The source region and melting mineralogy of high-titanium and low-titanium lunar basalts deduced from Lu-Hf isotope data

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Abstract—Five high-Ti basalts from the Apollo 11 and 17 landing sites have been analyzed for their hafnium isotope composition. These data serve to better constrain the hafnium isotope variation of the Moon’s mantle. Variations in initial \( \epsilon_{\text{Hf}} \) and \( \epsilon_{\text{Nd}} \) values of low- and high-Ti basalts imply that the source region mineral assemblages of these lunar magma types are distinct. Low-Ti basalts have higher initial \( \epsilon_{\text{Hf}} \) values, at a given \( \epsilon_{\text{Nd}} \) value, than high-Ti basalts. The differences in the hafnium and neodymium isotopic composition of low- and high-Ti basalts reflect the fact that the source of low-Ti basalts had a [Lu/Hf] ratio approximately four times greater than its [Sm/Nd] ratio. In contrast, the high-Ti source region had a smaller [Lu/Hf] and [Sm/Nd] ratio. If it is assumed that these basalts are partial melt of the Moon’s cumulate mantle, the differences between low- and high-Ti basalts can only be explained by these magma types being generated from melting sources with distinctly different mineral assemblages. The large Lu/Hf fractionations, relative to Sm/Nd fractionations, of low-Ti basalts can best be produced by an assemblage of olivine and orthopyroxene with trace amount of clinopyroxene that crystallized early in the history of the Lunar Magma Ocean (LMO). The subequal [Lu/Hf] and [Sm/Nd] fractionations of high-Ti basalts can be produced from a variety of ilmenite-bearing mineral assemblages.

Low- and high-Ti basalts have similar Lu/Hf ratios, approximately 0.6 times chondrite. The low Lu/Hf ratios measured for these mare magmas contrast sharply with the high Lu/Hf ratios (greater than chondritic) calculated for their sources from initial \( \epsilon_{\text{Hf}} \) values and an assumed chondritic bulk moon initial \( \epsilon_{\text{Hf}} \) value. The difference between the measured Lu/Hf of a lava, vs. the calculated Lu/Hf of its source, implies that during partial melting, Lu was preferentially retained in the residual source, relative to Hf. In order to explain the extreme fractionation of measured Lu/Hf ratios we suggest mare basalts can best be explained using a polybaric melting model. Initial melting of a garnet-bearing source followed by continued melting in the spinel stability field can produce the required Lu/Hf fractionations and produce a liquid that last equilibrated with a residuum of olivine and orthopyroxene.

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1. INTRODUCTION

Two approaches have been taken to infer the mineralogy of mare basalt source regions. One approach uses experimental studies to determine the pressure at which a mare magma type is multiply saturated (e.g., Longhi, 1992a, and references therein). These experimental studies assume that the pressure of multiple-mineral saturation is the depth at which the magma separated from its residuum and that the multiple-saturation phases represent the residual mineralogy. This experimental approach is useful but there are important limitations. First, if the investigated composition is not a primary magma, the multiple-saturation experiment will only produce a minimum depth estimate. Second, multiple-saturation experiments can only determine which phases last equilibrated with a magma. If melting was polybaric (i.e., occurred over a depth interval), the initial high-pressure history cannot be determined. Finally, source phases that were completely consumed during melting cannot be identified.

A second approach to evaluating the mineralogy of the Moon’s mantle is a forward modeling technique. This method assumes a bulk magma ocean composition and uses phase petrology to determine the sequence and proportion of minerals that would crystallize from the magma ocean (e.g., Taylor, 1978; Hughes et al., 1988; Snyder et al., 1992). The major and trace element compositions of these crystallizing cumulates are then calculated to determine the composition of different cumulate horizons in the Moon’s mantle. Based on comparisons between the composition of different mare basalt magma types and the calculated composition of cumulate layers in the mantle, it is possible to infer the mineralogy and depth of magma generation. Like the experimental method, this forward modeling method has its own pitfalls. The choice of the initial magma composition has an influence on crystallization sequence. Calculated trace element compositions are model-driven and depend on the type of crystallization model that is assumed (e.g., fractional or equilibrium), the degree of mineral-melt separation, and the mineral-melt distribution coefficients that are assumed. Moreover, inferring the cumulate horizon from which a mare magma is generated also requires assumptions to be made concerning the degree of melting, the melting proportions of different phases, and the type of melting that occurred (e.g., fractional vs. equilibrium fusion).

Our approach, based on hafnium isotopes and the unique sensitivity of Lu/Hf ratios to mineralogy, does not involve the same assumptions and uncertainties noted above and is distinct
from the experimental and forward modeling methods. The initial $\epsilon_{464}$ values measured for mare basalts are used to calculate the time-integrated Lu/Hf ratio that is required for the source. The calculated [Lu/Hf]$_{source}$ ratio is then compared to the Lu/Hf ratio that is measured for the basalt; differences between the source and magma Lu/Hf ratios reflect differences in bulk Lu/Hf partition coefficients, and the degree of partial melting. Because large, mineral-specific Lu/Hf fractionations may occur during partial melting, Lu/Hf variations allow inferences of distinctive source mineral assemblages. An analogous approach is not possible, for example, with the Sm/Nd isotope system, because large, mineral-specific Sm/Nd fractionations do not occur during partial melting.

2. Lu-Hf ISOTOPE SYSTEMATICS AND ANALYTICAL METHODS

2.1. The Lu-Hf Isootope System

The Lu-Hf isotope system is based on $\beta$ decay of $^{176}$Lu to $^{176}$Hf (t$_{1/2}$ = 3.57 $\pm$ 0.14 $\times$ 10$^7$; Patchett and Tatsumoto, 1980a; Tatsumoto et al., 1981). Variations in hafnium isotope compositions are expressed in ratio form as $^{176}$Hf/$^{177}$Hf, or in epsilon notation where $\epsilon_{Hf} = (\text{Hf}_{\text{sample}}/\text{Hf}_{\text{CHUR}}) - 1) \times 10^4.$ The $\epsilon_{Hf}$ expression is analogous to that used for the Sm-Nd isotope system. The Sm-Nd and Lu-Hf isotope systems are similar, in that the parent element of each system is typically more compatible during mafic magma production, relative to the daughter element. Residual mantle rocks tend to have high Lu/Hf and Sm/Nd ratios, which over time produce positive $\epsilon_{Hf}$ and $\epsilon_{Nd}$ values, if the Lu/Hf and Sm/Nd ratios are greater than chondrite. In contrast, crustal rocks on the Earth tend to have lower Lu/Hf and Sm/Nd ratios, which over time produce negative $\epsilon_{Hf}$ and $\epsilon_{Nd}$ values.

Temporal variation in the isotopic composition of a sample can be approximated by the relationship:

$$\epsilon_{Hf} = \epsilon_{Hf,0} + \epsilon_{Hf}(T_i - T_f) + \epsilon_{Hf},$$

where $\epsilon_{Hf,0}$ = $[(^{176}\text{Lu})/(^{176}\text{Hf})]_{\text{sample}} - [^{176}\text{Lu}/^{176}\text{Hf}]_{\text{CHUR}}$, $Q_{Hf} = (^{176}\text{Lu}/^{176}\text{Hf})_{\text{CHUR}}$, and $\epsilon_{Hf}$ is the notation defined for the Sm-Nd isotope system (e.g., DePaolo, 1988), and T is units of 10$^6$ years. An important observation is that $Q_{Hf}$ and $\epsilon_{Hf}$ are approximately equal (22.9 b.y.$^{-1}$ and 25.1 b.y.$^{-1}$, respectively). This indicates that the relative position of initial $\epsilon_{Hf}$ and $\epsilon_{Nd}$ values of samples are relatively insensitive to the age of the rocks and largely reflect differences in the Lu/Hf and Sm/Nd ratios of a source region. In other words, the relative placement of reservoirs such as crust, depleted mantle, and enriched mantle are the same on an $\epsilon_{Hf}$-$\epsilon_{Nd}$ diagram regardless of the age of the reservoirs (although their absolute separation in $\epsilon$ units will be largest for the present day), providing a powerful interpretive tool when hafnium and neodymium isotopes are both measured on the same samples.

2.2. Analytical Methods

Compared to the large number of neodymium isotope studies that have been conducted, the number of hafnium isotope studies is relatively small. In large part, this is a result of the formidable analytical challenges that are associated with hafnium isotope analysis, which include lengthy chemical processing to separate Hf from Zr and Ti, and the low ionization efficiency of Hf by conventional thermal-ionization mass spectrometry. Moreover, precise Lu/Hf determinations through isotope dilution, which are essential for determining initial hafnium isotope ratios on old rocks, can be difficult to obtain because of spike equilibration problems or incomplete sample dissolution. Problems associated with determining Lu/Hf ratios by isotope dilution include the large sample sizes required for hafnium isotope analysis (typically > 1 g of whole rock powder), as well as the different solubilities of Lu and Hf in different acids (e.g., Unruh et al., 1984). Lutetium, a lanthanide group element, is soluble in HCl but can form precipitates in the presence of hydrofluoric acid. In contrast, Hf is a high-field-strength element and requires hydrofluoric acid to avoid formation of precipitates; Hf is very insoluble in HCl.

