect fractional crystallization of the LMO and subsequent crystal-crystal and crystal-liquid separation; and
- 6) the depth in the cumulate pile at which melting must occur to generate a mare basalt parent magma.

Recently, SHEARER and PAPIKE (1989) stated that plagioclase separation during generation of the mare basalt source may not be required to give the requisite negative Eu anomaly found in some mare basalts. MCKAY et al. (1991) pointed out that although pigeonite may have a deeper negative Eu anomaly under lunar conditions than previously thought, it cannot account for the considerably larger negative Eu anomaly required for mare basalt sources. Trace-element modelling of the high-Ti mare basalts (BROPHY and BASU, 1990; SNYDER et al., 1990b, 1991b) has indicated the need for a source with a significant negative Eu anomaly. Furthermore, SHAFFER et al. (1991) have reached similar conclusions on the basis of modelling the evolution of La/Sm ratios in mafic cumulates with and without plagioclase fractionation (i.e., plagioclase fractionation and removal is required to achieve a cumulate which will yield a mare basalt Eu signature). For most model LMO compositions, pigeonite will not crystallize until late in the fractionation sequence, either subsequent or consequent to plagioclase becoming a liquidus phase. Therefore, fractionation of pigeonite only enhances the negative Eu anomaly of the mafic cumulates late in the sequence and hastens the point at which the required mare basalt source is generated.

Mare basalts are generally considered to have originated by the melting of a cumulate mantle formed during the crystallization of this LMO. However, the nature and chemistry of this cumulate mantle and the logistics of its origin have remained elusive. In fact, many workers have suggested that the stratigraphy of the lunar upper mantle has been overturned during subsolidus convection due to inherent density contrasts in the cumulate pile (HERBERT, 1980; RYDER, 1991; SPERA, 1992). This convective overturn has been proposed on mineral/chemical grounds to explain the relatively high Mg# ( $Mg^{2+}/(Mg^{2+} + Fe^{2+})$ ; as reflected in mare basalts) of late-stage ilmenite-bearing LMO cumulates. However, large-

scale convective overturn of the lunar mantle may not be required if equilibrium crystallization was an important process in LMO differentiation (SNYDER and TAYLOR, 1992).

Early in the evolution of the lunar mantle, when the LMO still was largely liquid, it is possible that vigorous convection was an important factor in crystallization. Such convection would allow crystals to remain suspended and in chemical equilibrium with the LMO for relatively long periods of time. During this interval, equilibrium crystallization would control the differentiation of the LMO. However, plagioclase began to crystallize and float later in the evolution of the LMO (when convection was less vigorous) and was effectively separated from the rest of the cumulate. At that time, fractional crystallization would take over as the controlling mode of differentiation.

In this paper, we will attempt to model the sources of the mare basalts, in particular the high-Ti basalts, by the consideration of a modified mafic cumulate source formed during the combined equilibrium and fractional crystallization of a Moon-wide magma ocean. A refined approach to previous work (SNYDER et al., 1991b) was adopted for modelling the evolution of the LMO. This includes not only mafic crystal accumulation and plagioclase flotation but the incorporation of both a small proportion of entrained plagioclase (2–5%) and trapped instantaneous residual liquid (TIRL; <5%). Due to inherent complexities, previous studies often model either trace elements or major elements but fail to integrate the two. In this study, both trace and major elements have been modelled in the evolution of the LMO. As shown in the sections that follow, major elements tightly constrain the composition of mare basalt sources and the pathways to their creation.

Our presentation of this mare basalt source model will be in two stages. First, we will consider the parameters which influence the initial LMO and its subsequent crystallization. Second, through a case study involving the high-Ti basalts, we will test this LMO model and its ability to generate viable mare basalt source regions.

#### BULK COMPOSITION OF THE LUNAR MAGMA OCEAN

HUGHES et al. (1988, 1989) reviewed several suggested LMO compositions and chose the "Standard Initial" composition of WARREN (1986a) to represent the major-element composition of the initial LMO. Choosing a different initial LMO composition can significantly alter the fractionation sequence, but the following generalizations still hold: (1) olivine is always the first liquidus phase and crystallizes alone over a very broad range (up to 40–65% fractional crystallization), and (2) the final total proportion of plagioclase after complete crystallization of the LMO never exceeds 21% (also see WARREN, 1986b, for a similar conclusion based upon modelling of pristine nonmare rocks). The "Standard Initial" composition of WARREN (1986a) is relatively extreme in its  $Al_2O_3$  content (7 wt% vs. <5 wt%; BUCK and TOKSOZ, 1980; RINGWOOD and KESSON, 1976; TAYLOR and BENICE, 1975), as indicated in Table 1. In practice, this leads to the appearance of plagioclase on the liquids earlier in the fractionation of the LMO (57 PCS (Per Cent Solid) vs. 70–80 PCS in other

Table 1: Previously Proposed Bulk Silicate Moon, Initial LMO Compositions

	Tay <sup>1</sup>	RSW1 <sup>2</sup>	RSW2 <sup>2</sup>	WSF <sup>3</sup>	B&T <sup>4</sup>	J&D <sup>5</sup>	MIA <sup>6</sup>	O'N <sup>7</sup>
SiO <sub>2</sub>	43.5	43.2	44.2	46.0	48.4	43.5	43.3	44.6
TiO <sub>2</sub>	0.30	0.30	0.42	0.3	0.40	0.29	0.39	0.17
Al <sub>2</sub> O <sub>3</sub>	6.0	3.7	5.2	7.00	5.00	5.8	7.58	3.9
FeO	13.0	12.2	13.5	12.4	12.9	16.1	13.0	12.4
MnO	—	0.16	0.18	—	—	—	0.15	0.17
MgO	32.0	36.9	31.9	27.6	29.0	29.9	29.1	35.1
CaO	4.5	3.03	4.2	5.5	3.83	4.6	6.13	3.3
K <sub>2</sub> O	0.01	—	—	0.06	—	—	0.01	—
Na <sub>2</sub> O	0.09	0.06	0.08	0.6	0.15	—	0.10	0.05
Cr <sub>2</sub> O <sub>3</sub>	—	0.32	0.36	0.5	0.30	—	0.30	0.47
Mg#	0.81	0.84	0.81	0.80	0.78	0.77	0.80	0.83

<sup>1</sup>TAYLOR (1982).<sup>2</sup>RINGWOOD et al. (1987).<sup>3</sup>WARREN (1986a).<sup>4</sup>BUCK and TOKSOZ (1980).<sup>5</sup>JONES and DELANO (1989).<sup>6</sup>MORGAN et al. (1978).<sup>7</sup>O'NEILL (1991).

models, including that presented herein). Assuming an initial LMO depth of 300–500 km, this early appearance of plagioclase is inconsistent with the proposed thickness of the mostly anorthositic lunar crust (60–100 km; TAYLOR, 1987; MUELLER et al., 1988). The later appearance of plagioclase (at 75–80 PCS) is therefore preferred.

Another area of contention is the bulk LMO Mg#. WARREN (1986a,b) insisted, based upon convincing geochemical and geophysical arguments, that the initial LMO Mg# must be between 0.8 and 0.9. TAYLOR (1982) quoted an Mg# of 0.84 for the primitive lunar mantle. On the basis of "pristine" glass compositions, JONES and DELANO (1989) argued for a bulk lunar Mg# of 0.82–0.84; whereas LONGHI (1981) preferred an Mg# of 0.86, based upon mineralogic and chemical criteria. MUELLER et al. (1988) suggested an upper mantle Mg# of 0.78, and O'NEILL (1991) has come out in favor of a bulk Moon Mg# of 0.83. For the present study, we will use an intermediate value of 0.82. Assuming a slightly higher or lower Mg# for the initial LMO (say 0.86 or 0.78, respectively) does not greatly affect either the crystallization sequence or the point of appearance of ilmenite in the crystallization sequence (as this occurs late, in any event).

The proposed chemical model assumes a starting composition for the LMO similar to the proposed bulk moon composition of BUCK and TOKSOZ (1980) (Table 2; modified Mg# = 0.82, Al<sub>2</sub>O<sub>3</sub> = 5.0 wt%, CaO = 3.83 wt%, TiO<sub>2</sub> = 0.4 wt%). This composition is intermediate in most respects (specifically Mg# and Al<sub>2</sub>O<sub>3</sub> content) to those which have been proposed. However, it must be pointed out that even this intermediate Mg# would be considered by some workers to be too low. Some would contend that the Mg# of the upper mantle of the Moon may even be as high as that of the Earth (approximately 0.89–0.90). An initial magma ocean depth of 400 km is also specified, though not necessary to the conclusions presented.

For the lithophile trace elements (including the REE), TAYLOR (1982) suggested the bulk Moon should have approximately 2.5× chondritic abundances. However, if it is presupposed that the bulk Moon underwent a previous dif-

ferentiation event which included the crystallization of olivine alone (TAYLOR, 1982; JONES and DELANO, 1989), the "residual" LMO (the initial LMO for our model) would then have been enriched further in these elements (as olivine has extremely low mineral/melt distribution coefficients for the lithophile elements; see MCKAY, 1986). With this rationale, a value of 3× chondritic is preferred for the trace elements (cf. HUGHES et al., 1988, 1989) and would be consistent with this differentiation event if approximately 20% of the total Moon had fractionated to yield olivine alone. Further, as a means of comparing our results with those of HUGHES et al. (1988, 1989), we continue to use a flat, 3× chondrites [or approximately 4–5× CI chondrites as per the recent compilations of WASSON and KALLEMEYN (1988) and ANDERS and GREVESSE (1989)] LILE pattern.

### RELEVANT PHASE EQUILIBRIA AND CRYSTALLIZATION PATH

The pseudoternary system olivine-plagioclase-silica was utilized to represent the crystallization path of the LMO at low pressure (Fig. 1). This ternary phase diagram allows an adequate representation of the lunar mantle and is preferred on the basis of the known mineralogy and phase equilibria of mare basalts.

The debate continues as to where crystallization occurs within a magma ocean. In recent years, in situ crystallization (i.e., crystallization at depth, near the base of a magma chamber) has become popular (LANGMUIR, 1989). The modelling presented in this paper assumes crystal nucleation at or near a boundary layer where the temperature gradient and heat flux are the greatest. Again, this assumption does

Table 2: Initial LMO and High-Ti Basalt Parent Compositions

	LMO	AP-11, B3 parent	AP-17, C parent
SiO <sub>2</sub>	48.4	39.6	38.5
TiO <sub>2</sub>	0.40	11.1	12.3
Al <sub>2</sub> O <sub>3</sub>	5.0	9.51	8.7
FeO	12.0	19.1	18.5
MnO	—	0.28	0.26
MgO	29.9	8.10	9.7
CaO	3.83	11.1	10.1
Na <sub>2</sub> O	0.13	0.36	—
K <sub>2</sub> O	0.04	0.05	0.07
Cr <sub>2</sub> O <sub>3</sub>	0.30	0.47	0.61
Mg#	0.82	0.43	0.48
Ba	10.8	95	67
Rb	1.0	0.80	1.2
Sr	34.2	144	160
Th	0.15	—	—
Nb	1.3	—	—
La	1.10	6.7	6.3
Ce	2.87	22.5	21.8
Nd	2.13	21.1	23.8
Sm	0.69	8.4	9.5
Eu	0.26	1.5	1.8
Gd	0.92	13.2	14.8
Tb	0.17	2.1	—
Dy	1.14	14.5	16.9
Er	0.75	9.7	9.7
Yb	0.74	8.5	8.8
Lu	0.11	1.2	1.2
Zr	17.1	—	—
Hf	0.51	—	—
Sc	23.4	—	—

Sources: LMO = BUCK and TOKSOZ (1980) for major elements; AP-11, B3 basalt = sample 10045 from AGRELL et al. (1970) and HASKIN et al. (1970); AP-17, C = average of 74245, 74275, 74247 for trace elements from RHODES et al. (1976), WARNER et al. (1979), and NEAL et al. (1990) for major elements.

