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INVITED REVIEW

Magmatic evolution of the Moon

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ABSTRACT

Although incomplete because of the imperfect and somewhat random sampling of rock types by the Apollo and Luna missions (1969–1976), the history of lunar magmatism has been reconstructed by numerous researchers over the past three decades. These reconstructions have illustrated the continuous nature of lunar magmatism (from 4.6 to ~2.0 Ga) and the large influence of early differentiation and catastrophic bombardment on lunar mantle dynamics, magmatism, and eruptive style. In this review, we group magmatism into multiple stages of activity based on sampled rock types and evaluate the models for each stage.

Stage 1 is early lunar differentiation and associated magmatism. Partial melting of the Moon soon after accretion was responsible for producing an anorthositic crust and a differentiated lunar interior. The extent of lunar melting and mantle processing depends strongly on the mechanisms that induced melting. Estimates for the time over which melting and crystallization occurred range from tens to hundreds of millions of years. Stage 2 is the disruption of lunar magma ocean cumulates. Soon after the crystallization of most of the lunar magma ocean, the cumulate pile experienced gravitational overturn. This resulted in transport of late-forming cumulates into the deep lunar mantle and mixing of magma ocean cumulates on a variety of scales. Stage 3 is the post-magma ocean highland magmatism. Whereas the ferroan anorthositic crust was probably produced during the crystallization of a magma ocean, the slightly younger Mg suite and alkali suite plutonic rocks may have been generated by decompressional melting of early magma ocean cumulates during cumulate pile overturn. A KREEP and crustal signature was incorporated into these primitive basaltic magmas through assimilation near the base of the lunar crust or through melting of a hybridized mantle. The alkali suite could represent either the differentiation products of Mg suite parental magmas or a separate, but contemporaneous episode of basaltic magmatism. Stage 4 is pre-basin volcanism. Sample analysis and remote sensing data indicate that early lunar volcanism (KREEP basalts and high-alumina basalts) was contemporaneous with periods of highlands plutonism and catastrophic bombardment of the lunar surface. The relationship between early stages of lunar volcanism and the contemporaneous plutonism is not clear. The KREEP basalts may be volcanic equivalents to both the Mg suite and alkali suite. Stage 5 is the late remelting of magma ocean cumulates and eruption of mare basalts. Basin-associated eruption of mare basalts occurred during and following the late stages of catastrophic bombardment. This volcanic activity was possibly an extension of the thermal event that initiated pre-basin volcanism. Mare basalts exhibit a wide range of composition resulting from near-surface fractionation of chemically distinct primary basaltic magmas. Most likely, mare basalts were produced by small to moderate degrees of partial melting of hybrid cumulate sources in the deep lunar mantle. Alternatively, the mixed chemical signatures observed in many mare basalts may be interpreted as indicating assimilation of late-stage, evolved cumulates by melts produced deep in the cumulate pile. The wide range of compositions exhibited by the mare basalts compared with earlier episodes of basaltic magmatism may reflect the thermal regime in the lunar mantle that limited the extent of partial melting and melt-source homogenization.

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INTRODUCTION

Important advances in our understanding of the evolution of the Moon were made possible by the six Apollo missions (1969–1972) and three Soviet unmanned missions (1970–1976). These missions returned more than 380 kg of samples from nine localities on the near side of the Moon (Taylor 1982). Intense study of these samples and several lunar meteorites (Warren 1994) has led to a basic understanding and appreciation of early planetary igneous activity and crustal evolution that has long since been erased from the Earth's surface.

IGNEOUS ROCKS ON THE MOON

One of the fundamental discoveries of the Apollo missions is that the Moon is made of a variety of igneous rock types that differ widely in their minerals, composition, and age. The most visible evidence of these differences is the existence of two distinct terrains on the Moon: the light-colored feldspathic rocks of the highlands and the dark basalts of the maria. Physically, the highlands are rough, intensely cratered, and rise up to 3 km above the flat maria.

The lunar crust is ~60 km thick on the near side, to perhaps as thick as ~86 km on the far side (Taylor 1982). The highland surface on the Moon has a bulk composition equivalent to anorthositic gabbro (26 to 28% Al_2O_3) and possibly represents the composition of the upper lunar crust. The minimum thickness of the upper crust is ~19 km. The observed enrichment in noritic components in basin ejecta with increasing basin size suggests that the bulk composition of the lower lunar crust is noritic (~20% Al_2O_3). The total lunar crust has a bulk composition that corresponds to anorthositic norite (Ryder and Wood 1977; Spudus and Davis 1986).