All approaches to avoiding problems of imprecise Lu/Hf ratios caused by precipitates is to spike the sample with a mixed Lu/Hf spike prior to dissolution. Our dissolution procedure is described in Scherer et al. (1997), which involves an initial evaporation of 29% HCl to volatilize much of the Si from the sample, followed by a prolonged (~4 days), high-temperature (190°C) dissolution of the sample using concentrated HF and HNO$_3$ in high-temperature bombs. The acid is then evaporated, and the sample is incubated with distilled H$_2$O to remove the HF in high-temperature bombs by periodically decanting the liquid and attacking remaining precipitates with fresh acid until a clear solution is obtained. Previous studies have avoided fully spiking a sample prior to dissolution, because of the poor precision of sample $^{176}$Hf/$^{177}$Hf ratios obtained when subtracting the spike contribution from the measured isotope ratios. This limitation is largely a function of the difference between the $^{176}$Hf/$^{177}$Hf of typical spikes (e.g., 0.145) and that of naturally-occurring Hf (e.g., 0.150), so that the spike values represent 1 part in 10,000 for the $^{176}$Hf/$^{177}$Hf ratio.

We have avoided the potentially large uncertainties in spike subtraction through use of a $^{178}$Hf spike that has a $^{176}$Hf/$^{177}$Hf ratio which is close to that of present-day chondritic compositions. This spike was made by adding a small amount of pure $^{176}$Hf to a $^{176}$Hf spike, producing a Hf spike that has a $^{176}$Hf/$^{177}$Hf of 0.2831 ± 2 (equal to an $\epsilon_{Hf}$ value of +11) and a $^{176}$Hf/$^{177}$Hf of 61.5. Comparison of a $^{176}$Hf spike from a $^{176}$Hf/$^{177}$Hf of 0.2831 ± 2 (equal to an $\epsilon_{Hf}$ value of +11) and $^{176}$Hf/$^{177}$Hf of 61.5. Comparison of a $^{176}$Hf spike from a $^{176}$Hf/$^{177}$Hf of 0.2831 ± 2 (equal to an $\epsilon_{Hf}$ value of +11) and $^{176}$Hf/$^{177}$Hf of 61.5. Comparison of the resulting $^{176}$Hf/$^{177}$Hf ratios determined from spiked rock-samples to the measured $^{176}$Hf/$^{177}$Hf ratio of unspiked rock-samples indicates that there is no bias between these two sets of measurements (Table 1; and Scherer et al., 1995; 1997). The minor correction for spike subtraction is robust for determining measured hafnium isotope ratios.

The chemical method for separating Hf from Zr and Ti follows the procedures outlined in Baragar et al. (1975). 1. A wet digestion technique to produce $^{176}$Hf ($\sim$90% yield) that is relatively free of Zr and Ti. This technique was slightly modified (Scherer et al., 1995) to produce a better separation of Hf from Ti for these high-Ti rocks. Lutetium was separated by initial bulk separation of the REE using cation-exchange resin and HCl, followed by separation of Lu from other REE, using cation-exchange resin and 2-methacrylic acid. Total procedural blanks were 0.1 ng for Lu and 4 ng for Hf, which are insignificant, and no correction was applied. Sample to blank ratios are greater than 1600 for Hf and greater than 3000 for Lu. Approximately 1 gram of sample was used for each analysis.

Lutetium was loaded onto single Re filaments using phosphoric acid, and a small amount of ion-exchange resin was added to the sample load to diminish any possible oxide signal. Lutetium was analyzed using a three-collector, static-analysis routine on a Micromass Sector 54 mass spectrometer. For four spectrometers, the furnace temperature was 2500°C, a single filament, as opposed to the triple filament assembly used by Patchett and Tatsumoto (1980b), made it easier to volatile any Yb that may be present on the Lu sample load. Isobarsic interferences caused by Yb were constantly monitored using mass 174, and small corrections, typically <0.5%, were made to $^{176}$Lu for any $^{176}$Yb present. As a further safeguard, all $^{176}$Lu/$^{177}$Lu analyses represent the average of three determinations run at progressively higher ion currents (typically $2 \times 10^{-13}$ A, $2 \times 10^{-12}$, and $2 \times 10^{-11}$ A at mass 175). The average ratios measured at different ion currents agree within error to one another. The average $^{175}$Lu/$^{177}$Lu measured for a pooled set of standards and unspiked samples is 37.65 ± 2.5, in agreement with the ratio used by Patchett and Tatsumoto (1980b).

Hafnium was loaded onto single Re filaments with boron acid and a small amount of ion exchange resin and run using a dynamic multicollector routine. Isobaric interferences caused by Lu were monitored at mass 175, but in no case were these interferences significant, indicating that our chemical separation methods completely separate Lu from Hf. All analyses were corrected for mass fractionation to a $^{176}$Hf/$^{177}$Hf ratio of 0.7325. The average value measured for the JMC-475 Hf standard during the course of this study was 0.282124 ± 43 (n = 4; 2-SD). This value is in agreement with the 0.282124 ± 43 reported by Baragar et al. (1987). This Hf isotope composition is incorrect for the bias between the USGS and University of Wisconsin instruments, but we have corrected for this bias in the reported $\epsilon_{Hf}$ values. This correction was done by setting the $\epsilon_{Hf}$ value of JMC-475 equal to
HF isotopic composition of the lunar mantle

Table 1: Lu and Hf concentrations and Hf isotopic composition of high-Ti mare basalts

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lu (ppm)</th>
<th>Hf (ppm)</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{182})Hf/(^{177})Hf</th>
<th>(^{178})Hf/(^{177})Hf</th>
<th>Ce*</th>
<th>(^{176})Hf/(^{177})Hf</th>
<th>Ce*</th>
<th>Ce*</th>
<th>Ce*</th>
<th>Age (Ga)</th>
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<tbody>
<tr>
<td>High-Ti mare basalts</td>
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<tr>
<td>78586.10</td>
<td>1.065</td>
<td>6.998</td>
<td>0.02225</td>
<td>0.282057 ± 0.0294</td>
<td>-25.8</td>
<td>3.5 ± 0.8</td>
<td>2.77</td>
<td>3.69</td>
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<tr>
<td>71543.16</td>
<td>1.039</td>
<td>6.635</td>
<td>0.02268</td>
<td>0.282124 ± 0.02111</td>
<td>-23.5</td>
<td>4.2 ± 0.8</td>
<td>4.61</td>
<td>3.69</td>
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<tr>
<td>77516.22</td>
<td>1.015</td>
<td>6.400</td>
<td>0.02254</td>
<td>0.282020 ± 0.0279</td>
<td>-27.1</td>
<td>1.4 ± 0.8</td>
<td>4.06</td>
<td>3.69</td>
<td></td>
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<tr>
<td>10044.634</td>
<td>2.618</td>
<td>15.857</td>
<td>0.02342</td>
<td>0.282282 ± 0.24</td>
<td>-17.9</td>
<td>8.2 ± 0.9</td>
<td>7.10</td>
<td>3.67</td>
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<tr>
<td>10047.219</td>
<td>2.430</td>
<td>14.664</td>
<td>0.02355</td>
<td>0.282310 ± 0.23</td>
<td>-16.9</td>
<td>8.9 ± 0.8</td>
<td>7.40</td>
<td>3.67</td>
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<tr>
<td>Ferrodiorites, Laramie Anorthosite Complex</td>
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<tr>
<td>PM-316 #1</td>
<td>0.595</td>
<td>4.694</td>
<td>0.01880</td>
<td>0.282339 ± 0.0234</td>
<td>-15.9</td>
<td>0.5 ± 1.2</td>
<td>1.43</td>
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<tr>
<td>PM-316 #2</td>
<td>0.994</td>
<td>4.874</td>
<td>0.01733</td>
<td>0.282353 ± 0.0174</td>
<td>-15.4</td>
<td>0.7 ± 0.6</td>
<td>1.43</td>
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<tr>
<td>SR-322 #1</td>
<td>0.450</td>
<td>5.411</td>
<td>0.01182</td>
<td>0.282189 ± 0.032</td>
<td>-21.2</td>
<td>0.3 ± 1.1</td>
<td>1.43</td>
<td></td>
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<tr>
<td>SR-322 #2</td>
<td>0.649</td>
<td>5.256</td>
<td>0.01214</td>
<td>0.282212 ± 0.027</td>
<td>-20.4</td>
<td>0.8 ± 1.0</td>
<td>1.43</td>
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<tr>
<td>SR-322 #3</td>
<td>0.448</td>
<td>5.500</td>
<td>0.01158</td>
<td>0.282185 ± 0.024</td>
<td>-21.3</td>
<td>0.4 ± 0.9</td>
<td>1.43</td>
<td></td>
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<tr>
<td>SR-322 #4 &amp; 5</td>
<td>unspiked</td>
<td>unspiked</td>
<td>unspiked</td>
<td>0.282183 ± 0.034</td>
<td>-20.8</td>
<td>0.282167 ± 0.20</td>
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a-age and initial \(^{143}\)Nd values from Paces et al. (1991)
b-age and initial \(^{143}\)Nd values from Snyder et al. (1994)

Ferrodiortites from the Laramie Anorthosite complex have Ti/Zr and Th/Hf ratios that are similar to those of high-Ti basalts (e.g., Mitchell et al., 1996). We used these samples as proxies for high-Ti basalts to evaluate the external reproducibility of initial \(^{143}\)Nd values and to assess whether spike-subtracted \(^{182}\)Hf/\(^{177}\)Hf ratios are biased relative to unspiked samples. Based on the analyses of the ferrodiorites, as well as analyses of spiked and unspiked normals, we believe that the spike subtraction is robust. The external reproducibility of initial \(^{143}\)Nd values of the ferrodiorites is ±0.6, and is typically better than the internal precision as calculated from 2-SE in-run statistics. The external reproducibility of Lu/Hf ratios is difficult to estimate because the large variation in the Lu/Hf ratio of sample SR-322 is positively correlated with \(^{182}\)Hf/\(^{177}\)Hf ratios, suggesting heterogeneity in different splits of bulk-rock powder. It has been suggested that Samples 10044 and 10047 are from the same sample (Beatty and Albare, 1997). The spread in initial \(^{143}\)Nd values of these samples is 0.7 epsilon units.