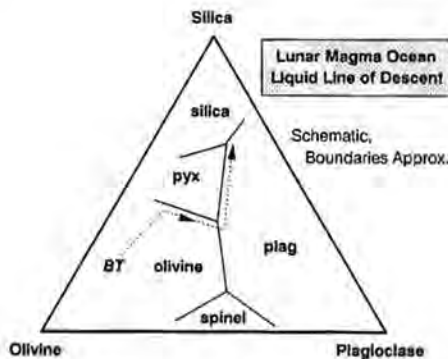


FIG. 1. Lunar magma ocean crystallization path (schematic) plotted on a modified olivine-plagioclase-silica ternary phase diagram (the so-called "Walker diagram") at 6 kbars of pressure.

not grossly affect the model, as equilibrium crystallization is expected to dominate during the first three-quarters of LMO evolution. Because of the increased heat flow and larger thermal gradient expected near the surface of a cooling body, the bulk of crystal nucleation and growth should occur within the upper and outer portions of the LMO. This location corresponds in the Moon to the upper portion of a bulk Moon-wide melt layer (the LMO) adjacent to the nascent lunar crust. Mafic crystals formed near the boundary layer would then become entrained in the vigorously convecting liquid. Once the temperature has decreased and the liquid has become sufficiently laden with crystals (78 PCS in the model presented herein), mafic crystals would begin to sink in response to their density contrast with the residual LMO, be carried downward by convection currents within the LMO, or otherwise be transferred to the lower reaches of the LMO "chamber" by a process which is a hybrid of both processes. Undoubtedly, it is true that during the crystallization of the LMO in this fashion, early formed crystals will not settle to the floor without interaction with the intervening magma. Resorption and recrystallization phenomena are possible, even likely, but are not considered to be within the scope of this paper.

The phase diagram of LONGHI (1981), depicted in Fig. 1, gives approximate proportions of the crystallizing minerals which are strikingly similar to those deduced by HUGHES et al. (1988, 1989) and by GROMET et al. (1981) in their study of late-stage KREEP formation from the LMO. In order to constrain more precisely the crystallization sequence for the LMO, the programs MAGPOX and MAGFOX (LONGHI, pers. comm.) were utilized. The model includes equilibrium crystallization for the first three-quarters of LMO differentiation during rapid convection (at 6 kb pressure), followed by low-pressure in situ fractional crystallization of the magma ocean. This latter phase was accompanied by trapping of 1% instantaneous residual liquid and entrainment of 1 modal% plagioclase (once it became a liquidus phase at 78 PCS). Most of the crystallizing plagioclase floated to form the lunar highlands. The fractionation sequence, as determined by using modified output from the programs MAGPOX and MAGFOX, is as follows:

- 1) 0–40 PCS = olivine;
- 2) 40–78 PCS = orthopyroxene (with some olivine resorption);
- 3) 78–86 PCS = 53% plagioclase + 25% olivine + 22% pigeonite;
- 4) 86–95 PCS = 38% clinopyroxene + 36% plagioclase + 26% pigeonite; and
- 5) 95–99.5 PCS = 34% pigeonite + 31% plagioclase + 24% clinopyroxene + 11% ilmenite.

This extended crystallization of orthopyroxene (up to 78 PCS) with only minor olivine resorption may at first seem counterintuitive. LONGHI (1978, 1980, 1981) has pointed out that, with increasing pressure, the reaction curve between olivine and orthopyroxene (opx) moves toward the olivine corner in the olivine-anorthite-silica pseudoternary phase diagram. This effectively decreases the proportion of olivine resorption with increasing pressure. At low pressures (near 1 bar), the olivine-opx phase boundary lies on the high-silica side of the Alkemade line, rendering this field boundary a reaction curve. As the liquid composition moves along this curve towards the reaction (peritectic) point, olivine reacts with the liquid to form opx. With increasing pressure, the olivine phase field shrinks relative to silica until, at between 5–6 kbar, the olivine-opx reaction curve lies along the low-silica side of the opx-plagioclase Alkemade line (LONGHI, pers. comm.). Thus, the olivine-opx reaction curve becomes a true cotectic field boundary curve. Likewise, the cotectic with plagioclase moves away from the olivine corner with increasing pressure, thereby delaying the appearance of plagioclase on the liquidus. These observations make the determination of the depth of crystallization (and eventual melting) of the LMO significant to developing crystallization paths for the LMO. Since the equilibrium crystallization calculations were performed at 6 kbar, the amount of olivine resorption is smaller than it would be at low pressure.

The stratigraphy of the upper mantle, which is a product of the equilibrium/fractional crystallization model presented here is similar to others previously presented (Fig. 2). The major difference is the relatively late appearance of clinopyroxene (after plagioclase) in the present model (at 86 PCS). Ilmenite begins to precipitate above 95 PCS, as opposed to 92–93 PCS in other models (SNYDER et al., 1991a,b; SNYDER and TAYLOR, 1992; TAYLOR, 1982). After approximately three-quarters of the LMO has crystallized, plagioclase becomes a liquidus phase. The bulk of this plagioclase is subsequently floated and does not become part of the mafic cumulate + trapped liquid pile. However, a small proportion of this plagioclase (2–5 wt%; see following sections) is entrained into the sinking cumulate mush. Generation of mare basalts from a cumulate containing only mafic minerals, left behind after plagioclase removal by flotation, is not a unique idea (see TAYLOR, 1982). However, previous geochemical models for the genesis of mare basalt sources do not include the TIRL component or an entrained plagioclase component.

#### INEFFICIENCY OF CRYSTAL-LIQUID AND CRYSTAL-CRYSTAL SEPARATION

A model is presented for crystallization of the LMO that is believed to more closely approximate both the early tur-

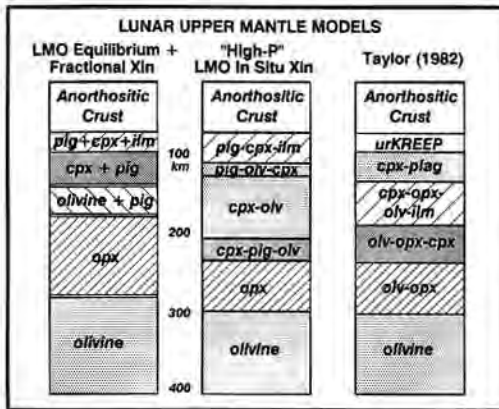


FIG. 2. Comparison of the lunar upper mantle presented in this paper with others presented recently (SNYDER *et al.*, 1991a,b; TAYLOR, 1982). Abbreviations (used throughout diagrams): cpx = clinopyroxene, ilm = ilmenite, olv = olivine, opx = orthopyroxene, pig = pigeonite, and plag = plagioclase.

bulent history of the LMO and its later, more quiescent phase when imperfect fractional crystallization was the controlling method of differentiation. In this model, the LMO is allowed to undergo equilibrium, followed by fractional crystallization. However, once fractional crystallization became the dominant mode of differentiation, both crystal-liquid and crystal-crystal separations were inefficient. At successively greater stratigraphic heights in the cumulate pile, a variable (1–5%) proportion of instantaneous residual liquid was trapped in the cumulate pile. In this model, 1 and 5% TIRL are used; and we assume that perfect adcumulates do not exist. The composition of the trapped liquid is calculated assuming equilibrium with the mafic cumulate and that, once trapped in the cumulate pile, the liquid does not communicate chemically with the residual LMO.

In terrestrial layered mafic intrusions, it is known that the separation efficiency of crystals of grossly different densities is seldom 100% (e.g., IRVINE, 1980). For instance, layers of troctolitic composition are much more common than those of purely anorthositic or dunitic composition. There is ample reason to believe that crystal-crystal separations in an LMO would be similarly inefficient, as evidenced by the lack of truly pure anorthosites in the highlands crust. Once plagioclase becomes a liquidus phase, a small proportion of this less dense mineral is also entrained into the sinking mafic cumulate mush. In turn, entrainment of a small proportion of more dense material (mafic cumulate) is likely to be entrained into the floating plagioclase cumulate. Furthermore, plagioclase is a difficult mineral to nucleate. It is likely that it would nucleate heterogeneously, preferring a mineral surface for a nucleation site. With this in mind, we will model the LMO cumulate pile with both the TIRL component ( $\leq 5\%$ ) and, once it begins to crystallize, a small amount ( $\leq 5\%$ ) of entrained plagioclase.

The presence of trapped liquid elevates the extremely low REE abundances found in the mafic cumulates alone. Only prohibitively low degrees of melting ( $\leq 1\%$ ) of this mafic cumulate could generate a liquid with the appropriate abun-

dances of the REE present in mare basalts. However, at these extremely low degrees of partial melting, derived liquids would be extremely LREE-enriched and would not match the analyzed REE patterns of mare basalts. Therefore, the addition of this TIRL component has a two-fold effect: (1) to elevate the REE abundances of the total cumulate pile; and (2) after plagioclase starts to crystallize, it is integral to the generation of the requisite negative Eu anomaly in the mare basalt source region.

#### CHOICE OF MINERAL/MELT PARTITION COEFFICIENTS

SNYDER *et al.* (1990b, 1991b) attempted to model the evolution of the LMO cumulate-liquid system using trace-element partition coefficients published by HUGHES *et al.* (1988, 1989). We initially used the values given by HUGHES *et al.* (1988, 1989) in order to compare more closely differences in these two models, devoid of differences in model details. However, we have since searched the literature and have generated a list of partition coefficients which we consider to represent more closely the lunar scenario.

Recent work by MCKAY *et al.* (1991) on the partitioning of trace elements between pigeonite and basaltic liquid has led to conclusions which are significantly different from those previously published (MCKAY *et al.*, 1986). The same is true for the recent work of PHINNEY and MORRISON (1990) on the plagioclase-melt system. It appears that advances in both theory and method have led to the determination of much lower (sometimes by as much as an order of magnitude) mineral/melt partition coefficients for certain trace elements. We have compiled a list of partition coefficients, which is both up to date (at the time of publication) and, where the data are available, represents the best estimate of partitioning behavior under lunar conditions. This list is compiled in Table 3, and these values were used throughout this study. A glaring deficiency of this list is the lack of available data on trace elements other than the REE. Because the REE are a coherent group of elements which behave in a predictable fashion, interpolation of abundances within the group is straightforward. However, estimation of mineral/melt  $k_D$ s for the other trace elements is difficult, and attendant errors are large.

#### MODELLING OF MAJOR ELEMENTS DURING LMO EVOLUTION

Modelling of major elements was undertaken in this study to provide stringent constraints on the plausible processes involved in mare basalt source generation, and LMO liquid evolution. The methods employed are similar to those from the pioneering work of HANSON and LANGMUIR (1978) and LANGMUIR and HANSON (1980) on terrestrial rocks. The algorithms of J. Longhi (pers. commun.) were followed throughout for the major elements.