Because of intense bombardment early in the history of the Moon, the upper lunar crust is fractured and brecciated to depths of 20 km and is covered by thick polymict ejecta and comminuted melt sheets (≥2 km). To unravel the complex history of the lunar crust, it is essential to separate rocks that formed from endogenous igneous activity from those formed by impact-generated melting. According to Warren (1985, 1993), the least ambiguous means of identifying pristine rocks that were neither formed nor contaminated by impact is by analysis of siderophile elements. This contamination results from the fact that most meteorites have high siderophile contents, and rocks formed by impact can usually be identified by elevated concentrations of these elements. Warren (1993) presented a list of pristine rock candidates.

Pristine highland rocks can be divided into two major groups based on bulk rock molar ratios Na/(Na + Ca) vs. Mg/(Mg+Fe) (Fig. 1) (Warren 1986). These two major rock types, the ferroan anorthosites and the Mg suite, are cumulates from distinct parental magmas (Warner et al. 1976; James 1980; Raedeke and McCallum 1980; Taylor et al. 1991). The ferroan anorthosites have been divided into several subgroups based on compositional variations in minerals (James et al. 1989). The Mg suite consists of dunites, troctolites, norites, and gabbronorites. Taylor et al. (1993) estimated that the Mg suite makes up ~20% of the lunar crust and that the ferroan anorthosites make up most of the rest. More evolved plutonic rocks (granites, rhyolites, and felsites) and pre-mare basalts also occur in the highlands. These rock types are uncommon in the returned sample suite, and their relationship to the more voluminous highland rock types is still debated.

The pristine highland rocks and the breccias and regolith derived from them make up the lunar highlands (~83% of the lunar surface). The remaining ~17% of the lunar surface or ~1% of the lunar crust is composed of mare basalts (Head 1976; Head and Wilson 1992). The mare basalts are enriched in FeO and TiO_2 and have higher CaO/Al_2O_3 ratios than highland rock types (Taylor et al. 1991). These chemical differences reflect the enrichment of highland rocks in plagioclase. Mare basalts display a large variation in TiO_2 content, and the TiO_2 content is used in their classification. The three major groups of mare basalts are high-Ti basalts (>9 wt% TiO_2), low-Ti basalts (1.5–9 wt% TiO_2), and very-low-Ti (VLT) basalts with TiO_2 wt% <1.5 (Taylor et al. 1991). Some of the important chemical characteristics of mare basalts are shown in Figure 2.

The temporal aspects of lunar magmatic events represented by these rock types are summarized in Figure 3. Early stages of magmatism, represented primarily by the ferroan anorthosites (~4.6–4.4 Ga), were responsible for the early differentiation of the Moon through the formation of the lunar crust and the mantle. Contemporaneous and subsequent periods of early crust production (4.5–3.9 Ga) correspond to the formation of Mg suite plutonic rocks, the alkali suite, and granitic rocks. The Mg suite magmas, which apparently intruded the early anorthositic crust, may have formed layered mafic complexes (James 1980) similar to the Stillwater Complex, Montana. During these early periods of magmatism, crust formation, and lunar mantle evolution, the Moon also experienced catastrophic bombardment. Periods of pre-mare volcanic activity overlapped periods of magmatism associated with the generation of the plutonic highland rock types and preceded many episodes of

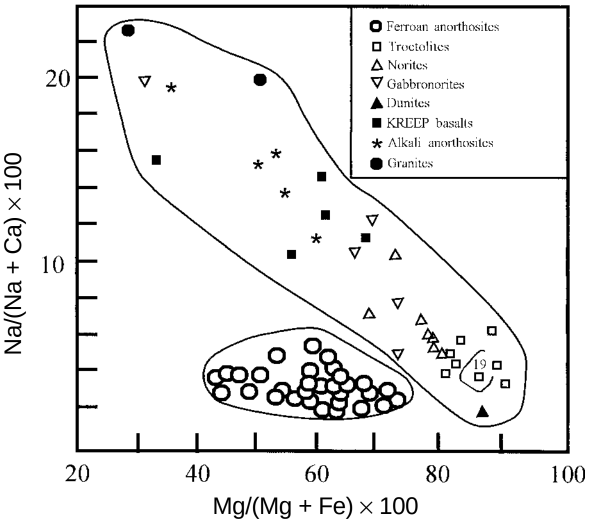


FIGURE 1. Compositional variability of the various lunar highland magmatic suites illustrated in a plot of molar Na/(Na+Ca) × 100 vs. Mg/(Mg+Fe) × 100 for bulk rock compositions (after Warren 1986).