-23.47 (Patchett and Tatsumoto, 1980a; Tatsumoto et al., 1981); this is equivalent to a present-day \(^{182}\)Hf/\(^{177}\)Hf ratio for CHUR of 0.282788 for the University of Wisconsin laboratory.

3. PREVIOUS GEOCHEMICAL AND EXPERIMENTAL RESULTS FOR LOW- AND HIGH-TITANIUM BASALTS

3.1. General Features of the Lunar Mantle

Most theories regarding the evolution of the Moon suggest that soon after accretion, the outer ~4000km of the moon was completely melted, forming a magma ocean. As this magma ocean (LMO) crystallized, dense mafic phases sank and lighter plagioclase floated to the surface. The end result of crystallization of the LMO was formation of a differentiated planet, consisting of a plagioclase-rich crust and an upper mantle composed of mafic-mineral cumulates (Smith et al., 1970; Wood et al., 1970). Bottom layers in the stratified cumulate mantle consisted of the first phases to crystallize from the LMO, whereas the upper part of the mantle consisted of phases that crystallized at lower temperatures. The predicted mineralogy of the cumulate layers is model driven and depends on the assumed bulk composition of the LMO and if crystallization of the LMO was fractional or equilibrium, or took place at high or low pressures, or both. However, most petrologic investigations suggest that the lower cumulate mantle consists of olivine and orthopyroxene, whereas the upper cumulate mantle also includes pigeonite and clinopyroxene (e.g., Longhi, 1981). Late in the crystallization of the LMO, ilmenite was a liquidus phase, which resulted in production of a Ti-rich upper mantle. Importantly, some workers have noted that the ilmenite-rich horizon would not be in density equilibrium and may have sunk into the lunar mantle to its density compensation depth (e.g., Ringwood and Kesson, 1976; Spera, 1992). However, it is largely unknown if sinking of the ilmenite horizon was large scale and completely stirred and mixed the cumulate mantle, or if such effects only resulted in localized disruption of the cumulate layers of the Moon’s mantle.

Using the LMO model, it is possible to make general predictions regarding the composition of the Moon’s cumulate mantle. Early-formed cumulates will have lower incompatible-element contents relative to later formed cumulates, reflecting the progressive enrichment of incompatible-element contents in the evolving LMO liquid. In contrast, the Mg number and compatible element contents of early formed cumulates will be higher than in later formed cumulates (until ilmenite begins to crystallize), reflecting the progressive depletion of compatible elements in an evolving LMO.

3.2. The Origins of Mare Basalts

The first models developed for explaining the compositional diversity of mare basalts suggested that low-Ti basalts were derived from deep olivine and pyroxene cumulates and that high-Ti basalts were derived from upper cumulates composed of olivine, pyroxenes, and ilmenite (e.g., Green et al., 1971; Longhi et al., 1974; Walker et al., 1975). This model provided an explanation for the high incompatible element contents of high-Ti basalts compared to low-Ti basalts, because later formed cumulates have higher incompatible element contents than earlier formed contents. Moreover, the pressures determined by multiple saturation experiments on low- and high-Ti basalts agree with forward models of the crystallization of the LMO which predicted that ilmenite bearing cumulates would have formed at 6 kbars (approximately 120 km on the Moon), and olivine low-Ca pyroxene cumulates would occur at 12–20 kbars (200–400 km on the Moon).

Critical evaluation of the cumulate remelting model left
unexplained several important features. For example, remelting of early- vs. late-formed cumulates would predict that low-Ti basalts should have higher Mg numbers and compatible trace element contents as compared to high-Ti basalts; low- and high-Ti basalts have similar Mg numbers and Ni and Cr contents (e.g., Ringwood and Kesson, 1976). Moreover, additional multiple saturation experiments on compositions that more closely represent primary magmas indicate that both low- and high-Ti basalts were derived from an olivine + low-Ca pyroxene source at similar pressures (15–25; Green et al., 1975; Walker et al., 1975; Delano, 1980). In order to explain the similar equilibration pressures and Mg numbers of low- and high-Ti basalts, Ringwood and Kesson (1976) proposed that the dense upper ilmenite layer of the cumulate mantle sank into the mantle, partially mixing with the deeper high-Mg cumulates. By placing the ilmenite-bearing cumulates into the deeper reaches of the Moon’s mantle, it is possible to explain the similar depths of multiple saturation of low- and high-Ti basalts. Partial mixing of the low- and high-Ti basaltic source regions also explains the similar Mg numbers of these more basaltic magmas. Other workers proposed that high-Ti basalts were derived from low-Ti basaltic magmas that had selectively assimilated the upper ilmenite layer (Anderson, 1971; Hubbard and Minear, 1975).

The major difficulty of the Ringwood and Kesson (1976) model is that theoretical calculations and experimental results indicate that a high-Ti magma at pressures greater than 20 kbars is more dense than their surrounding crystalline mass (Delano, 1990; Agee and Circone, 1995). The high densities of high-Ti magmas would suggest that magmas derived from melting an ilmenite-bearing mantle should sink and never be erupted onto the lunar surface. However, Hess (1991) suggested that this density problem may be overcome by entrainment of additional material in the rising mantle diapir.

The ilmenite assimilation model had been criticized in the past because it was not believed that low-Ti basaltic magmas had the required thermal energy to assimilate ilmenite (e.g., Ringwood and Kesson, 1976). However, recent experimental results have clearly shown that the thermal budget of low-Ti basalts is sufficient to assimilate ilmenite (Wagner and Grove, 1997).

4. NEODYMIUM AND HAFNIUM ISOTOPE DATA

The lunar samples analyzed in this study serve to complement the more extensive studies of Patchett and Tatsumoto (1981) and Unruh et al. (1984). Specifically, we selected high-Ti basalts from the Apollo 11 and 17 landing sites which have neodymium isotope compositions that differ from those of the Apollo 11 and 17 samples analyzed by Unruh et al. (1984). Two Apollo 11 samples that have high initial εNd values were analyzed to compare with the hafnium isotope compositions of Apollo 11 samples that have low εNd values (analyzed by Unruh et al., 1984). Three Apollo 17 samples that have low εNd values were analyzed to compare with the high εNd samples analyzed by Unruh et al. (1984). These additional data serve to better define the variations in εHf and εNd values for high-Ti basalts.

The five analyzed samples are considered to be pristine based on petrography and chemical composition. Samples 10044 and 10047 are type B1 basalts (e.g., Jerde et al., 1994) and petrographic descriptions are reported in Beatty and Albee (1978). Samples 78586, 71545, and 77516 are type B2 basalts (e.g., Neal et al., 1990) and petrographic descriptions are reported in Warner et al. (1975) and Neal and Taylor (1993).

The low-Ti basalts from the Apollo 12 and 15 landing sites define a large range of initial εHf values (+8 to +46) and εNd values are positively correlated (Fig. 1, data from Patchett and Tatsumoto, 1981 and Unruh et al., 1984). The Apollo 12 ilmenite basalts have the highest εNd and εNd values, +46 and +10, respectively. A linear regression of all low-Ti samples produces the equation: εHf = 3.79εNd + 5.48 (r² = 0.97). It remains unclear why this array does not pass through the origin, but Unruh et al. (1984) suggest it may be a result of underestimating the initial 176Hf/177Hf of CHUR, or that the Lu/Hf and Sm/Nd ratios of the low-Ti basalt sources were influenced by different minerals.

In contrast to the wide range in εHf variations of low-Ti basalts, high-Ti basalts have a much smaller range in εHf values (+10 to +13), even though their range in εNd values is similar (Fig. 1). High-K, high-Ti basalts have higher εHf values at a given εNd value compared to low-K, high-Ti basalts. Low-K, high-Ti basalts define a positive correlation of initial εHf and εNd values; a linear regression of all the analyzed low-K, high-Ti basalts produces the equation of εHf = 0.95εNd + 1.00 (r² = 0.62). Relative to low-Ti basalts, the slope of the εHf and εNd correlation line is very shallow. For comparison, the slope of the εNd and εNd variations of the terrestrial mantle (defined by oceanic island basalts) is intermediate at 1.36 (Johnson and Beard, 1993; Fig. 1).

The differences between the initial εHf and εNd values of low-Ti and high-Ti basalts are not a result of differences in the ages of the reservoirs from which these samples were derived because the expected rates of change in εNd and εNd values are approximately equal. This fortuitous property of the Sm-Nd and Lu-Hf isotope systems is a result of the trade-off between the low isotopic abundance of 176Lu and its short half-life, relative to the higher isotopic abundance of 146Sm and its longer half-life, which results in QHf ~ QNd, as noted above. Therefore, variations in the slope of initial εNd and εNd values are a result of long-term differences in the Lu/Hf and Sm/Nd ratios of their source. The low-Ti source region is characterized by Lu/Hf fractionations that are approximately four times greater than those of Sm/Nd fractionations (Fig. 2) as inferred by their εHf and εNd relations. In contrast, the high-Ti source region is characterized by subequal fractionations of Lu/Hf and Sm/Nd (Fig. 2).