The major oxides MgO, FeO, and SiO<sub>2</sub> are constrained in the first 40% of LMO crystallization by their stoichiometry in the mineral olivine (HANSON and LANGMUIR, 1978):

$$X^{\text{MgO}} + X^{\text{FeO}} = 0.6667, \quad (1)$$

where 0.6667 is the mole fraction of octahedral sites in olivine,

Table 3: Mineral/Melt Partition Coefficients

Element	oliv <sup>1</sup>	opx <sup>2</sup>	gpx <sup>3</sup>	plg <sup>4</sup>	plau <sup>5</sup>	ilm <sup>6</sup>
Sc	.27	1.6	1.6	1.6	.0071	1.5
Cr	3	5	5	[5]	.0332	4.2
Co	5	1.3	1.2	[1.3]	.0202	4.3
Ni	30	2.5	2.5	[2.5]	[0.01]	3.5
Rb	.0001	[0.023]	.0257	[0.0026]	0.088	[0.001]
Sr	.0001	.018	.097	.0020	1.61	[0.001]
Zr <sup>9</sup>	0.013	0.063	0.063	0.063	0.0128	0.406
Nb <sup>10</sup>	0.0001	[0.02]	0.02	[0.02]	0.005	0.80
Ba	.0001	0.013	0.013	0.013	0.686	0.005
La	.0001	.007	.023	.0009	.0418	.0024
Ce	.0001	.009	.039	.00172	.0302	.0019
Nd	.0001	.014	.104	.0058	.0236	.0012
Sm	.0005	.022	.17	.011	.0170	.0023
Eu	.0007	.015	.16	.0068	1.2	.0009
Gd	.001	.037	.20	.021	.0105	.0053
Tb	.002	.048	.21	.027	.0095	.0082
Dy	.003	.060	.23	.034	.0089	.013
Er	.008	.10	.26	.055	.0077	.031
Yb	.019	.170	.29	.087	.0065	.057
Lu	.03	.22	.30	.11	.0068	.070
Hf	.013	0.063	0.063	0.063	.0128	.406
Th	0.03	0.13	0.13	0.13	.0208	0.55

References: <sup>1</sup>McKAY (1986); <sup>2</sup>1200°C run from WEILL and McKAY (1975) and as estimated from their Fig. 6; <sup>3</sup>McKAY et al. (1986); <sup>4</sup>McKAY et al. (1991); <sup>5</sup>PHINNEY and MORRISON (1990); <sup>6</sup>NAKAMURA et al. (1986); <sup>7</sup>SHIMIZU (1974); <sup>8</sup>McKAY and WEILL (1977); <sup>9</sup>Zr partition coefficients were assumed to be the same as those of Hf; <sup>10</sup>Nb data were taken or estimated from McCALLUM and CHARETTE (1978). Many REE "data" are interpolated from data in these sources; <sup>11</sup>Eu value for opx is estimated assuming a similar magnitude Eu anomaly as pigeonite and that for plagioclase is from WEILL and McKAY (1975). Some "gaps" in the data base (especially in Sc, Ba, Ta, Hf, and Th) have been filled in by previously used estimates generated in NEAL et al. (1989). Rb data for opx and pigeonite were calculated assuming Rb/Sr ratios given in ARTH (1976) and Sr partition coefficients from above. Transition metal (Cr, Co, Ni) partition coefficients were taken from VILLEMANT et al. (1981), BIRD (1971), and LINDSTROM (1976).

with SiO<sub>2</sub> making up the other one-third mole fraction; and by the exchange reaction distribution coefficient,

$$K_D = (X_{\text{olivine}}^{\text{FeO}} * X_{\text{liquid}}^{\text{MgO}}) / (X_{\text{olivine}}^{\text{MgO}} * X_{\text{liquid}}^{\text{FeO}}). \quad (2)$$

Combining Eqns. (1) and (2) leads to the following useful equation:

$$C_{\text{olivine}}^{\text{MgO}} = (0.6667)(C_{\text{liquid}}^{\text{MgO}}) / (C_{\text{liquid}}^{\text{FeO}} * K_D + C_{\text{liquid}}^{\text{MgO}}), \quad (3)$$

where  $C$  = concentration, and  $X$  = mole fraction of MgO or FeO in the liquid or olivine as designated in the superscripts and subscripts, respectively. Finally, once a "package" of crystals has fractionated out of the liquid, the residual liquid may be calculated using the following "mole balance" equation:

$$C_{\text{res liq}}^{\text{MgO}} = [C_{\text{init liq}}^{\text{MgO}} - (C_{\text{olivine}}^{\text{MgO}})(X)] / (1 - X), \quad (4)$$

where the subscripts "res liq" and "init liq" represent the residual liquid and initial (starting) liquid, respectively. Again, analogous equations may be derived for both FeO and SiO<sub>2</sub>.

As crystallization exceeds approximately 50%, orthopyroxene will also exhibit control over these oxides. The stoichiometry of orthopyroxene allows the derivation of the following analogous equation:

$$X^{\text{MgO}} + X^{\text{FeO}} = 0.50, \quad (5)$$

as well as a similar  $K_D$  equation. The exchange-reaction distribution coefficients for olivine/liquid and pyroxene/liquid have been determined experimentally and found to be nearly constant with changes in liquid composition and pressure. The  $K_D$  for olivine/liquid is roughly 0.32 (ROEDER and EM-

SLIE, 1970; and appended in LONGHI, 1981) and was shown to vary measurably with the TiO<sub>2</sub> of the liquid (DELANO, 1990). Those for orthopyroxene/liquid and clinopyroxene/liquid are 0.30 and 0.28, respectively (LONGHI, 1981). Finally, and in a similar fashion, the succeeding minerals on the liquidus, as the LMO crystallizes, also control major-element evolution according to stoichiometry. Plagioclase and clinopyroxene both exhibit controls on Ca and Si; plagioclase affects changes in Al; and clinopyroxene complicates the evolution of Mg and Fe in both the LMO and its accumulating minerals.

#### MAJOR-ELEMENT EVOLUTION OF THE LMO

Using the proposed fractionation assemblage, starting composition, and exchange reaction distribution coefficients from the literature and discussed above, major elements have been modeled for the proposed LMO up to 99.5 [PCS by volume, as originally defined by MORSE (1979)]. Above about 98 PCS, it becomes exceedingly difficult to model the major element evolution, due to uncertainties in exact mineral compositions and proportions (e.g., one starts to exhaust major oxides if the mineral compositions and proportions are not well known). Also, it is likely that any residual liquid which is greater than 98 PCS will have already split into two immiscible liquids (HESS et al., 1975) if that liquid has continued to follow a Fenner trend for fractional crystallization.

The evolution of the major oxides and Mg# of the residual LMO are shown in Fig. 3. It can be seen that MgO decreases continuously from start to end. SiO<sub>2</sub> is nearly constant for the first 90 PCS of LMO evolution and then decreases until ilmenite becomes a liquidus phase, where it increases. CaO increases steadily to a maximum of 12 wt% at about 85 PCS, and Al<sub>2</sub>O<sub>3</sub> increases to a maximum of approximately 14 wt%, where plagioclase becomes a liquidus phase (78 PCS), at which points both begin to decrease. The point at which the LMO reaches saturation with respect to ilmenite affects the shape of the curves for FeO and Al<sub>2</sub>O<sub>3</sub>, but in opposing manners. Note that Al<sub>2</sub>O<sub>3</sub> is in the range of 10–12 wt% between 90–95 PCS. The oxide TiO<sub>2</sub> shows a similar relation to FeO,

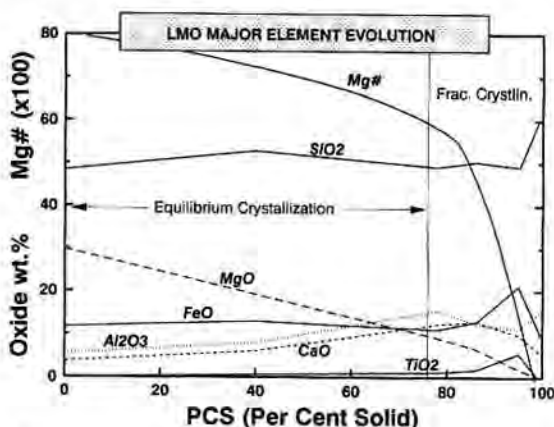


FIG. 3. A plot of wt% oxides (and Mg# × 100) vs. PCS for the LMO liquid. This plot indicates the evolution of the LMO as the ocean fills with cumulates.

increasing rapidly, until ilmenite reaches the liquidus (at 93–95 PCS), and then dropping off rapidly. Extended equilibrium crystallization of the LMO leads to the “buffering” of the liquid at relatively high Mg#. At 86 PCS, the Mg# of the liquid has still only fallen to 0.49 (Fig. 3). The Mg# of the mafic cumulate at 86 PCS is 0.78. However, from this point on, the Mg# of the liquid and the corresponding Mg# of the mafic cumulate drops off markedly (due to fractional crystallization becoming the dominant process for differentiation of the liquid). Ilmenite does not become a liquidus phase until 95 PCS, at which point the Mg# of the liquid is 0.15 and that of the corresponding mafic solid is 0.34.

The mafic cumulates (Fig. 4) show sympathetic trends to the melt, albeit in a more stepwise fashion. It has been stated previously that the LMO could have convected vigorously for most of its differentiation history, thus keeping crystallizing minerals suspended. Therefore, it is likely that the lower three-quarters of the lunar upper mantle consists of nearly equal proportions of olivine and orthopyroxene. In this case, the trends seen in Fig. 4 would be smooth up to 78 PCS. Figure 4 indicates the major-element trends if olivine and orthopyroxene precipitated separately, forming a dunite layer at the base and an orthopyroxene layer above. Again, it is important to point out the precipitous drop in MgO and Mg# above 86 PCS, once clinopyroxene comes on the liquidus. Note also the expected jump in FeO and TiO<sub>2</sub> once ilmenite begins to crystallize at 95 PCS.

For instructive and display purposes, the composition of the LMO cumulate pile will be dealt with in a three-step manner in proceeding discussions. First, the composition of the mafic cumulate alone will be presented; then the mafic cumulate + varying percentages of TIRL; and finally, the complete model featuring the mafic cumulate + TIRL + varying proportions of entrained plagioclase. By using this stepwise approach to the presentation, we hope to indicate the relative importance and effect of each component at successive stages in the evolution of the LMO and lunar mantle.

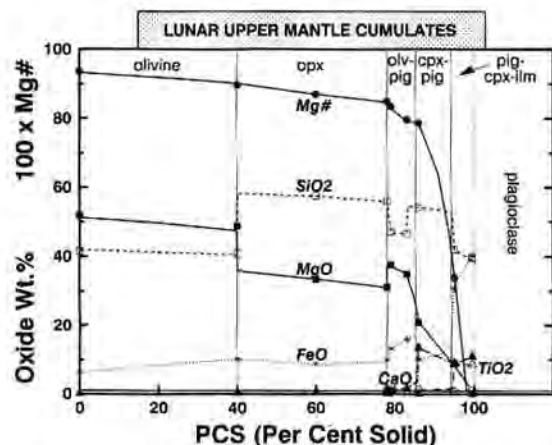


FIG. 4. A plot of wt% oxides (and Mg#) vs. PCS for the mafic LMO cumulate pile (including only 1 wt% plagioclase once it comes on the liquidus). Open triangles are for TiO<sub>2</sub>; closed triangles are for CaO.