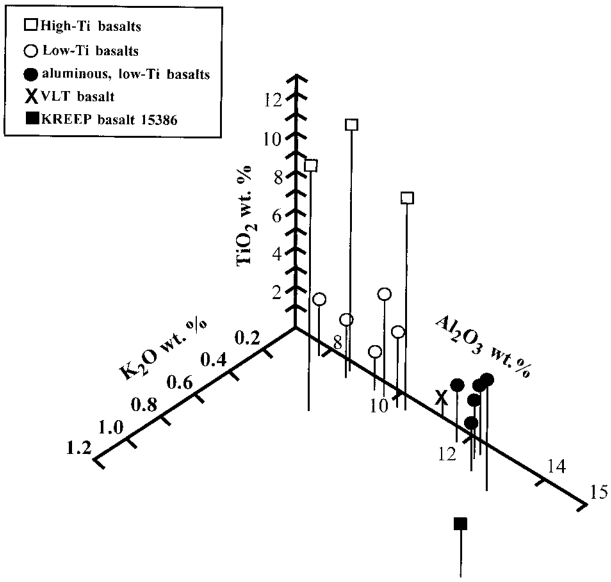


FIGURE 2. Compositional variability exhibited in the mare and KREEP basalts expressed in a plot of weight percents of K_2O-Al_2O_3-TiO_2.

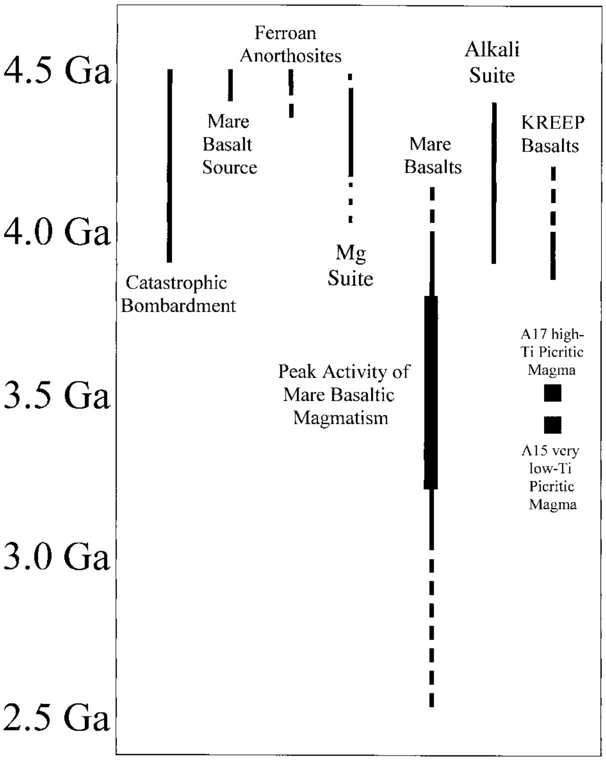


FIGURE 3. Bar diagram depicting age span for major magmatic events on the Moon. Age data are from Papanastassiou and Wasserburg (1971), McKay et al. (1979), Carlson and Lugmair (1988), Nyquist and Shih (1992), Shih et al. (1992), Head and Wilson (1992, 1997), and Snyder et al. (1995, 1999).

basin formation. These early periods of volcanic activity are represented primarily by clasts of KREEP and high-Al basalts in the lunar regolith and by "cryptomare" rocks identified by remote sensing (Head and Wilson 1992, 1997). Mare basaltic magmatism followed catastrophic bombardment and basin formation and continued to ~2.0 Ga. Mare basaltic magmatism reached its peak between 3.8 and 3.2 Ga.

In the following discussion of petrogenetic models for lunar magmatism, we grouped magmatism into stages of activity: Stage 1—Early Lunar Differentiation and Associated Magmatism; Stage 2—Disruption of Lunar Magma Ocean Cumulates; Stage 3—Post-Magma Ocean Highland Magmatism; Stage 4—Pre-Basin Volcanism; and Stage 5—Mare Basalts, Late Remelting of Magma Ocean Cumulates. The reader is reminded that this grouping of lunar magmatic activity primarily is based on sampled rock types returned by manned and robotic missions to the Moon. In most cases, lunar magmatism was a continuous expression of previous dynamic and thermal global processes.