5. DISTRIBUTION COEFFICIENTS

The combined isotopic and trace element melting models that we present below are dependent on the mineral-melt partition coefficients that are used. Therefore, we briefly review our rationale for choosing the partition coefficients used in this study. The critical parameters for our melting models are the bulk D-value ratios of Lu/Hf and Sm/Nd (K(D,Lu/Hf) and K(D,Sm/Nd)). We have assembled a database of published partition coefficients determined by either experimental methods, or by analysis of phenocryst-matrix pairs, that report Lu and Hf or Sm and Nd distribution coefficients for phases that are likely to
have been important in lunar petrogenesis, including olivine, orthopyroxene, clinopyroxene, pigeonite, ilmenite, plagioclase, garnet, and apatite. We have limited this database to studies that were conducted on basaltic- to intermediate-rock compositions and use this to determine the median D$_{Lu}$, D$_{Sm}$, K$_{DJ(Lu/Hf)}$, and K$_{DJ(Sm/Nd)}$ values; the D$_{Hf}$ and D$_{Nd}$ values were adjusted to match the median K$_{DJ(Lu/Hf)}$ and K$_{DJ(Sm/Nd)}$. Importantly, because our melting calculations are only dependent on elemental ratios, uncertainties introduced by variable absolute values for partition coefficients due to changes in pressure, temperature, and bulk composition will be minimized. This minimization is due to the fact that although absolute partition coefficients are dependent upon compositional, temperature, and pressure effects, the absolute values for D$_{Lu}$, D$_{Hf}$, D$_{Sm}$, and D$_{Nd}$ generally change in the same direction. Hence, models that involve trace element ratios, or isotopic variations, which also reflect variations in trace element ratios, will not be as strongly affected as those that are based upon absolute concentrations.

Figure 3 shows the range of measured K$_{DJ(Sm/Nd)}$ and K$_{DJ(Lu/Hf)}$ values for different phases, the median K$_{DJ}$ values used in this study, and the K$_{DJ}$ values used by Unruh et al. (1984) in their study of the neodymium and hafnium isotope composition of mare basalts. All of the investigated phases have K$_{DJ(Lu/Hf)}$ > 1, except ilmenite, which has a K$_{DJ(Lu/Hf)}$ < 1. However, K$_{DJ(Sm/Nd)}$ of ilmenite is similar to most of the other silicate phases, which implies that ilmenite fractionation may produce large Lu/Hf fractionations without perturbing the Sm/Nd system. Another important phase that can produce large variations in the hafnium isotope system is garnet (e.g., Salters and Hart, 1989; Beard and Johnson, 1993). Garnet is one of the few phases in which Lu is a compatible element (D$_{Lu}$ > 1), making its K$_{DJ(Lu/Hf)}$ > 10. We do not show a field of K$_{DJ}$-values for pigeonite because there are few investigations that have determined the D$_{Lu}$ and D$_{Hf}$ values of pigeonite. However, based on crystal chemical grounds it is believed that pigeonite partition coefficients should have values that are intermediate to those of clinopyroxene and orthopyroxene (e.g., Shearer et al., 1989). We have used the K$_{DJ(Lu/Hf)}$ for pigeonite determined by Haskin and Korotev (1977), which is a value that is intermediate relative to orthopyroxene and clinopyroxene. Also a K$_{DJ(Sm/Nd)}$ = 1.9 for pigeonite was used, as determined by the studies of McKay et al. (1991). A field for the range of apatite K$_{DJ}$-values is not shown in Fig. 3 because there are few investigations that have determined the D$_{Hf}$ in apatite. Apatite-melt partitioning of REE has been studied (e.g., Watson and Green, 1981; McKay et al., 1992), and we have used the REE D-values determined by Watson and
Green (1981) for a basanite liquid. We have used a $D_{\text{HF}}$ of 0.15 which is consistent with the $K_{D_{\text{Lu/Hf}}}$ determined by Fujimaki (1986).

The pioneering hafnium isotope studies of lunar materials by Patchett and Tatsumoto (1981) and Unruh et al. (1984) were conducted prior to most of the investigations that determined $D_{\text{Lu}}$ and $D_{\text{Hf}}$ values. As such, the partition coefficients used by Unruh et al. (1984) are based mainly on the mineral-matrix studies conducted by Fujimaki et al. (1984). For orthopyroxene and ilmenite, the values determined by Fujimaki et al. (1984) are among the most extreme $K_{D_{\text{Lu/Hf}}}$ values determined. For example, the $K_{D_{\text{Lu/Hf}}}$ for orthopyroxene used by Unruh et al. (1984) is 25; this value is similar to the $K_{D_{\text{Lu/Hf}}}$ of garnet. In contrast to this extreme orthopyroxene $K_{D_{\text{Lu/Hf}}}$ value, most recent investigations have determined that $K_{D_{\text{Lu/Hf}}}$ is significantly smaller; the median $K_{D_{\text{Lu/Hf}}}$ is 5, and all values, excluding Fujimaki et al. (1984), are less than 10. The $K_{D_{\text{Lu/Hf}}}$ for ilmenite used by Unruh et al. (1984) is also an extreme value, relative to other determinations of $K_{D_{\text{Lu/Hf}}}$ for ilmenite. Fujimaki et al. (1984) determined that Hf is a compatible element in ilmenite ($D_{\text{Hf}} = 1.8$), which makes this determination the lowest $K_{D_{\text{Lu/Hf}}}$ determined for ilmenite ($K_{D_{\text{Lu/Hf}}} = 0.03$). More recent experimental results indicated that Hf is a moderately incompatible element in ilmenite ($D_{\text{Hf}} = 0.4$), and that ilmenite $K_{D_{\text{Lu/Hf}}}$ is 0.18 (Nakamura et al., 1986; McKay et al., 1986a).

6. HAFNIIUM ISOTOPE AND LUNAR TRACE ELEMENT MODELS

The following trace element modeling is based largely on our preferred set of partition coefficients for $K_{D_{\text{Lu/Hf}}}$ and $K_{D_{\text{Sm/Nd}}}$, and the assumption that the bulk LMO was initially chondritic in composition. We graphically illustrate in
Fig. 4 shows how the Lu/Hf ratio of a basalt's source region may be calculated for a given initial $\epsilon_{Hf}$ value of the lava, and an estimate of the age and isotopic composition of the original source region. For these calculations we have assumed that the LMO crystallized at 4.4 Ga with an initial chondritic CHUR (chondritic uniform reservoir) isotopic composition ($\epsilon_{Nd} = \epsilon_{Hf} = 0$; e.g., Carlson and Lugmair, 1981). If the LMO crystallized sooner, as suggested by recent geochronologic data on ferroan anorthosites (e.g., Alibert et al., 1994; Lee et al., 1997), the Lu/Hf and Sm/Nd source ratios that we have calculated will be maximum estimates. If the LMO had negative $\epsilon_{Nd}$ and $\epsilon_{Hf}$ values, the calculated Lu/Hf and Sm/Nd source ratios are minimum estimates. For example, for the assumption that the LMO crystallized at 4.55 Ga (rather than at 4.4 Ga), then the Lu/Hf source ratio for the Apollo 12 ilmenite basalt would be 4.5% lower. For the assumption that the LMO had an $\epsilon_{Hf}$ value of $-1.5$ at 4.4 Ga (rather than 0), then the calculated source ratio of Apollo 12 ilmenite basalt would be 2.4% higher. Considering the extreme variations in hafnium isotopic composition measured for low- and high-Ti basalts, the slight imprecision in calculated Lu/Hf source ratios is small. Therefore, these models are relatively insensitive to variations in the age and the assumed isotopic composition of the LMO, within the bounds that have been discussed.

We have modeled the source mineralogy of high- and low-Ti basalts using two methods. Method 1 predicts the possible minerals that could have crystallized from the LMO to produce the Lu/Hf and Sm/Nd ratios of their source as calculated using initial neodymium and hafnium isotope compositions. Method 2 attempts to define the melting mineralogy required to produce the Lu/Hf and Sm/Nd ratios measured in mare basalts, relative to the calculated Lu/Hf and Sm/Nd ratios of their sources.

6.1. Method 1: Source Region Mineralogy Inferred from Lu/Hf and Sm/Nd Source Ratios

The $\epsilon_{Hf}$ and $\epsilon_{Nd}$ variation of low-Ti basalts implies that their source had a [Lu/Hf]$_{m}$ ratio which was four times greater than its [Sm/Nd]$_{m}$ ratio (Figs. 1 and 2). In contrast, the slope of the $\epsilon_{Hf}$-$\epsilon_{Nd}$ correlation of high-Ti basalts implies that high-Ti basalts were derived from a source that had subequal [Lu/Hf]$_{m}$ and [Sm/Nd]$_{m}$ ratios (Figs. 1 and 2). Following the method outlined by Unruh et al. (1984), we calculate possible mineral assemblages for sources that are capable of producing these different Lu/Hf to Sm/Nd fractionations using the simple relationship: $C_{m} = D \times C_{L}$, where $C_{m}$ is the concentration of an element in the cumulate mantle, $C_{L}$ is the concentration of that element in the LMO liquid at the time the cumulates formed, and D is bulk mineral/melt partition coefficient. The bulk D-value is the sum of the products of the distribution coefficient of a phase and the proportion of the phase in the cumulate. The effect of trapped melt on the cumulate composition is modeled by assuming that trapped melt has a distribution coefficient of 1 for all elements.