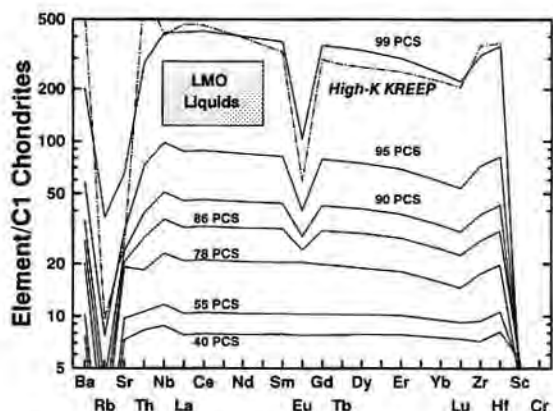


FIG. 5. Plot of a suite of modeled trace elements at various stages (40, 55, 78, 86, 90, 95, and 99 PCS, where PCS = per cent solid) in the evolution of the LMO liquid normalized to the C1 Chondrites. High-K KREEP (WARREN, 1989) is also plotted for comparison.

#### TRACE-ELEMENT EVOLUTION OF THE LMO

Trace-element evolution of the residual liquid and the mafic-cumulate pile + trapped liquid + entrained plagioclase pile are illustrated in Figs. 5–8 and labelled with the respective PCS values. Extended equilibrium crystallization (for 78 PCS) is nearly indistinguishable from the trace-element evolution of the LMO displayed by simple fractional crystallization (SNYDER et al., 1991b; SNYDER and TAYLOR, 1992). This is due to the fact that most trace elements are incompatible with mafic minerals. Due to the low  $K_{D,S}$ , resulting in low bulk  $D_s$  for the cumulates, equilibrium crystallization mathematically reduces to fractional crystallization.

As pointed out by many previous workers (e.g., GROMET et al., 1981; HUGHES et al., 1988), REE evolution of the LMO residual melt becomes more and more KREEPy as crystallization of the LMO continues (Figs. 5 and 7a). Once plagioclase starts to crystallize, the LMO residual melt takes on a pronounced negative Eu anomaly. Because of the continual removal of mafic minerals (which generally fractionate the LREE relative to the HREE), the LMO also begins to show LREE enrichment as the final dregs of melt are approached. NYQUIST et al. (1976) have also pointed this out and concluded that the LMO may have started out LREE-enriched. If this is so, then this places even more stringent restrictions on the percentage of melting of cumulate sources produced from such an ocean. It is evident in Fig. 5 that late-stage, magma-ocean, residual liquids are similar to lunar high-K KREEP (WARREN, 1989).

On the other hand, the mafic cumulate pile alone displays LREE depletion throughout its evolution (Fig. 6a), but also begins to take on a pronounced negative Eu anomaly once plagioclase comes on the liquidus. Addition of only 1% TIRL boosts the total REE (by an order of magnitude for the LREE) and moderates the LREE depletion but maintains the strong negative Eu anomaly (Fig. 6b and 7a). Addition of 5% or greater TIRL all but eliminates the LREE-depleted signature (which is required for the mare basalt source, as will be shown later) of the cumulate pile. The degree of LREE depletion,

and change in the Eu anomaly, of the cumulate pile with varying percentages of TIRL during evolution of the LMO is shown in Fig. 6a–c and Fig. 7a. Above 90 PCS, the mafic cumulate takes on a pronounced LREE-depleted signature, due in part to the crystallization of clinopyroxene and pi-

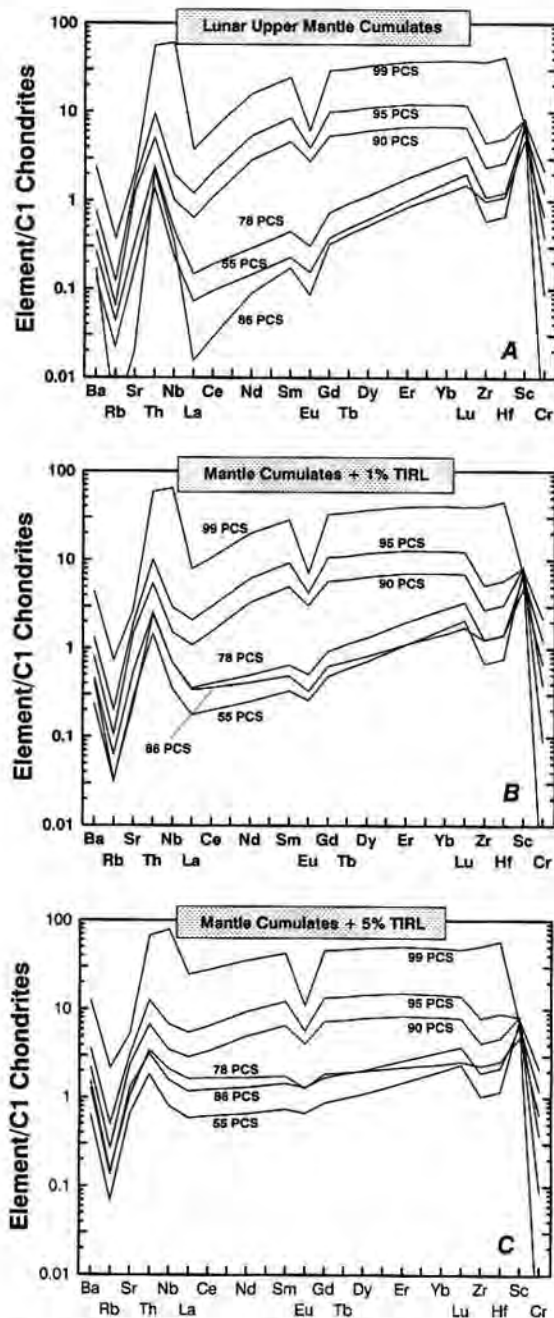


FIG. 6. Plots of a suite of trace elements for evolution of the mafic cumulate pile: (a) LMO mafic cumulates alone; (b) LMO mafic cumulates with 1% trapped liquid (TIRL); (c) LMO mafic cumulates with 5% TIRL. PCS values are indicated on each figure.

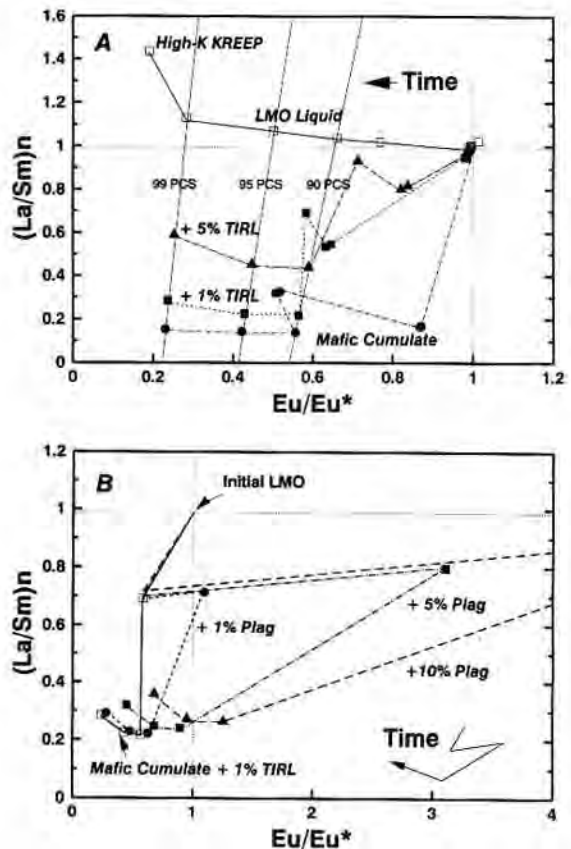


FIG. 7. Plots of trace element ratios vs.  $Eu/Eu^*$  (depth of Eu anomaly) on the abscissa: (a) with  $(La/Sm)_n$  plotted on the ordinate for the mafic cumulates, and mafic cumulates + 1 and 5% TIRL, along with a plot of LMO liquid evolution and high-K KREEP (WARREN, 1989); and (b) with  $(La/Sm)_n$  plotted on the ordinate for mafic cumulates + 1% TIRL and 1, 5, and 10% entrained plagioclase.

geonite. Melts taken from above this level would retain relatively depleted Nd isotopic signatures over time. Above about 78 PCS (after plagioclase becomes a liquidus phase), the Eu anomaly varies little with the proportion of trapped liquid in the cumulate. However, the degree of LREE depletion (as indicated by the ratio  $(La/Sm)_n$ ) is drastically decreased by addition of just a few percent of TIRL (Fig. 7a). Above about 5% TIRL added, the mafic cumulate after 90 PCS is effectively swamped by the LREE pattern of the LMO residual liquid.

It is evident at this point that the modelling encounters two competing problems. In order to get the absolute abundances of the REE in the mare basalt source up to levels that will make it appropriate for melting to generate high-Ti basalts, an added component with higher REE (TIRL) must be included. However, the TIRL component has either a chondritic LREE pattern or is slightly LREE-enriched and therefore must make up a small proportion of the source in order to maintain the required overall LREE-depleted signature.



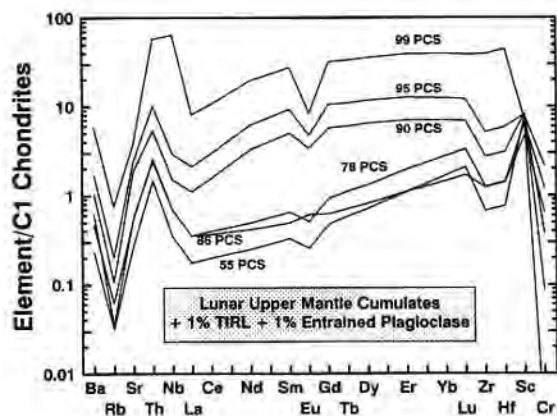


FIG. 8. Plot of a suite of trace elements for evolution of the mafic cumulate pile + 1% TIRL and 1% entrained plagioclase.

Also notable are the large depletions of Sc, Sr, and Rb in the evolving LMO liquid (Fig. 5). This is mirrored by the enrichment of Sc in the earliest LMO mafic cumulates. Niobium and Th indicate enrichments in the mafic cumulate throughout, with Th becoming relatively more depleted above 95 PCS once ilmenite starts to crystallize. Chromium is depleted early in the LMO liquid and is enriched in the earliest cumulates (where chromite is likely to crystallize with olivine). Chromium becomes more depleted in the cumulate with PCS. Barium becomes relatively more depleted in the liquid with time (especially above 78 PCS, where plagioclase starts to crystallize). Above 78 PCS, Rb partitions into the residual LMO (i.e., is incompatible with the mafic cumulate) increasing the Rb/Sr ratio with PCS value (Fig. 5). Hafnium is depleted in the cumulate early but becomes even relative to the normalized REE above 95 PCS, when ilmenite is crystallizing.