STAGE 1: EARLY LUNAR DIFFERENTIATION AND ASSOCIATED MAGMATISM

Models for lunar differentiation and associated magmatism

Three basic models have been proposed for the early differentiation of the Moon and the formation of the early lunar crust: heterogeneous accretion (Ringwood 1972; Gast 1972; Papanastassiou and Wasserburg 1971), serial magmatism (Wetherill 1975; Walker 1983; Longhi and Ashwal 1985; Shih et al. 1993), and lunar magma ocean (Wood et al. 1970; Smith et al. 1970; Taylor and Jakes 1974; Shirley 1983; Warren 1985; Fig. 4).

In models advocating limited post-accretionary differentiation of the Moon (Fig. 4a), the compositional variability in the lunar mantle, the source for lunar magmas, was attributed to heterogeneous accretion. The lunar crust, rich in refractory elements such as Al and the REE, was accreted late in the formation of the Moon. These types of models were abandoned in the early 1970s because they failed to account for numerous observations that implied a differentiated nonchondritic mantle: (1) the 4.4–4.35 Ga Nd-Sm and U-Pb differentiation ages for the source regions of the mare basalts; (2) nonchondritic signature of mare basalts; (3) correlation between the total REE abundance and the magnitude of the Eu anomaly in basalts; (4) ubiquitous negative Eu anomaly in basalts that do not have plagioclase as a liquidus or near-liquidus phase; and (5) significant isotopic and chemical differences between high-Ti and low-Ti mare basalt sources.

As an alternative to accretion to account for a differentiated and heterogeneous Moon, Wetherill (1975) envisioned a differentiation model in which accretional melting of the Moon was restricted, creating isolated magma chambers that crystallized to form large layered intrusions consisting of mafic cumulates (source for mare basalts) capped with anorthosites (lunar highlands). An extensive primitive lunar mantle remained unprocessed during these events. Walker (1983) and Longhi and Ashwal (1985) expanded this type of lunar differentiation model. They suggested that convection in the lunar mantle resulted in the separation of the anothosites from layered intru-

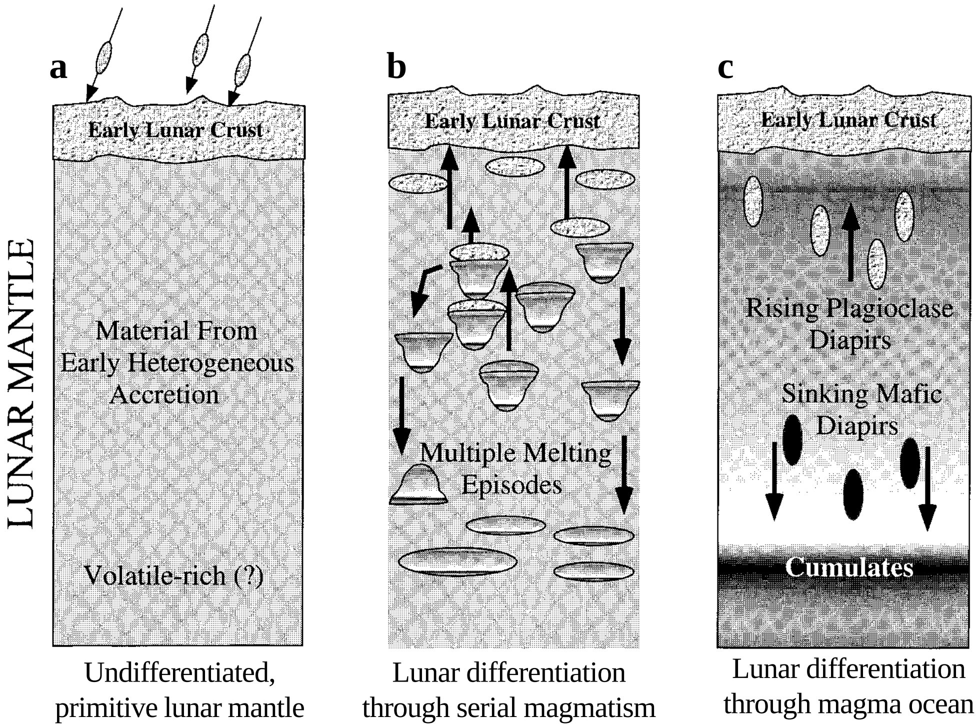


FIGURE 4. Models for the early evolution of the lunar mantle. (**a**) Undifferentiated lunar mantle. (**b**) Lunar differentiation through serial magmatism. (**c**) Lunar differentiation through magma ocean formation and crystallization.