The concentration of an element in the LMO ($C_{L}$) has been evaluated by calculating the concentrations of Lu, Hf, Sm, and Nd in the LMO using the crystallization sequence determined by Snyder et al. (1992). This trace element modeling indicates that the [Lu/Hf]$_{m}$/[Sm/Nd]$_{m}$ ratio of the LMO will only fractionate to a value of 0.85 at 99.5% crystallization; in the first 90% of crystallization of the LMO, the [Lu/Hf]$_{m}$/[Sm/Nd]$_{m}$ ratio will only decrease to 0.9 (Fig. 5). Crystallization of the LMO decreases both the Sm/Nd and Lu/Hf ratios of the fractionating liquid. Therefore, it is highly unlikely that the large differences between $\epsilon_{Hf}$ and $\epsilon_{Nd}$ correlations of low- and high-Ti basalts are caused solely by changes in the composition of the LMO as it crystallized.
Taylor, 1978), then the source of low-Ti basalts must be dominated by orthopyroxene to explain the \([\text{Lu/Hf}]_n/[\text{Sm/Nd}]_n\) ratios. Crystalization of an assemblage of olivine (ol), orthopyroxene (opx), and minor clinopyroxene (cpx) produces a reasonable fit to the high Lu/Hf and Sm/Nd source ratios calculated for the Apollo 12 ilmenite basalts (Fig. 6). The lower Lu/Hf source ratios of Apollo 12 feldspathic and pigeonite basalts and the Apollo 15 basalts can be reproduced by the same assemblage if a small proportion of trapped melt (0–10%) is included (Fig. 6). Clinopyroxene is required in the source of low-Ti basalts to produce the linear array of Lu/Hf and Sm/Nd source ratios; exclusion of clinopyroxene produces a strongly convex upward array that does not match the linear positive correlation between \([\text{Lu/Hf}]_n\) and \([\text{Sm/Nd}]_n\) source ratios of low-Ti basalts.

6.1.2. High-titanium basalts

Inferences regarding the mineralogy of the high-Ti basal source region are more difficult to make because there are a variety of possible mineral assemblages that can produce a source that has a low \([\text{Lu/Hf}]_n/[\text{Sm/Nd}]_n\) ratio. This is largely a reflection of the fact that ilmenite has a \(K_{\text{D}[\text{Sm/Nd}]} < 1\), but a \(K_{\text{D}[\text{Sm/Nd}]}\) that is similar to most silicate phases. Therefore, a variety of phases that produce large \([\text{Lu/Hf}]/[\text{Sm/Nd}]\) fractionations (e.g., orthopyroxene or pigeonite) can be in the source of high-Ti basalts; if ilmenite is included to buffer the Lu/Hf fractionation to low values in the Moon’s upper mantle cumulate pile. If we exclude ilmenite (ilm) as a possible phase, then the source of high-Ti basalts must be dominated by clinopyroxene, because it does not strongly fractionate Lu/Hf relative to Sm/Nd.

A cumulate layer consisting of ol + opx + cpx + ilm ± trapped melt (0–6%) that crystallized from the LMO after >95% of the magma ocean had crystallized can produce the shallow slope of \([\text{Lu/Hf}]_n\) to [Sm/Nd] source ratios of high-Ti basalts. (Fig. 6). Assuming a large degree of crystallization of the LMO is consistent with ilmenite being a liquidus phase late in the crystallization history of the LMO (e.g., Snyder et al., 1992). Other possibilities such as pigeonite replacing opx in the high-Ti basal source are equally consistent with Lu/Hf and Sm/Nd source ratio data. However, we conclude that a small proportion of ilmenite must be in the source region of high-Ti basalts to buffer the \([\text{Lu/Hf}]_n\) source ratio to low values without affecting [Sm/Nd] source ratios.

6.1.3. Hybridization models

It has recently been shown that the exceptionally large variation in Ti contents (0.3–16 wt% TiO₂) of picritic lunar glasses can be produced by assimilation of ilmenite by low-Ti (1 wt%) picritic magmas derived from an ol + opx source (e.g., Wagner and Grove, 1997). Variations in the initial \(\epsilon_{\text{Hf}}\) and \(\epsilon_{\text{Nd}}\) values of low and high-Ti basalts can be used to evaluate this hypothesis. Based on the \(K_{\text{D}[\text{Hf/Hf}]}\) and \(K_{\text{D}[\text{Sm/Nd}]}\) for ilmenite, one would predict that it will have a low Lu/Hf ratio (resulting in low \(\epsilon_{\text{Hf}}\) values over time) and a Sm/Nd ratio similar to other silicate phases such as clinopyroxene (resulting in positive \(\epsilon_{\text{Nd}}\) values over time). Calculated trends for bulk mixing between a parental low-Ti picritic magma (Apollo 15 green glass) and an
ilmenite component are shown in Fig. 7a. Assimilation of ilmenite can produce the high Ti contents and low Lu/Hf source ratios of high-Ti basalts by 30% assimilation of ilmenite-rich + cpx cumulate. However, such an assimilation process produces near vertical mixing arrays on a plot of Lu/Hf and Sm/Nd source ratios at low (<10%) amounts of ilmenite assimilation (Fig. 7b). In contrast, the [Lu/Hf]_s and [Sm/Nd]_s source ratios of low-Ti basalts defines a distinct linear array that is subnormal to the hypothetical ilmenite assimilation mixing arrays, implying low- and high-Ti basalts were derived from different long-lived sources. Stated in another way, the moderate Ti contents of Apollo 12 ilmenite and pigeonite basalts (6wt% TiO₂) are difficult to reconcile with their initial ε_Hf values of +46 and +20, respectively, by a hybridization model. Small amounts of ilmenite assimilation should produce large variations in initial ε_Hf values making it impossible to preserve the strong positive 4 to 1 correlation between [Lu/Hf]_s and [Sm/Nd]_s source ratios. The hafnium isotope system has the unique ability to resolve the distinct mineralogic differences that existed between the low- and high-Ti basalt source regions.

6.2. Method 2: Melting Mineralogy Inferred from Measured Lu/Hf Lava Data

If the Moon’s upper mantle was formed by crystallization of the LMO, and has since remained static, the source mineralogy calculated from the isotopic data should be identical to the melting mineralogy that is required to produce the observed Sm/Nd and Lu/Hf ratios measured in mare basalts. If the mineralogy required to produce the isotopic composition of the source, and the melting mineralogy required to produce the Sm/Nd and Lu/Hf ratios of mare basalts are different, then we must conclude that the history of the Moon’s upper mantle has been dynamic, involving a change in the source-region mineral assemblage.

The melting mineralogy required to produce low- and high-Ti basaltic magmas can be determined by comparing the measured [Lu/Hf]_s of the lava, to the calculated [Lu/Hf]_s of the source (determined by evolving the source from CHUR to the initial ε_Hf value of the lava), which is essentially the angle between the evolution curves for a particular sample illustrated in Fig. 4. Differences between the measured [Lu/Hf]_s of the lava, relative to the inferred [Lu/Hf]_s of the source, are a result of the degree of partial melting and bulk K_D(Lu/Hf). The assumptions and limitations of using the Lu/Hf isotope system to calculate the melting mineralogy of a source region are discussed in Beard and Johnson (1993) and include assumptions regarding the age and isotopic composition of the reservoir from which a source was derived, recent mantle metasomatism, and the effect that low-pressure crystal fractionation has in changing the measured Lu/Hf ratio of the derived lava.

For this study, two important effects that must be evaluated are low-pressure crystal fractionation, and the influence that trapped melt from the initial crystallization of the LMO (i.e., mesostasis in the cumulate mantle) may have on melting mineralogy. The effect that low-pressure crystal fractionation will have on the Lu/Hf and Sm/Nd ratio of a mare basalt can be
assessed using the low-pressure liquidus phases of basalts. Low-Ti basalts are likely to have fractionated a silicate mineral assemblage (e.g., Neal and Taylor, 1992), which would produce a lava with Lu/Hf and Sm/Nd ratios that are lower than the primary magma composition (Fig. 8). In contrast, high-Ti basalts would have fractionated silicates along with a titanium
The second complicating factor in evaluating the melting mineralogy of mafic basaltics is the effect that trapped melt, from the crystallization of the cumulate mantle, will have on the mineralogy of a source region. We do not consider it feasible that trapped melt from the LMO will remain as an amorphous material in the upper mantle from the time it was trapped in the cumulate mantle until the time of mafic basalt magmatism (on the order of 100's of million years). Unruh et al. (1984) considered that a trapped melt component would recrystallize to the same mineral assemblage with which it was equilibrated. However, such an assumption disregards the possibility that trace phases may also crystallize to accommodate incompatible elements (e.g., P producing apatite or whitlockite). Because such trace phases can play an important role in producing extreme incompatible element fractionations (e.g., Watson and Green, 1981), their role in magma genesis must be investigated. We have modeled the role of trapped melt by assuming that the trapped melt equilibrated and recrystallized with the surrounding minerals (e.g., Unruh et al., 1984) and by assuming that some of the trapped melt also crystallized to apatite (ap). This modeling approach may not represent natural conditions, but it does provide a means to evaluate natural conditions by establishing their limits.

The melting mineralogy calculations presented below seek to match the Lu/Hf ratios measured in mafic basaltics by partial melting of a cumulate mantle with a Lu/Hf ratio determined by the initial ε_{Hf} value of the measured basalt. We have also made analogous calculations using the Sm/Nd system. A successful model should predict similar degrees of partial melting for both the Lu/Hf and Sm/Nd systems. We consider these calculations robust because the variations in Lu/Hf ratios measured in mafic basaltics are large relative to their calculated Lu/Hf source ratios. For example, low- and high-Ti basaltics have Lu/Hf ratios that are less than chondrite ([Lu/Hf]_ch = 0.6), but their source(s) all have [Lu/Hf]_ch ratios greater than one. This large difference between the calculated Lu/Hf source ratio vs. the measured Lu/Hf ratio of the lava implies that the melting mineralogy was strongly Lu retentive, relative to Hf. In contrast, the Sm/Nd ratios of high- and low-Ti basaltics range from values that are slightly less than chondrite, to values that are greater than chondrite. The positive initial ε_{Hf} values of low- and high-Ti basaltics imply that all these basaltics were derived from sources with [Sm/Nd]_ch > 1. The differences between Sm/Nd ratios of the sources and the measured Sm/Nd ratios of the lavas indicate that low to moderate Sm/Nd fractionation occurred during partial melting to produce low- and high-Ti basaltics.