The entrainment of a small proportion of plagioclase into the cumulate source has some interesting effects. If the proportion is small (<5%), there will be little effect on the major-element composition of the bulk source; however, if plagioclase was an early melting phase, and melting percentages were low, it could exert an important influence on the major-element composition of the derivative melt. Even small amounts of plagioclase will have a significant effect on both the Eu and Sr values of the bulk cumulate (Figs. 7b, 8). Europium and Sr are more compatible with plagioclase than with mafic minerals by over an order of magnitude (Table 3). Thus, as seen in Figs. 7b and 8a, addition of an entrained plagioclase component drastically alters the requisite negative Eu anomaly and Rb/Sr ratio of the source. In fact, addition of only 10% plagioclase to the source regions of the mare basalts will effectively delete the required negative Eu anomaly. By this observation, any model involving addition of plagioclase into the source is tightly constrained to be much less than 10 wt% plagioclase in the cumulate.

#### HIGH-Ti MARE BASALT SOURCE REGIONS: A CASE STUDY

In order to demonstrate the applicability of this model, we have chosen as our case study an important group of basalts—

the high-Ti mare suite. It is felt that if the model is successful in explaining this complex suite of basalts, modelling of the petrogenesis of other basalt types will be even more readily accomplished. However, before we can proceed with modelling the high-Ti basalts, a parental liquid (mantle-derived magma, unmodified during ascent and emplacement) must be determined.

#### Determining Parental Mare Basalt Magmas

SNYDER et al. (1991a) have pointed out that the major- and trace-element abundances of the most primitive (i.e., lowest incompatible element abundances) basalts from the Apollo 11 low-K (B3) suite and from the Apollo 17 (Type C) high-Ti suites are practically indistinguishable (Table 2). It is proposed that both suites were derived from similar source regions by similar processes yielding a similar, if not a common, parent magma. This parental magma had trace-element abundances approximately 10–15× the bulk Moon (TAYLOR, 1982), was LREE-depleted, and had a pronounced negative Eu anomaly. Further evidence for the similarity of the Apollo 11 B3 and Apollo 17 Type C sources is shown by the modelling of HUGHES et al. (1989). Though they did not delineate primitive parental liquids, the chemical composition of their calculated sources for high-Ti Apollo 11 B3 and Apollo 17 Type C basalts are almost identical.

The most primitive compositions in each high-Ti mare basalt suite have been determined, either by calculation or simple inspection. In delineating parental magma compositions, several criteria (as follows) must be met:

- 1) The sample under consideration as a parent should be relatively free of phenocrysts, or the phenocryst compositions must be somehow "subtracted out" in order to arrive at a liquid composition.
- 2) It should be relatively fine-grained or vitrophyric as to make it a reliable candidate for a liquid.
- 3) It should contain the lowest incompatible element abundances (relative to other phenocryst-free samples) of that basalt type.
- 4) It should contain relatively high abundances of compatible elements.
- 5) It should exhibit, in general, the more primitive compositions (high Mg#) found within each suite.

These conventions help alleviate confusion in the mineralogy and chemical composition of the calculated sources that may be inherent in simply averaging a group of samples (HUGHES et al., 1988, 1989, and many others), especially those samples which may have experienced post-magma generation evolutionary processes (e.g., fractional crystallization, AFC (assimilation, fractional crystallization), and accumulation of phenocrystic phases). We consider the parental compositions, selected using these criteria, to represent a best estimate of the parental magma for each suite at this time. It must be pointed out that these parental magma compositions may or may not be primary magmas. The Mg#s are quite low. However, if one assumes that the LMO crystallized from bottom to top and was not well mixed during this process, the source region for the late-stage, high-Ti basalts would have an evolved composition.

DELANO (1980, 1986) has indicated that basaltic glasses are the most primitive (highest Mg#) expression of volcanism on the Moon. However, a genetic link between the basaltic glasses and actual mare basalts has yet to be elucidated (DELANO, 1986). The relationship and relevance of any experimental studies of basaltic glasses to the problem of mare basalt genesis must be considered equivocal at this time. Therefore, actual mare basalts from within each group have been chosen to delineate a parental magma for that group.

### Major Element Constraints on the High-Ti Source

A continual problem with approaches similar to that outlined above is the seeming lack of consistency of the trace-element modelling with major-element constraints. Using only trace-element constraints, we have shown (along with many others, including UNRUH et al., 1984; HUGHES et al., 1988) that approximately 1–5% melting of an LMO cumulate source region, after 95 PCS (when ilmenite is a liquidus phase in the LMO), with only pyroxenes, olivine, and ilmenite in the residue will generate magmas parental to high-Ti mare basalts. However, important constraints are also placed on the character and depth of the mare basalt source region (or degree of fractional crystallization of the LMO) by the major element compositions (and, specifically, the Mg#) of the evolving cumulates and residual liquid. For the high-Ti basalts, the Mg# of the parent is approximately 0.45 (average of high-Ti basalt parent values in Table 2). Therefore, any predominantly mafic cumulate which was melted to form these basalts must have had an Mg# in excess of 0.60. If the parent is derived from the melting of LMO cumulates, and modal or nearly modal melting is assumed, then cumulates above approximately 85 PCS must be ruled out, as the Mg# of the cumulate is too low. Addition of a small proportion of trapped residual LMO liquid (1–5%) with a much lower Mg# would exacerbate the problem.

Other major-element considerations would appear to indicate further discrepancies in such a simple model. In general, none of the mafic minerals contain an appreciable amount of  $Al_2O_3$ . The parental magmas to the high-TiO<sub>2</sub> basalts contain approximately 9 wt% and up to 12 wt%  $Al_2O_3$ . Thus, the source must contain an Al-bearing component. Clinopyroxene may contain an appreciable proportion of Al. However, even if only clinopyroxene were melted from the source, the clinopyroxene would have to exhibit an enrichment in  $Al_2O_3$  well beyond that observed in mare basalts. Therefore, it is essential that a small proportion (1–2 wt%) of plagioclase was entrained into the source and subsequently melted out during mare basalt magma formation.

In order to account for the high-Ti tenor of these mare basalts, ilmenite is required in the source. However, ilmenite does not precipitate from the LMO until late (at 95 PCS), when the Mg# of the mafic cumulate is 0.34 (Fig. 3). As mentioned above, high-Ti mare basalts exhibit a range of Mg#s; the most primitive samples have Mg#s of approximately 0.48 (JERDE et al., 1992). Therefore, any corresponding mafic source must have an Mg# above this value. In our model, the LMO precipitates a mafic solid with an Mg# of 0.79 at 86 PCS. The radial distance between this layer and the ilmenite-bearing layer at 95 PCS is only 20–30 km in a

300–400-km deep magma ocean. If small-scale convective overturn occurred during the formation of the mantle, then a proportion of the ilmenite-containing layer at 95 PCS could have sunken to, and mixed with, the mafic cumulate from the 86 PCS level. An 80:20 mixture of an 86 PCS cumulate with 95 PCS cumulate yields a mafic cumulate with an Mg# of 0.73. The mineralogy of this mafic cumulate is as follows: 46% pigeonite + 43% olivine + 7% clinopyroxene + 3% ilmenite + 1% plagioclase (and a small (1–2%) proportion of interstitial trapped KREEPy liquid from the 95 PCS level). This mafic cumulate is a viable source for the high-Ti basalts.

Major-element considerations also indicate the need for chromite in the source, in order to generate magmas with elevated  $Cr_2O_3$ . Small amounts of chromite can be easily accommodated in the REE modelling due to the overall low abundances in this mineral. However, in order to generate the inordinately high TiO<sub>2</sub> in these basalts, nonmodal melting with a greater proportion of ilmenite melting out of the source is also required. In fact, experimental studies indicate that ilmenite must not be retained in the residue after melting to form the high-Ti mare basalts (GREEN et al., 1975).

### Trace-element Constraints on the High-Ti Source

Assuming that the parental liquids for the high-Ti basalts have trace-element compositions as indicated in Table 2, an upper limit is imposed on the La/Sm of the source. Due to the fact that all plausible mafic minerals have LREE-depleted partition coefficient patterns, the source which must have melted to generate the high-Ti basalts must also have been LREE-depleted. The lower the degree of melting, the greater the LREE depletion required in the source. Furthermore, if these basaltic melts have undergone fractional crystallization prior to extrusion, the REE abundances observed would be elevated relative to the parent, thus allowing higher degrees of melting of the source. Assuming, as a lower limit, that the minerals left in the residue of melting to form the high-Ti parent had a bulk mineral/melt distribution coefficient ( $D$ ) for the REE of  $10^{-4}$ , or effectively zero (i.e., the REE are totally incompatible with the residue), minimum REE abundances in the source residue may be calculated.

This calculation was performed at 1, 3, 5, and 10% melting for both equilibrium and fractional fusion (when  $D$  approaches 0, they are effectively the same). The minimum residue abundances of the REE (represented in Fig. 9a by Sm) for each degree of melting were then plotted along the minimum  $(La/Sm)_n$  line for the source, as determined by the  $(La/Sm)_n$  of the high-Ti basalt parent (Fig. 9a). Abundances of the REE (again represented by Sm) were then calculated for the plausible residual minerals in equilibrium with the LMO liquid at 90 and then 95 PCS (Fig. 10) using the following simple relation:

$$c_{solid} = c_{liquid} * D.$$

Only clinopyroxene has the requisite abundance of Sm as well as the appropriate  $(La/Sm)_n$  at 80 PCS for  $\geq 1\%$  melting. At 90 PCS, orthopyroxene plots just above the 1% melting line (Fig. 9b). At 95 PCS (Fig. 10), clinopyroxene plots well into the field of permissible partial melting residues, orthopyroxene plots near the 3% residue-of-melting line, and pi-

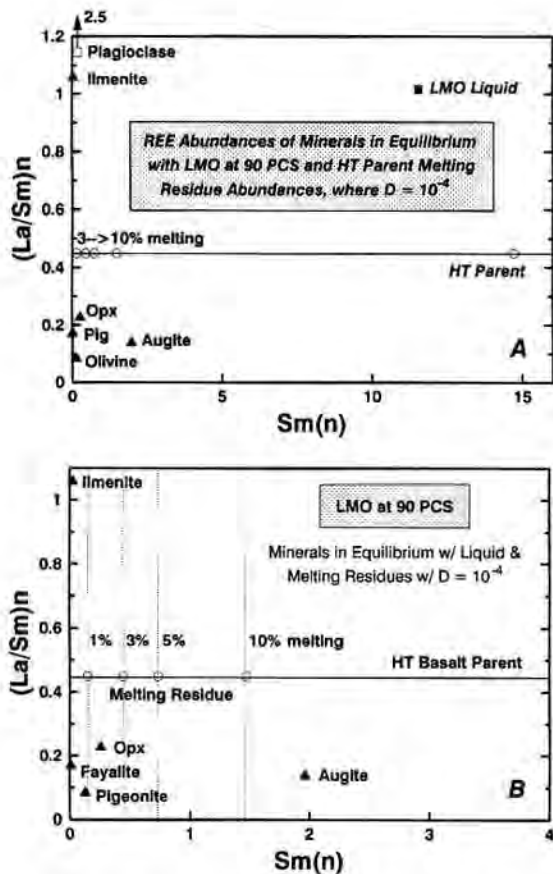


FIG. 9. This series of plots is, in effect, two separate plots superimposed. First, minerals in equilibrium with the calculated LMO liquid at 90 and 95 PCS are plotted. Second, residues of melting were calculated from the actual high-Ti basalt parent. These residue compositions are plotted as open circles on plot (a) but thereafter are indicated by dotted vertical lines. The point of these plots is to determine what degree of melting of a high-Ti source is allowed for different source mineralogies. (a) Plot of  $(La/Sm)_n$  vs.  $Sm(n)$  for residues of melting of a high-Ti basalt parent, assuming  $D = 10^{-4}$  (effectively zero), at 1, 3, 5, and 10% melting (open circles on horizontal line). Essentially these circles give minimum Sm abundances for any source in equilibrium with the high-Ti basalt parent magma. The horizontal line separates permissible sources having  $(La/Sm)_n <$  than that of the high-Ti parent and those above this line which are not permissible. Model minerals (closed triangles) give the composition of the minerals in equilibrium with the LMO liquid at this PCS value (90 PCS). (b) A blow-up of a portion of (a) indicating mineral compositions in equilibrium with the LMO liquid (plotting off the diagram to higher  $Sm(n)$ ) and residue-of-melting lines at 1–10% melting ( $D = 10^{-4}$ ).

geonite plots just above the 1% residue-of-melting line. It must be reemphasized that these model bulk residues are calculated assuming  $D = 10^{-4}$  and therefore must be considered minima. It is likely that pigeonite would actually plot below the 1% residue-of-melting line at reasonable  $D$  values. However, this does not mean that minerals that plot below these minimum source abundance lines cannot occur in the residue, but that their proportion in the residue must be rel-

atively minor. As the proportion of these low-abundance phases is added to the model residue, a greater proportion of some high-abundance phase must be included. This high-abundance phase could be clinopyroxene or the LMO residual liquid. However, because the LMO residual liquid composition lies above the maximum  $(La/Sm)_n$  line (Fig. 9a), it must be a proportionately minor component. At greater degrees of melting, more severe constraints are placed on the mineralogy of the residue, requiring more and more clinopyroxene. It is apparent that either clinopyroxene must have been an important phase in the source, and/or the high-Ti mare basalt source(s) underwent low percentages (1–3%) of melting.

These conclusions are consistent with previous Sr isotopic modelling presented by NYQUIST *et al.* (1976). Furthermore, modelling of Nd and Hf isotopes presented by UNRUH *et al.* (1984) also suggests clinopyroxene in the residue of high-Ti basalt sources (albeit in a much smaller proportion). However, our modelling is not in total agreement with that of NEAL *et al.* (1990) and PACES *et al.* (1991). On the basis of Rb-Sr isotopic systematics for the Apollo 17 Type C basalts, NEAL *et al.* (1990) and PACES *et al.* (1991) have stated that clinopyroxene in the source must have been exhausted during melt generation. Unfortunately, this argument cannot be definitively rebutted due to the uncertainties in lunar-relevant mineral/melt partition coefficients for Rb and Sr. We still contend, based on the trace-element arguments presented above, that a clinopyroxene-bearing residue is required.

A residue from melting the mafic cumulate pile can be calculated by using the bulk mineral/melt partition coefficients for the REE (Table 3). This is shown in Fig. 11 for 1–5% melting with only pigeonite, fayalite, and augite left in the residue. Based on experimental evidence, GREEN *et al.* (1975) have stated that ilmenite should be exhausted from the Apollo 17 Type C high-Ti basalt source. This observation is not inconsistent with the modelling presented herein. In Fig. 11, a field for these residue compositions is shown in reference to the 80:20 mixture of cumulates from the 86 PCS level with those at the 95 PCS level. With the addition of this TIRL component, abundances of this mixed LMO cumulate

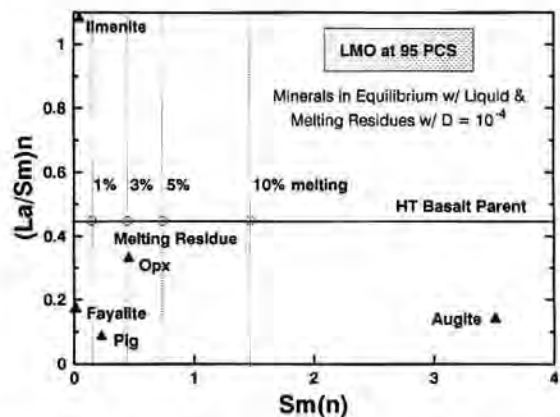


FIG. 10.  $(La/Sm)_n$  vs.  $Sm(n)$  at 95 PCS. Designations are the same as in Fig. 9.

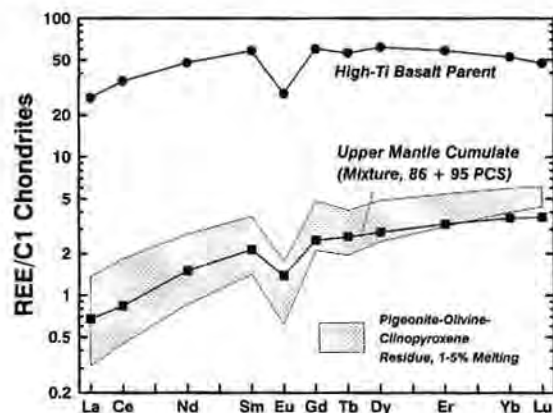


FIG. 11. Plot of the REE (normalized to the initial LMO) for the modeled lunar mantle cumulates (80:20 mixture of 86 PCS:95 PCS) with reference to the high-Ti basalt parent pattern. A field of residues-of-melting for 1–5% melting with pigeonite-olivine-clinopyroxene in the residue and the high-Ti basalt as the derived melt are also indicated.

just achieve the requisite abundances for the residue. Our complete model of mafic cumulates + TIRL + entrained plagioclase can achieve a near-perfect fit to the calculated 1–5% melting residues (Fig. 11), both in terms of the shape of the LREE, HREE, and Eu anomaly, the total REE abundances, and other trace-element abundances.

By varying the percentages of TIRL and entrained plagioclase and the degree of fractional crystallization of the LMO, a wide range of sources which are LREE-depleted to varying degrees and exhibit a negative Eu anomaly may be generated (Fig. 11). Using this scheme, a spectrum of plausible sources for mare basalts may be delineated. Further studies will concentrate on modelling the sources of other mare basalt types.

## DISCUSSION

It is debatable whether the parental magma compositions determined from actual mare basalt suites represent primary magmas. For the modelling presented here, we have assumed that the least-fractionated sample in each group represents not only the parental magma composition for that suite but a primary mantle-derived magma as well. A source residue was then calculated from that composition and then, through successive iterations, an attempt was made to match this residue composition with upper mantle cumulate compositions produced by crystallizing a proposed LMO liquid. If, instead, these hypothesized parental magmas are fractionates of a more primitive melt, then a few conclusions may be quickly drawn. First, the primary melt must have had lower LILE abundances than even the most primitive mare basalt parental composition. Therefore, higher degrees of melting of the source would be allowed. Second, in order to maintain the major-element balance, a larger proportion of plagioclase and a larger proportion of cumulates from a more Mg-enriched layer would be required in the source. The composition of such hypothetical primary magmas is difficult to ascertain. Primitive pyroclastic glasses have already been ruled out as

there is presently no proven connection between them and most mare basalts. BINDER (1982) has speculated that these hypothetical primary melts are related to mare basalts by the simple fractionation of olivine. However, this again assumes that the source for the high-Ti mare basalts is more primitive than what is calculated in this study. An important conclusion of the present study (though not uniquely ours) is that the source region for the high-Ti mare basalts was indeed fractionated relative to other mare basalt sources. This conclusion is in agreement with the high-Ti tenor of these rocks, as ilmenite does not crystallize in the LMO until advanced stages of fractionation.

An olivine-orthopyroxene multiple saturation point was determined at 24 kbars pressure for the Apollo 15 high-Ti (13.8 wt% TiO<sub>2</sub>) red glass (DELANO, 1980, 1986). This observation would suggest that the source of this chilled liquid was located at 450 km depth in the Moon. BINDER (1982) has noted that mare basalt magmas may not have been multiply saturated. Instead, he has suggested that melting was controlled by olivine alone and thus argues for melting at depths shallower than 200 km for all mare basalts. TAYLOR (1982) has also stated that melting at depths greater than 400 km is unlikely. This is due to the fact that at depths greater than this value, mare basalt melts would be in equilibrium with clinopyroxene and garnet. This eclogitic assemblage would have a density which is not consistent with the current understanding of the lunar mantle. Furthermore, melts from a source containing garnet in the residue would not yield melts with trace-element patterns resembling mare basalts. Also, ilmenite does not become a liquidus phase until late in the fractionation sequence of the LMO. The choice of any reasonable bulk LMO composition (such as the four given in Table 1) does not affect this conclusion. Only the substitution of a drastically Ti-enriched (2–3 wt% TiO<sub>2</sub>) initial LMO composition would allow the fractionation of ilmenite early in the sequence (and thus at great depths).

The depth-of-origin problem can be stated succinctly: Experimental data (on high-Ti glasses) suggests derivation from great depths (450 km); modelling of any reasonable LMO composition does not achieve a high-Ti source until late in the sequence (and at shallow depths). This dichotomy has led to the promulgation of "exotic" processes for the formation of high-Ti basalt sources (RINGWOOD and KESSON, 1976; HUGHES et al., 1988, 1989; SHEARER et al., 1991). This process includes massive convective overturn of the lunar mantle, late in its evolution. The overturn would allow late-stage, ilmenite-rich cumulates to be brought down to the depths indicated by the experimental studies of GREEN et al. (1975) and DELANO (1980). Furthermore, this convective overturn would allow late-stage, low-Mg# (ilmenite-rich) cumulates to mix with early, high-Mg# cumulates. The mechanical mixture would then yield a cumulate with an intermediate Mg# and (possibly?) ilmenite on the liquidus, as required for the high-Ti basalt progenitor. However, this mixture would then have REE abundances which would be far too low to generate the high-Ti basalts by partial melting. The addition of a small component of a KREEPy liquid was then proposed to elevate the REE abundances of the source (HUGHES et al., 1988, 1989). Recently, SPERA (1992) has attempted to model magma transport phenomena in the

evolution of the LMO and has subsequently come out in support of the convective overturn hypothesis.

There are several problems with this proposed scenario. First, an unequivocal linkage between basaltic glasses and mare basalts has yet to be made. Therefore, the experimental studies of DELANO (1986), however elegant and eminently applicable to basaltic glass petrogenesis, may have little bearing on the evolution of high-Ti mare basalts. Second, isotopic studies of NYQUIST *et al.* (1976, 1977) can be interpreted as obviating extensive mixing in the sources of mare basalts. Third, there are serious objections to the computer modelling which has been reported recently in support of this scenario (SPERA, 1992). The initial conditions of this modelling assume a semicrystalline magma ocean which consists of two layers: a lower, thicker, less dense (Mg-, olivine-, and opx-rich) cumulate overlain by a denser (high-Ti, high-Fe, ilmenite-rich) cumulate. In this scenario, convective overturn would occur. Though intriguing, this modelling is overly simplistic (indeed, SPERA (1992) states this drawback) and possibly misleading.

Perhaps the most poignant objection to the convective overturn hypothesis is that it is unlikely that this situation would ever occur. As the LMO crystallizes and ilmenite comes on the liquidus, small-scale instabilities will develop; and overturn will occur locally. Cycles of ilmenite-rich accumulation, followed by local convective overturn, could operate repeatedly while ilmenite remains a liquidus phase. Extremely thick ilmenite-rich layers (comprising the upper 10% of the crystallized magma ocean, as per SPERA, 1992) would never develop due to this local and persistent density feedback mechanism. Furthermore, localized small-scale overturn events would likely not affect the gross, primary stratigraphy of the Moon, especially since ilmenite does not become a liquidus phase until very late (>93–95 PCS) in the crystallization of the LMO. The effect of these local density feedback cycles would be to smear out the fine-scale structure of the last 10% of LMO crystallization. However, if convective overturn of the total LMO, late in its crystallization, is proven to be a viable alternative, we would contend that this overturn has not completely obliterated the chemistry of these distinct layers or sources. Instead, these sources could have remained largely intact, albeit displaced to different levels in the LMO.