**Table 2.** D-values used for trace element models.

<table>
<thead>
<tr>
<th>Element</th>
<th>olivine</th>
<th>opx</th>
<th>pigeonite</th>
<th>pgr</th>
<th>ilmenite</th>
<th>plagioclase</th>
<th>garnet</th>
<th>apatite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sm</td>
<td>0.0043</td>
<td>0.072</td>
<td>0.037</td>
<td>0.29</td>
<td>0.0094</td>
<td>0.017</td>
<td>0.25</td>
<td>4</td>
</tr>
<tr>
<td>Nd</td>
<td>0.0024</td>
<td>0.077</td>
<td>0.0195</td>
<td>0.17</td>
<td>0.0076</td>
<td>0.024</td>
<td>0.15</td>
<td>2</td>
</tr>
<tr>
<td>La</td>
<td>0.039</td>
<td>0.065</td>
<td>0.123</td>
<td>0.43</td>
<td>0.073</td>
<td>0.007</td>
<td>8.05</td>
<td>3</td>
</tr>
<tr>
<td>Hf</td>
<td>0.011</td>
<td>0.012</td>
<td>0.030</td>
<td>0.24</td>
<td>0.406</td>
<td>0.0034</td>
<td>1.22</td>
<td>0.15</td>
</tr>
<tr>
<td>Sm/Nd</td>
<td>1.762</td>
<td>1.552</td>
<td>1.897</td>
<td>1.670</td>
<td>1.237</td>
<td>0.720</td>
<td>1.660</td>
<td>2.000</td>
</tr>
<tr>
<td>Lu/Hf</td>
<td>3.491</td>
<td>5.458</td>
<td>4.100</td>
<td>1.833</td>
<td>0.180</td>
<td>2.000</td>
<td>29.59</td>
<td>20.00</td>
</tr>
</tbody>
</table>

The partition coefficients represent median values (as discussed in text) from the following references: for olivine: Fujimaki et al. (1984), McKay (1986), Francalanci (1989), and Kennedy et al. (1993); for orthopyroxene: Schmider and Philpotts (1970), Fujimaki et al. (1984), Irving and Frey (1984), and Kennedy et al. (1993); for clinopyroxene: Haskin and Korotov (1977), Dostal et al. (1983), Fujimaki et al. (1984), McKay et al. (1986b), Dunn (1987), Francalanci (1989), Hart and Dunn (1993), Hauri et al. (1994), Johnson (1994), and Skulski et al. (1994); for plagioclase: Fujimaki et al. (1984) and Snyder et al. (1992); for garnet: Irving and Frey (1978); Fujimaki et al. (1984); Hauri et al. (1994), and Johnson (1994); references for the partition coefficients of pigeonite, ilmenite and apatite are reported in the text.
Table 3. Mode and melting proportions used for low-Ti basalt melting mineralogy models.

<table>
<thead>
<tr>
<th>Model 1</th>
<th>olivine</th>
<th>cpx</th>
<th>cpx</th>
<th>garnet</th>
<th>spinel</th>
<th>apatite</th>
<th>Cumulative melting %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stage 1 mode</td>
<td>0.7780</td>
<td>0.2000</td>
<td>0.0220</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>stage 1 melt</td>
<td>-1.4050</td>
<td>0.2000</td>
<td>2.2050</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>1.00</td>
</tr>
<tr>
<td>stage 2 mode</td>
<td>0.8000</td>
<td>0.2000</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>stage 2 melt</td>
<td>0.2000</td>
<td>0.8000</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Model 1A</td>
<td>stage 1 mode</td>
<td>0.7702</td>
<td>0.1980</td>
<td>0.0218</td>
<td>...</td>
<td>...</td>
<td>0.0100</td>
</tr>
<tr>
<td>stage 1 melt</td>
<td>-2.3850</td>
<td>0.2000</td>
<td>2.1850</td>
<td>...</td>
<td>...</td>
<td>1.0000</td>
<td>1.00</td>
</tr>
<tr>
<td>stage 2 mode</td>
<td>0.8020</td>
<td>0.1980</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>stage 2 melt</td>
<td>0.2000</td>
<td>0.8000</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Model 2</td>
<td>stage 1 mode</td>
<td>0.7580</td>
<td>0.2000</td>
<td>0.0220</td>
<td>0.0200</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>stage 1 melt</td>
<td>-2.4050</td>
<td>0.2000</td>
<td>2.2050</td>
<td>1.0000</td>
<td>...</td>
<td>...</td>
<td>0.05</td>
</tr>
<tr>
<td>stage 2 mode</td>
<td>0.7739</td>
<td>0.2000</td>
<td>0.0110</td>
<td>...</td>
<td>0.0151</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>stage 2 melt</td>
<td>-2.0000</td>
<td>0.8000</td>
<td>2.2000</td>
<td>...</td>
<td>0.0000</td>
<td>...</td>
<td>1.00</td>
</tr>
<tr>
<td>stage 3 mode</td>
<td>0.7878</td>
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<td>...</td>
<td>...</td>
<td>0.0152</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>stage 3 melt</td>
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<td>0.8000</td>
<td>...</td>
<td>...</td>
<td>...</td>
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</tr>
<tr>
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<td>stage 1 mode</td>
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<td>0.2000</td>
<td>0.0200</td>
<td>0.0200</td>
<td>...</td>
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<td>0.8000</td>
<td>...</td>
<td>...</td>
<td>...</td>
<td>0.0000</td>
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</tbody>
</table>

The source mineral assemblages that do not contain apatite are consistent with the inference that any trapped melt equilibrated with the surrounding minerals, as assumed by Urruh et al. (1984). The mineral assemblages that contain apatite are inferred to represent a source in which part of the trapped melt recrystallized to produce apatite. All melting models are designed to leave a residual phase assemblage of ol + cpx; consistent with high-Mg lunar glass multiple saturation experiments.

Table 4. Modes and melting proportions used for high-Ti basalt melting mineralogy models.

<table>
<thead>
<tr>
<th>Model 3</th>
<th>olivine</th>
<th>cpx</th>
<th>cpx</th>
<th>ilmenite</th>
<th>garnet</th>
<th>spinel</th>
<th>apatite</th>
<th>cumulative melting %</th>
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<td>-1.2050</td>
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<td>...</td>
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<td>...</td>
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<td>0.2000</td>
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<td>...</td>
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<td>...</td>
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<tr>
<td>Model 4A</td>
<td>stage 1 mode</td>
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<td>0.0980</td>
<td>0.0190</td>
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<td>0.1250</td>
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<td>...</td>
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<td>0.0000</td>
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<td>0.0200</td>
<td>0.0250</td>
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<td>...</td>
<td>...</td>
<td>0.0000</td>
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</tbody>
</table>

The source mineral assemblages that do not contain apatite are consistent with the inference that any trapped melt equilibrated with the surrounding minerals, as assumed by Urruh et al. (1984). The mineral assemblages that contain apatite are inferred to represent a source in which part of the trapped melt recrystallized to produce apatite. All melting models are designed to leave a residual phase assemblage of ol + cpx; consistent with high-Mg lunar glass multiple saturation experiments.
We have chosen to model the Lu/Hf and Sm/Nd ratios of mare magmas using an incremental, nonmodal, batch melting model, similar to the melting models that have been developed for mid-ocean-ridge basalt. The melting modes we have used were arbitrarily chosen, largely because the experimental investigations needed to characterize such melting reactions have yet to be conducted (e.g., Longhi, 1992b). However, in order to be consistent with experimental investigations that indicate primary low- and high-Ti magmas are multiply saturated with olivine and orthopyroxene at high pressure, the melting modes we have used were designed to leave a residual mineralogy of olivine and orthopyroxene at small degrees of partial melting (e.g., <5%). The modal and melting mineral proportions that were used for low- and high-Ti basalt are reported in Tables 3 and 4, respectively.

6.2.1. Low-titanium basalts

Melting calculations that use the source mineralogy inferred from the isotopic data (see above) for low-Ti basalts (ol + opx + cpx) indicate that the measured Lu/Hf ratios of the low-Ti basalts, excluding the Apollo 12 ilmenite samples, can be produced at small degrees of melting (0.5–2%; Fig. 9). The measured Lu/Hf ratios of the Apollo 12 ilmenite basalts are too low, as compared to their high calculated Lu/Hf source ratios, to be produced by partial melting of the proposed ol + opx + cpx low-Ti basalt source. Similar calculations using the Sm-Nd isotope system reveal that between 1 and 2% partial melting of an ol + opx + cpx source can produce the measured Sm/Nd ratios of the low-Ti basalts (Fig. 9b). Note that this incremental melting model leaves a residual mineralogy of only ol + opx at >1% melt extraction.