## CONCLUSIONS

Mantle cumulates produced after approximately 95–96% fractional crystallization of the LMO, and which contain 1–5% trapped instantaneous residual liquid and 2–5% entrained plagioclase, have the requisite REE abundances and relative proportions to generate magmas parental to the high-Ti mare basalts (Fig. 11). As a lower limit, 1–10% batch melting of the LMO mafic cumulates is assumed. Higher degrees of batch melting may indeed occur but would require even higher percentages of fractional crystallization of the LMO in order to generate the appropriate source. The degree of fractional crystallization required may vary somewhat, dependent upon the amount of trapped instantaneous residual liquid in the cumulate. Due to the relative proportions of the REE as calculated in the high-Ti mare basalt source (*i.e.*, LREE-depleted with nearly flat HREE), one constraint does seem to be in-

escapable: The LMO liquid must have reached a point in its crystallization where clinopyroxene was on the liquidus [as also pointed out by NYQUIST *et al.* (1976) and PACES *et al.* (1991)]. Major-element considerations also point to the need for ilmenite (and chromite?) and plagioclase (entrained in the sinking mafic cumulate mush) in the source. Furthermore, in order for the source to retain both the LREE-depleted signature and the negative Eu anomaly, the trapped-liquid component and percentage of plagioclase must be proportionately small (as the TIRL is flat to LREE-enriched, and plagioclase exhibits a large positive Eu anomaly).

Extensive equilibrium crystallization of the LMO will not alter its trace-element evolution but will have an important "buffering" effect on the major elements, specifically Mg and Fe. Whether equilibrium crystallization was as prolonged as presented herein is debatable. However, it is plausible that equilibrium crystallization played a major role in the differentiation of the LMO.

The modelling presented in this paper is partially consistent with that performed by HUGHES *et al.* (1989) on Apollo 17 high-Ti orange glasses and Apollo 15 yellow-brown glass, and Apollo 12 mare basalts (HUGHES *et al.*, 1988), respectively. Their work also suggested extensive fractional crystallization of the LMO with subsequent low degrees of remelting of the cumulates to generate these basalts. They also pointed to the need for a trapped late-stage (KREEPy) LMO liquid in the cumulates. Although our work confirms the need for a trapped liquid component in the source, a simple in-situ trapped instantaneous residual liquid rather than a more evolved variant (HUGHES *et al.*, 1989) will suffice for the high-Ti basalts. A more complicated model involving large-scale convective overturn of the LMO cumulate pile and mixing with a late KREEPy liquid (as per HUGHES *et al.*, 1988) is not required. However, in order to achieve a source with an appropriate Mg#, localized mixing in the upper tenth of the upper mantle is required. A proportion of the ilmenite-containing cumulate from the 95 PCS level could have sunken to the 86 PCS level and mixed with the olivine-orthopyroxene cumulate there (*cf.* RINGWOOD and KESSON, 1976). An 80:20 mixture of this 86 PCS cumulate with the 95 PCS cumulate would yield a source with an Mg# of 0.73. The mineralogy of this mafic cumulate is as follows: 46% pigeonite + 43% olivine + 7% clinopyroxene + 3% ilmenite + 1% plagioclase (with 1–2% trapped KREEPy liquid from the 95 PCS level). Nonmodal melting of such a source, whereby most of the liquid, plagioclase, and ilmenite were exhausted, is a likely scenario for generating the high-Ti basalts.

The source region for a group of high-Ti mare basalts has been successfully modelled using our proposed approach. In general, mare basalt source regions can be generated by simple fractional crystallization of the LMO (accompanied by bulk flotation of plagioclase) and settling of the mafic phases with a small amount of trapped instantaneous residual liquid and entrained plagioclase. The model presented herein is chemically and petrologically feasible; however, it awaits further studies to determine whether it is consistent with other physical and dynamic models of the LMO.

*Acknowledgments*—We would like to thank Frank Spera for a preprint of his work on lunar transport phenomena; John Longhi for assistance

in using his program MAGFOX; and many others for insightful comments and discussions at the Mare Basalt Volcanism Workshop, held in Dallas in October, 1990. We would especially like to thank Paul Hess for critical comments and discussions of this work during all stages of its formulation. A careful review by Eric Jerde improved an earlier version of this manuscript. Thoughtful reviews by Scott Hughes, Chip Shearer, Abhijit Basu, and Roman Schmitt improved the quality and focus of the manuscript and are greatly appreciated. This research was supported by NASA grant NAG 9-415 to L. A. Taylor.

Editorial handling: R. A. Schmitt

## REFERENCES

- AGRELL S. O., SCOON J. H., MUIR I. D., LONG J. V. P., MCCONNELL J. D. C. and PECKETT A. (1970) Observations on the chemistry, mineralogy, and petrology of some Apollo 11 lunar samples. *Proc. Apollo 11 Lunar Sci. Conf.*, pp. 93-128.
- ANDERS E. and GREVESSE N. (1989) Abundances of the elements: Meteoritic and solar. *Geochim. Cosmochim. Acta* 53, 197-214.
- ARTH J. G. (1976) Behavior of trace elements during magmatic processes—a summary of theoretical models and their applications. *J. Res. USGS* 4, 41-47.
- BINDER A. B. (1982) The mare basalt magma source region and mare basalt magma genesis. *Proc. 13th Lunar Planet. Sci. Conf.*, pp. A37-A53.
- BIRD M. L. (1971) Distribution of trace elements in olivines and pyroxenes—an experimental study. Unpubl. Ph.D. thesis, Univ. Missouri at Rolla.
- BROPHY J. G. and BASU A. (1990) Europium anomalies in mare basalts as a consequence of mafic cumulate fractionation from an initial lunar magma. *Proc. 20th Lunar Planet. Sci. Conf.*, pp. 25-30.
- BUCK W. R. and TOKSOZ M. N. (1980) The bulk composition of the moon based on geophysical constraints. *Proc. 11th Lunar Planet. Sci. Conf.*, pp. 2043-2058.
- DELANO J. W. (1980) Chemistry and liquidus phase relations of Apollo 15 red glass: Implications for the deep lunar interior. *Proc. 11th Lunar Planet. Sci. Conf.*, pp. 251-288.
- DELANO J. W. (1986) Pristine lunar glasses: Criteria, data, and implications. *Proc. 16th Lunar Planet. Sci. Conf.*, pp. D201-D213.
- DELANO J. W. (1990) Buoyancy-driven melt segregation in the Earth's Moon: I. Numerical results. *Proc. 20th Lunar Planet. Sci. Conf.*, pp. 3-12.
- GREEN D. H., RINGWOOD A. E., HIBBERSON W. O., and WARE N. G. (1975) Experimental petrology of Apollo 17 mare basalts. *Proc. 6th Lunar Sci. Conf.*, pp. 871-893.
- GROMET L. P., HESS P. C., and RUTHERFORD M. J. (1981) An origin for the REE characteristics of KREEP. *Proc. 12th Lunar Planet. Sci. Conf.*, pp. 903-913.
- HANSON G. N. and LANGMUIR C. H. (1978) Modelling of major elements in mantle-melt systems using trace element approaches. *Geochim. Cosmochim. Acta* 42, 725-741.
- HASKIN L. A., ALLEN R. O., HELMKE P. A., PASTER T. P., ANDERSON M. R., KOROTEV R. L., and ZWEIFEL K. A. (1970) Rare earths and other trace elements in Apollo 11 lunar samples. *Proc. Apollo 11 Lunar Sci. Conf.*, pp. 1213-1231.
- HERBERT F. (1980) Time-dependent lunar density models. *Proc. 11th Lunar Planet. Sci. Conf.*, pp. 2015-2030.
- HESS P. C., RUTHERFORD M. J., GUILLEMETTE R. N., RYERSON F. J., and TUCHFELD H. A. (1975) Residual products of fractional crystallization of lunar magmas: An experimental study. *Proc. 6th Lunar Sci. Conf.*, pp. 895-909.
- HUGHES S. S., DELANO J. W., and SCHMITT R. A. (1988) Apollo 15 yellow-brown glass: Chemistry and petrogenetic relations to green volcanic glass and olivine-normative mare basalts. *Geochim. Cosmochim. Acta* 52, 2379-2391.
- HUGHES S. S., DELANO J. W., and SCHMITT R. A. (1989) Petrogenetic modeling of 74220 high-Ti orange volcanic glasses and the Apollo 11 and 17 high-Ti mare basalts. *Proc. 19th Lunar Planet. Sci. Conf.*, pp. 175-188.
- IRVINE T. N. (1980) Magmatic density currents and cumulus processes. *Amer. J. Sci.* 280A, 1-58.
- JERDE E. A., SNYDER G. A., TAYLOR L. A., LIU Y.-G., and SCHMITT R. A. (1992) Chemical composition of Apollo 11 mare basalt rocks: Evidence for sample heterogeneity in 10057. *Lunar Planet. Sci. Conf. XXXIII*, 613-614.
- JONES J. H. and DELANO J. W. (1989) A three-component model for the bulk composition of the moon. *Geochim. Cosmochim. Acta* 53, 513-527.
- LANGMUIR C. H. (1989) Geochemical consequences of *in situ* crystallization. *Nature* 340, 199-205.
- LANGMUIR C. H. and HANSON G. N. (1980) An evaluation of major element heterogeneity in the mantle sources of basalts. *Phil. Trans. Roy. Soc. London A297*, 383-407.
- LINDSTROM D. J. (1976) Experimental study of the partitioning of the transition metals between cpx and coexisting silicate liquids. Unpubl. Ph.D. thesis, Univ. Oregon.
- LONGHI J. (1978) Pyroxene stability and the composition of the lunar magma ocean. *Proc. 9th Lunar Planet. Sci. Conf.*, pp. 285-306.
- LONGHI J. (1980) A model of early lunar differentiation. *Proc. 11th Lunar Planet. Sci. Conf.*, pp. 289-315.
- LONGHI J. (1981) Preliminary modeling of high pressure partial melting: Implications for early lunar differentiation. *Proc. 12th Lunar Planet. Sci. Conf.*, pp. 1001-1018.
- LONGHI J. and ASHWAL L. D. (1985) Two-stage models for lunar and terrestrial anorthosites: Petrogenesis without a magma ocean. *Proc. 15th Lunar Planet. Sci. Conf.*, pp. C571-C584.
- MCCALLUM I. S. and CHARETTE M. P. (1978) Zr and Nb partition coefficients: Implications for the genesis of mare basalts, KREEP, and sea floor basalts. *Geochim. Cosmochim. Acta* 42, 859-869.
- MCKAY G. A. (1986) Crystal/liquid partitioning of REE in basaltic systems: Extreme fractionation of REE in olivine. *Geochim. Cosmochim. Acta* 50, 69-79.
- MCKAY G. A. and WEILL D. F. (1977) KREEP petrogenesis revisited. *Proc. 8th Lunar Sci. Conf.*, pp. 2339-2355.
- MCKAY G. A., WAGSTAFF J., and YANG S.-R. (1986) Clinopyroxene REE distribution coefficients for shergottites: The REE content of the Shergotty melt. *Geochim. Cosmochim. Acta* 50, 927-937.
- MCKAY G. A., WAGSTAFF J., and LE L. (1991) REE distribution coefficients for pigeonite: Constraints on the origin of the mare basalt europium anomaly, III (ext. abstr.). In *Mare Volcanism & Basalt Petrogenesis Workshop*: LPI Tech. Rept. 91-03, pp. 27-28.
- MORGAN J. W., HERTOGEN J., and ANDERS E. (1978) The Moon: Composition determined by nebular processes. *The Moon and the Planets* 18, 465-478.
- MORSE S. A. (1979) Kiglapait geochemistry: I. Systematics, sampling, and density. *J. Petrol.* 20, 555-590.
- MUELLER S., TAYLOR G. J., and PHILLIPS R. J. (1988) Lunar composition: A geophysical and petrological synthesis. *J. Geophys. Res.* 93, 6338-6352.
- NAKAMURA Y., FUJIMAKI H., NAKAMURA N., TATSUMOTO M., MCKAY G., and WAGSTAFF J. (1986) Hf, Zr, and REE partition coefficients between ilmenite and liquid: Implications for lunar petrogenesis. *Proc. 16th Lunar Planet. Sci. Conf.*, pp. D239-D250.
- NEAL C. R., TAYLOR L. A., HUGHES S. S., and SCHMITT R. A. (1990) The significance of fractional crystallization in the petrogenesis of Apollo 17 Type A and B high-Ti basalts. *Geochim. Cosmochim. Acta* 54, 1817-1833.
- NEAL C. R., TAYLOR L. A., SCHMITT R. A., HUGHES S. S., and LINDSTROM M. M. (1989) High alumina (HA) and very high potassium (VHK) basalt clasts from Apollo 14 breccias. Part 2—whole rock geochemistry: Further evidence for combined assimilation and fractional crystallization within the lunar crust. *Proc. 19th Lunar Planet. Sci. Conf.*, pp. 147-161.
- NYQUIST L. E., BANSAL B. M., and WIESMANN H. (1976) Sr isotopic constraints on the petrogenesis of Apollo 17 mare basalts. *Proc. 7th Lunar Sci. Conf.*, pp. 1507-1528.
- NYQUIST L. E., BANSAL B. M., WOODEN J. L., and WIESMANN H. (1977) Sr-isotopic constraints on the petrogenesis of Apollo 12 mare basalts. *Proc. 8th Lunar Sci. Conf.*, pp. 1383-1415.
- O'NEILL H. ST. C. (1991) The origin of the Moon and the early