In order to directly compare the melting calculations for both the Lu-Hf and Sm-Nd isotope systems we have plotted these melting fields on a plot of $\delta_{\text{Lu/Hf}}$ vs. $\delta_{\text{Sm/Nd}}$. The $\delta_{\text{Lu/Hf}}$ parameter is a way of normalizing the measured Lu/Hf ratio of the lava to the calculated Lu/Hf source ratio and is defined as: $\delta_{\text{Lu/Hf}} = [\text{Lu/Hf}]_{\text{lava}}/[\text{Lu/Hf}]_{\text{source}}$. This definition is similar to the definition of Salters and Hart (1989); $\delta_{\text{Sm/Nd}}$ is defined in an analogous fashion. A plot of this type collapses the melting fields shown on Fig. 9a and b to a single line and shows the fit between calculated amounts of partial melting as determined using the Sm-Nd and Lu-Hf isotope systems (Fig. 9c). Partial melting of between 1 and 5% calculated from their initial $\varepsilon_{\text{Nd}}$ values (y-axis). The shaded field shows the same melting field, as above, for the Sm-Nd system; the lines indicate 1, 1.5, 2, and 3% (from left to right) partial melting. Clinopyroxene is exhausted at 1% partial melting. The low Lu/Hf ratios of the Apollo 15 and Apollo 12 feldspathic and pigeonite basalts can be produced at 0.5 to 2% partial melting; the low Lu/Hf ratios of the Apollo 12 ilmenite basalts cannot be produced from incremental batch melting of this pyroxenite source. (b) Plot of the $\delta_{\text{Sm/Nd}}$ measured for low-Ti basalts (x-axis) vs. their $\delta_{\text{Lu/Hf}}$. (c) Plot of the $\delta_{\text{Lu/Hf}}$ and $\delta_{\text{Sm/Nd}}$ values of low-Ti basalts (the $\delta_{\text{Lu/Hf}}$ value is a way of normalizing the Lu/Hf ratio of the lava to that of the Lu/Hf ratio of the source; see text for discussion). The curve shows the melting fields from Fig. 9A and 9B. The solid portions of the curve indicate a residuum of ol + opx + cpx; the solid portion indicates a residuum of ol + opx. Incremental melting a pyroxenite source 1.5–2% can produce the observed Lu/Hf and Sm/Nd ratios of the Apollo 15 and Apollo 12 ilmenite and one Apollo 15 basalt. The other low-Ti basalts require larger Lu/Hf fractionations than can be produced by incremental melting of a pyroxenite source.
1.5% of the low-Ti basalt source produces an excellent fit to the measured Sm/Nd and Lu/Hf ratios for the Apollo 12 feldspathic basalt and two of the three analyzed Apollo 15 basalts. The other low-Ti basalts have Lu/Hf ratios that are too low as compared to their calculated Lu/Hf source ratios and require melting a source region that contains a Lu retentive phase relative to Hf such as garnet.

Figure 10 shows four possible melting curves, the $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ of low-Ti basalts, and the $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ of low-Ti basalts corrected for crystal fractionation (arrowhead on Fig. 10). The crystal fractionation corrections are minor and are based on assuming 50% fractionation of a silicate assemblage as shown in Fig. 8. The modes and melting proportions for the melting curves labeled Model 1.1A, 2, and 2A are reported in Table 3. Models 1 and 1A show the affect of melting an ol + opx + cpx source that is apatite free (Model 1; this is the same source as shown in Fig. 9c) and apatite bearing (Model 1A). Apatite (ap) is a Lu retentive phase relative to Hf (Table 2) but Sm and Nd are compatible elements in apatite. The melting curve for incremental melting of an ol + opx + cpx + ap source (Model 1A; Fig. 10; Table 3) is similar in shape to an apatite free source (Model 1; Fig. 10; Table 3), the main difference is that larger degrees of partial melting are required to produce the same change in $\delta$[Sm/Nd]$_n$ as compared to the apatite free source; melting an apatite bearing source cannot reproduce the measured low Lu/Hf and moderate Sm/Nd ratios of the Apollo 12 ilmenite and pigeonite basalts.

The melting curves labeled Model 2 and 2A are for melting an ol + opx + cpx source that contains a small amount of garnet (2%) and had either a 0.5% (Model 2) or 1% (Model 2A) partial melt extracted in the garnet stability field prior to conversion of garnet to spinel continued melting in the spinel stability field. Such a polybaric melting model that includes a small amount of melting in equilibrium with garnet, followed by continued melting in equilibrium with spinel is similar to models of the petrogenesis of mid-ocean ridge basalts (e.g., Salters and Hart, 1989). These garnet/spinel melting curves reproduce the observed $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ values of the Apollo 12 ilmenite and pigeonite basalts and are consistent with the magmas last equilibrating with an olivine plus orthopyroxene residuum.

An important observation is that the garnet that is inferred in the source of Apollo 12 ilmenite and pigeonite low-Ti basalts cannot be from the initial crystallization of the LMO. Inclusion of garnet that crystallized from the initial LMO in the inferred low-Ti source would produce a source that has a [Lu/Hf]$_n$ to [Sm/Nd]$_n$ ratio that is much greater than that observed for low-Ti basalts. The implication of this observation is that the low-Ti ol + opx + cpx source must have formed garnet at the expense of aluminous pyroxenes by metamorphic equilibration.

6.2.2. High-titanium basalts

Although partial melting of an exclusively olivine and orthopyroxene source can produce the Lu/Hf and Sm/Nd ratios measured in high-Ti basalts, a source composed ex-
Fig. 11. (a) Plot of the $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ values of high-Ti basalts. The curves show the calculated aggregate liquid compositions for incremental melting of an ol + opx + cpx + ilmenite + apatite source (models and melting proportions reported in Table 4). The different symbols on the curves show percentage of partial melt as a function of each 0.5% increment of melt extraction (total melting of 1, 1.49, 1.99, 2.49, 2.96% from top to bottom as indicated on each curve), the solid lines indicate a residual mineral assemblage of olivine and orthopyroxene, the dashed portion of the curves have residual ol + opx + other phases as indicated in Table 4. See text for description of the different melting models. Polybaric melting of an ilmenite-bearing source can reproduce the low Lu/Hf and moderate Sm/Nd ratios of high-Ti basalts if the melting modes of ilmenite and clinopyroxene are high. Partial loss of the incrementally generated high-Ti magmas (e.g., Model 4) is required to reproduce the negative $\delta$[Sm/Nd]$_n$ values of some high-Ti basalts. (b) Plot of the $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ values of high-Ti basalts. The curves show the aggregate liquid compositions for initial melting of an ol + opx + cpx + ilmenite + garnet source in the garnet stability field, followed by continued melting in the spinel stability field. Initial melting in equilibrium with garnet does not require the high melting modes of ilmenite and clinopyroxene (as inferred for model 4 and 4A) to reproduce the low Lu/Hf ratios of high-Ti basalts.

exclusively of ol + opx is inconsistent with the low $\epsilon_{Hf}$ values measured for high-Ti basalts. Including a small amount of ilmenite in an ol + opx source will produce a source region that can match the low $\epsilon_{Hf}$ values measured for high-Ti basalts because ilmenite has a $D_{Lu/Hf} < 1$ (inclusion of ilmenite in the high-Ti source region also explains the high-Ti contents of this mare magma type). However, melting an ilmenite-bearing mantle makes it exceedingly difficult to produce a magma with a low Lu/Hf ratio, because ilmenite is Hf-retentive relative to Lu.

Figure 11a shows the calculated $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ of high-Ti basalts, the $\delta$[Lu/Hf]$_n$ and $\delta$[Sm/Nd]$_n$ values corrected for crystal fractionation, and four calculated melting curves for an ol + opx + cpx + ilm + ap source (Fig. 11a, Table 4). Because high-Ti basalts fractionated an assemblage of olivine + Ti-oxide phase, primary high-Ti
magnas were likely to have Lu/Hf ratios that are lower than those measured on these fractionated samples, implying that the $\delta$[Lu/Hf]$_m$ values of the plotted symbols are minimum values (Fig. 11a). The $\delta$[Lu/Hf] values for high-Ti basalt range from 0.37 to 0.62 and their $\delta$[Sm/Nd]$_m$ values range from 0.15 to -0.1. Negative $\delta$[Sm/Nd]$_m$ values imply that the measured Sm/Nd ratio of the sample is greater than the inferred Sm/Nd ratio of the source.

The melting curves labeled Model 3 and 3A are the calculated accumulated liquids for an ilmenite bearing ol
+ opx + cpx source ± apatite (Model 3A contains apatite) that would produce an ol + opx residuum after 2% melt extraction. The melting curves labeled Model 4 and 4A are the calculated accumulated liquids for the same source as in Models 3 and 3A but these melting curves reflect production of an ol + opx residuum after only 1% melt extraction. Rapid exhaustion of ilmenite and clinopyroxene from this source was modeled by assuming that the large melting proportions required to exhaust ilmenite and clinopyroxene were offset by precipitation of extensive amounts of olivine and orthopyroxene. Moreover, the Model 4 melting curve also reflects the affect of partial loss of magma generated during decompression melting in equilibrium with ilmenite. Partial loss of a high-Ti magma at pressure is a likely phenomenon because high-Ti magmas are more dense than the solid (e.g., Delano, 1990). Loss of early formed incremental magma batches prior to accumulation and storage of the accumulated liquid provides a reasonable method of reproducing the negative δ[Sm/Nd]₀ values measured for some high-Ti basalts. Melting an ilmenite bearing ol + opx + cpx source can reproduced the low Lu/Hf and moderate Sm/Nd ratios measured for high-Ti basalts if ilmenite and clinopyroxene can be consumed at low degrees of partial melting (e.g., Models 4 and 4A; Fig. 11a). Alternatively, if the first melt fractions were derived from an ol + opx + cpx + ilm source that also contained a small percentage of garnet (2%), followed by continued melting in the spinel stability field then the low Lu/Hf ratios of the high-Ti basalts can also be reproduced without requiring extensive olivine precipitation (Model 5A, 5B, 5C; Fig. 11b). Regardless of the inferred melting mineralogy and melting modes, the high-Ti basalts with negative δ[Sm/Nd]₀ values can only be produced by loss of early ilmenite-rich magma from the upwelling mantle diapir.

### 7. DISCUSSION

#### 7.1. Inferences Derived from Lu/Hf and Sm/Nd Source Ratio Calculations, Method 1

High- and low-Ti basalts define distinct variations in initial εᵤᵣ and εᵣₙd values which imply that the Lu/Hf and Sm/Nd source ratios of low- and high-Ti basalts are different. Low-Ti basalts define a Lu/Hf and Sm/Nd source ratio of \([\text{Lu/Hf}]₀ = 4.12\) and \([\text{Sm/Nd}]₀ = 2.91\) whereas high-Ti basalts define a correlation with \([\text{Lu/Hf}]₀ = 1.12\) and \([\text{Sm/Nd}]₀ = 0.08\). These differences in the Lu/Hf and Sm/Nd source ratios of high- and low-Ti basalts imply that the low- and high-Ti basalt sources must be composed of different minerals and that these minerals must have been stored and not significantly mixed with one another since the time the sources formed. Of particular significance is the high y intercept of the low-Ti basalt initial εᵤᵣ - εᵣₙd correlation. Low-Ti basalts define a positive correlation of \(εᵤᵣ = 3.79 + 5.48\), and the fact that this correlation line does not intersect the origin (i.e., \(εᵤᵣ = εᵣₙd = 0\)) requires that the low-Ti source was derived from a strongly fractionated source early in the evolution of the Moon and was not later mixed. In contrast, correlations in εᵤᵣ and εᵣₙd variations for terrestrial samples (e.g., Johnson and Beard, 1993) typically pass through or close to the origin reflecting the continual mixing and homogenization of sources on a tectonically active planet.

The variation in calculated Lu/Hf and Sm/Nd source ratios of low-Ti basalts can best be modeled by crystallization of an olivine and orthopyroxene assemblage with a small amount of clinopyroxene (2%) and variable amounts of trapped liquid (0–10%) that crystallized from the LMO at 0 to 80% solid (Fig. 6). The variation in calculated Lu/Hf and Sm/Nd source ratios for high-Ti basalts can be modeled by crystallization of a variety of silicate phases and ilmenite from the LMO after 95% solidification of the LMO (Fig. 6). A good fit to the Lu/Hf and Sm/Nd source ratios of high-Ti basalt can be obtained by a cumulate composed of olivine, orthopyroxene, trace amounts of ilmenite and clinopyroxene (±2% each), and 0–6% trapped liquid.

Hybridization models that suggest the variation in the Ti contents of mare magmas are produced from assimilation of ilmenite by primary low-Ti picritic magmas can reproduce the inferred variations in Lu/Hf source ratios and Ti contents of high-Ti basalts. However, assimilation of ilmenite in low-Ti magmas of moderate Ti contents (e.g., 6 wt% TiO₂) should produce vertical arrays in plots of initial εᵤᵣ vs. εᵣₙd (Fig. 7). Low-Ti basalts do not show evidence for such a mixing relationship; hence, we conclude that assimilation of ilmenite can-
not be the operating mechanism in producing the variation in Ti contents of mare magmas.

7.2. Inferences Derived from Melting Mineralogy Calculations, Method 2

The melting mineralogy calculations that have been presented are based on using a starting mineral composition inferred from the Lu/Hf and Sm/Nd ratio calculations and then modeling the liquid composition by assuming an incremental melting model, similar to models that were developed for the generation of mid-ocean-ridge basalts. The melting modes were chosen to produce a residual mineralogy of olivine and orthopyroxene, consistent with multiple saturation experiments on picritic lunar glasses.

Low-Ti basalts have low Lu/Hf ratios, but their calculated Lu/Hf source ratios are high indicating that there were large Lu/Hf fractionations during partial melting. Melting an ol + opx + cpx source between 1 and 2% can reproduce the measured Lu/Hf and Sm/Nd ratios of the Apollo 12 feldspathic basalt and two of the three analyzed Apollo 15 basalts. The low Lu/Hf ratios of the other Apollo 15 and Apollo 12 ilmenite and pigeonite basalts cannot be reproduced unless it is assumed that the original source started melting in the garnet stability field, followed by continued melting in the spinel stability field (Fig. 10). This inferred polybaric melting history is similar to that proposed for the origin of mid-ocean ridge basalts (e.g., Salter and Hart, 1989).

The Lu/Hf and Sm/Nd ratios of high-Ti magmas can be reproduced by partial melting of an ilmenite bearing ol + opx + cpx source if the melting proportions of ilmenite and clinopyroxene are much greater than 1 and significant amounts of olivine and/or orthopyroxene are precipitated during melting (Fig. 11a; Model 4 and 4A; Table 4). If the melting proportions are not this high, then high-Ti basalts also require initial melting of a garnet bearing source followed by continued melting in the spinel stability field (e.g., Fig. 11b). In either case, the negative δ[Sm/Nd]₀ values and high δ[Lu/Hf]₀ values of some high-Ti basalts can only be explained by partial loss of incremental liquids prior to storage and transport of the aggregate liquid in an upwelling mantle column during decompressive melting. Considering the high densities of ilmenite-saturated liquids, partial loss of this high-Ti magma is likely (e.g., Delano, 1990).

The garnet that is inferred in the source of low-Ti basalts (and perhaps high-Ti basalts) cannot have crystallized from the LMO because it would produce a source with a Lu/Hf to Sm/Nd that is higher than that inferred from the initial ε[H] and εNd values measured for low- and high-Ti basalts. Rather, we propose that the garnet was produced at the expense of aluminous pyroxenes and/or spinel via metamorphic reequilibration. Presumably metamorphic reequilibration was a result of the convective overturn of the cumulate layer caused by sinking of the ilmenite layer. The inference that many of the lunar basalts started melting in the garnet stability field indicates that the initial site of melt generation of mare magmas may be deeper than originally determined from multiple saturation experiments, which indicate last equilibration of an olivine orthopyroxene residuum at 20kbars (400km on the moon). The minimum pressure for garnet stability in terrestrial pyroxenite compositions at solidus temperatures is estimated to be 13–17 kbars (e.g., Hirschman and Stolper, 1996 and references there in). The pressures required for garnet stability in a pyroxenite contrast with a minimum pressure of 25 kbars for garnet stability at liquidus temperatures for a peridote composition (e.g., Hirschman and Stolper, 1996, and references there in). A bulk Moon composition is probably intermediate to that of terrestrial peridotites and pyroxenites (e.g., Taylor, 1982), and we infer that a lunar pyroxenite garnet would probably be stable at some intermediate pressure (e.g., 20 kbars; 400km depth in the Moon; e.g., Warren, 1985). Alternatively, synthesis experiments for a lunar pyroxenite composition indicate that at solidus temperatures garnet pyroxenite would be stable at 28–30kbars (e.g., Ringwood and Essene, 1970) which is equivalent to depths of approximately 600km. Computational models indicate that gravitational sinking of the upper ilmenite-bearing layer to depths as great as 550 km can be accomplished in 500 m.y. (e.g., Spera, 1992).

We suggest that a working model, similar to that proposed by Ringwood and Kesson (1976) for diapirc sinking of the dense ilmenite layer into the deep lunar mantle, can explain most of the combined Lu/Hf and Sm/Nd isotope data (Fig. 12). We favor a model that requires only limited mixing of the deep lunar mantle with the plume melting ilmenite blobs because the initial ε[H] and εNd data of low- and high-Ti basalts are only consistent with distinct mineralogic compositions that have remained unmixed since their inception (Fig. 12b). Polybaric melting of these different sources probably started in the garnet stability field and continued into the spinel stability field (Fig. 12c). The amount of melting in the garnet stability field is minimal (0.5–1%) and the amount of garnet required in the source is small (~2%). For all of these models, final melting was accomplished at higher levels (spinel stability) and final separation of the liquid was accomplished with a residua of only olivine and orthopyroxene. For some high-Ti magmas, loss of the first formed high-Ti melt is required, presumably caused by the high density of ilmenite-saturated magmas, to reproduce their negative δ[Sm/Nd]₀ values and high positive δ[Lu/Hf]₀ values.

8. CONCLUSIONS

The polybaric melting models presented above are sensitive to their input parameters, which include mineral-melt partition coefficients, the melting modes, and initial source mineral proportions. For the low-Ti melting models a variety of melting modes and starting mineral proportions (all consistent with a source that had [Lu/Hf]₀ ratios 4 times greater [Sm/Nd]₀ ratios) were investigated. These models all yield similar results. In order to explain the low Lu/Hf ratios (0.6 times chondrite) measured for Apollo 12 ilmenite and pigeonite basalts (derived from a source with [Lu/Hf]₀ ratios > 1.8), the initial site of melting must be in equilibrium with garnet. The melting models for high-Ti basalts are more difficult to interpret, largely because variation in melting modes can exert a strong control on the calculated aggregate liquid compositions (e.g., compare Models 3 and 4; Fig. 11a). In part, the uncertainty in the high-Ti basalt melting models reflects the addition of a fourth phase to the ol + opx + cpx composition (ilmenite) which has the effect of increasing the degrees of freedom in the model.
Clearly, as suggested by Longhi (1992b) polybaric melting models must be refined with further experimental results. However, we wish to stress that the following conclusions are independent of the melting calculations we have proposed. First, the hafnium and neodymium isotope compositions of low- and high-Ti basalts can only be explained by long-lived, mineralogically distinct sources. Second, the low Lu/Hf ratios of high- and low-Ti basalts require significant fractionation of Lu from Hf because these basalts were derived from a source with high Lu/Hf ratios. Such extreme Lu/Hf fractionations require either very small degrees of partial melting or melting of a source that has a Lu retentive phase such as garnet. Third, the moderate Sm/Nd ratios of high- and low-Ti basalts requires little fractionation of Sm from Nd because these were derived from a source that had Sm/Nd ratios only slightly greater than chondritic values. The moderate fractionation of Sm from Nd is not consistent with small degrees of partial melting.

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