- history of the Earth—A chemical model. Part I. The Moon. *Geochim. Cosmochim. Acta* 55, 1135–1157.
- PACES J. B., NAKAI S., NEAL C. R., TAYLOR L. A., HALLIDAY A. N., and LEE D.-C. (1991) A strontium and neodymium isotopic study of Apollo 17 high-Ti mare basalts: Resolution of ages, evolution of magmas, and origins of source heterogeneities. *Geochim. Cosmochim. Acta* 55, 2025–2043.
- PHINNEY W. C. and MORRISON D. A. (1990) Partition coefficients for calcic plagioclase: Implications for Archean anorthosites. *Geochim. Cosmochim. Acta* 54, 1639–1654.
- RHODES J. M., HUBBARD N. J., WIESMANN H., RODGERS K. V., BRANNON J. C., and BANSAL B. M. (1976) Chemistry, classification, and petrogenesis of Apollo 17 mare basalts. *Proc. 7th Lunar Sci. Conf.*, pp. 1467–1489.
- RINGWOOD A. E. and KESSON S. E. (1976) A dynamic model for mare basalt petrogenesis. *Proc. 7th Lunar Sci. Conf.*, pp. 1697–1722.
- RINGWOOD A. E., SEIFERT S., and WANKE H. (1987) A komatiite component in Apollo 16 highland breccias: Implications for the nickel-cobalt systematics and bulk composition of the Moon. *Earth Planet. Sci. Lett.* 81, 105–117.
- ROEDER P. L. and EMSLIE R. F. (1970) Olivine-liquid equilibrium. *Contrib. Mineral. Petrol.* 29, 275–289.
- RYDER G. (1991) Lunar ferroan anorthosite and mare basalt sources: The mixed connection. *Geophys. Res. Lett.* 18, 2065–2068.
- SHAFFER E., BROPHY J. G., and BASU A. (1991) La/Sm ratios in mare basalts as a consequence of mafic cumulate fractionation from an initial lunar magma. *Proc. 21st Lunar Planet. Sci. Conf.*, pp. 325–330.
- SHEARER C. K. and PAPIKE J. J. (1989) Is plagioclase removal responsible for the negative Eu anomaly in the source regions of mare basalts. *Geochim. Cosmochim. Acta* 53, 3331–3336.
- SHEARER C. K., PAPIKE J. J., GALBREATH K. C., and SHIMIZU N. (1991) Exploring the lunar mantle with secondary ion mass spectrometry: A comparison of lunar picritic glass beads from the Apollo 14 and Apollo 17 sites. *Earth Planet. Sci. Lett.* 102, 134–147.
- SHIH C.-Y. and SCHONFELD E. (1976) Mare basalt genesis: A cumulate-remelting model. *Proc. 7th Lunar Sci. Conf.*, pp. 1757–1792.
- SHIMIZU N. (1974) An experimental study of the partitioning of K, Rb, Cs, Sr, and Ba between clinopyroxene and liquid at high pressure. *Geochim. Cosmochim. Acta* 38, 1789–1798.
- SHIRLEY D. N. (1983) A partially molten magma ocean model. *Proc. 13th Lunar Planet. Sci. Conf.*, pp. A519–A527.
- SMITH J. V., ANDERSON A. T., NEWTON R. C., OLSEN E. J., WYLLIE P. J., CREWE A. V., ISAACSON M. S., and JOHNSON D. (1970) Petrologic history of the moon inferred from petrography, mineralogy, and petrogenesis of Apollo 11 rocks. *Proc. Apollo 11 Lunar Sci. Conf.*, pp. 897–925.
- SNYDER G. A. and TAYLOR L. A. (1992) Imperfect fractional crystallization of the Lunar Magma Ocean and formation of the lunar mantle: A "realistic" chemical approach. In *Workshop on the Physics and Chemistry of Magma Oceans from a Bar to 4 Mbar* (ed. C. AGEE and J. LONGHI): LPI Tech. Rept. pp. 41–42.
- SNYDER G. A., TAYLOR L. A., and NEAL C. R. (1990a) Geochemical constraints (and pitfalls) on remelting of lunar magma ocean cumulates for the generation of high-Ti mare basalts (ext. abstr.). In *Mare Volcanism & Basalt Petrogenesis Workshop*: LPI Tech. Rept. 91-03, pp. 51–52.
- SNYDER G. A., TAYLOR L. A., and NEAL C. R. (1990b) The sources of mare basalts: A model involving lunar magma ocean crystallization, plagioclase flotation, and trapped instantaneous residual liquid (ext. abstr.). In *Mare Volcanism & Basalt Petrogenesis Workshop*: LPI Tech. Rept. 91-03, pp. 53–54.
- SNYDER G. A., TAYLOR L. A., and NEAL C. R. (1991a) Determining parental magmas for mare basalts: An interim proposal and synthesis (ext. abstr.). *Lunar Planet. Sci. XXII*, 1295–1296.
- SNYDER G. A., TAYLOR L. A., and NEAL C. R. (1991b) The sources of mare basalts revisited: A model involving lunar magma ocean crystallization, plagioclase entrainment, and trapped instantaneous liquid. *Lunar Planet. Sci. XXII*, 1297–1298.
- SNYDER G. A., TAYLOR L. A., and NEAL C. R. (1992) Combined equilibrium and fractional crystallization of a magma ocean and formation of the upper mantle of the Moon. *Lunar Planet. Sci. XXIII*, 1325–1326.
- SPERA F. J. (1992) Lunar magma transport phenomena. *Geochim. Cosmochim. Acta* 56, 2253–2265.
- TAYLOR S. R. (1982) *Planetary Science: A Lunar Perspective*. Lunar Planet. Inst.
- TAYLOR S. R. (1987) The origin of the Moon. *Amer. Sci.* 75, 469–477.
- TAYLOR S. R. and BENCE A. E. (1975) Evolution of the lunar highland crust. *Proc. 6th Lunar Sci. Conf.*, pp. 1121–1141.
- UNRUH D. M., STILLE P., PATCHETT P. J., and TATSUMOTO M. (1984) Lu-Hf and Sm-Nd evolution in lunar mare basalts. *Proc. 14th Lunar Planet. Sci. Conf.*, pp. B459–B477.
- VILLEMANT B., JAFREZIC H., JORON J. L., and TREUIL M. (1981) Distribution coefficients of major and trace elements: Fractional crystallization in the alkali basalt series of Chaîne des Puys (Massif Central, France). *Geochim. Cosmochim. Acta* 45, 1997–2016.
- WALKER D. (1983) Lunar and terrestrial crust formation. *Proc. 14th Lunar Planet. Sci. Conf.*, pp. B17–B25.
- WALKER D., LONGHI J., and HAYS J. F. (1975) Differentiation of a very thick magma body and implications for the source regions of mare basalts. *Proc. 6th Lunar Sci. Conf.*, pp. 1103–1120.
- WARNER R. D., TAYLOR G. J., CONRAD G. H., NORTHROP H. R., BARKER S., and KEIL K. (1979) Apollo 17 high-Ti mare basalts: New bulk compositional data, magma types, and petrogenesis. *Proc. 10th Lunar Planet. Sci. Conf.*, pp. 225–247.
- WARREN P. H. (1985) The magma ocean concept and lunar evolution. *Ann. Rev. Earth Planet. Sci.* 13, 201–240.
- WARREN P. H. (1986a) Anorthosite assimilation and the origin of the Mg/Fe-related bimodality of pristine moon rocks: Support for the magmasphere hypothesis. *Proc. 16th Lunar Planet. Sci. Conf.*, pp. D331–D343.
- WARREN P. H. (1986b) The bulk-moon MgO/FeO ratio: A highlands perspective. In *Origin of the Moon* (W. K. HARTMANN et al.) pp. 279–310. Lunar Planet. Inst.
- WARREN P. H. (1989) KREEP: Major element diversity, trace-element uniformity (almost). In *Workshop on Moon in Transition: Apollo 14, KREEP, and Evolved Lunar Rocks*: LPI Tech. Rept. 89-03, pp. 149–153. Lunar Planet. Inst.
- WASSON J. T. and KALLEMEYN G. W. (1988) Compositions of chondrites. *Phil. Trans. Roy. Soc. London* A325, 535–544.
- WEILL D. F. and MCKAY G. (1975) The partitioning of Mg, Fe, Sr, Ce, Sm, Eu, and Yb in lunar igneous systems and a possible origin of KREEP by equilibrium partial melting. *Proc. 6th Lunar Sci. Conf.*, pp. 1143–1158.
- WOOD J. A., DICKEY J. S., MARVIN U. B., and POWELL B. N. (1970) Lunar anorthosites and a geophysical model of the moon. *Proc. Apollo 11 Lunar Sci. Conf.*, pp. 965–988.