sions and storage of the anothosites in the outer portions of the Moon. The coeval mafic component was buried in the mantle at depths greater than the sampling depth of multi-ring impact basins (30–60 km). At such depths, these mafic cumulates could have mixed with primitive lunar mantle assemblages that did not participate in the localized melting (Fig. 4b). The strength of this model is that it accounts for both the anorthositic lunar highlands and a differentiated lunar mantle (negative Eu anomaly, nonchondritic mantle) without extensive melting of the Moon. It does has several weakness, however (Hess 1989; Shearer and Papike 1993). First, this model does not adequately explain the diversity of mantle sources for the mare basalts (e.g., TiO_2 variability). Second, in comparison with terrestrial analogs, most of the larger layered intrusions do not have large anorthosite units. Third, there is very little evidence for the existence of a primitive lunar mantle that was involved in lunar magmatism. Fourth, the apparent consistency of the KREEP component is not compatible with basalt crystallization in relatively small, isolated magma chambers (Warren 1985). This latter argument is less compelling because the Lunar Prospector mission documented the preponderance of KREEP on the near side of the Moon (Feldman et al. 1998).

Wood et al. (1970) and Smith et al. (1970) suggested that the Moon experienced global melting and fractionation early in its history. In its simplest manifestation, this model postulates that the ferroan anorthositic crust was produced by flotation of plagioclase in a magma ocean. The mafic components coeval with the plagioclase do not appear in the crust but sank to the base of the magma ocean. These mafic cumulates provided the source for later lunar magmas (Fig. 4c). The magma ocean model has become widely accepted in lunar science because it is consistent with many of the fundamental lunar observations that could not be explained by an undifferentiated Moon (e.g., Nd-Sm and U-Pb differentiation ages, nonchondritic signatures of the lunar mantle, and the correlation between total REE abundance and the magnitude of the negative Eu anomaly in lunar rocks). Since originally proposed by Wood et al. (1970), Smith et al. (1970), and Taylor and Jakes (1974), numerous problems with the details of the magma ocean model have been recognized. There are indications that the sources of heat to melt the outer few hundreds of km of the Moon were insufficient or require very specific circumstances (Wetherill 1976; Ringwood 1976; Stevenson 1980). Another problem is that the layer cake cumulate pile resulting from the crystallization of the magma ocean implies that mare basalt-type should be related systematically to depth (Taylor and Jakes 1974). For example, the high-Ti mare basalts (incompatible-element-enriched) originated from shallow, late-cumulate sources, whereas the very low-Ti mare basalts (incompatible-element-depleted) were derived from deep, early cumulate sources. This simple model contradicts high-pressure experiments that indicate basalt composition is not related to depth of origin. Finally, magma ocean models do not readily explain some of the overlapping ages for the ferroan anorthositic suite and some Mg suite rocks as reported by Shih et al. (1993) and Borg et al. (1998).

Lunar magma ocean

Formation of the lunar magma ocean. The extent and duration of early lunar melting depends upon the mechanism providing primordial heat (Wetherill 1975, 1981; Alven and Arrhenius 1976; Sonett and Reynolds 1979; Hostetler and Drake 1980; BVSP 1981). Potential sources for primordial heat that have been advocated are accretionary heating from large impacts (Safronov 1978; Wetherill 1975, 1980, 1981; Ransford 1982), electromagnetic induction due to enhanced solar winds during early intense solar activity (Herbert et al. 1977), short-lived radioactive species, enhanced solar luminosity, core formation, tidal dissipation, and inherited heat (Binder 1978).

Many of these potential sources for heat would not have initiated substantial planetary melting or require unlikely circumstances. Core formation would have raised the Moon's mean temperature less than 10 °C (Solomon 1980; Warren 1985). Enhanced solar luminosity would have provided additional lunar heat only to the Moon's surface and would require the Moon to accrete quickly to be effective for any surface melting (Warren 1985). Electromagnetic induction potentially could melt the outer 400 km of the Moon (Warren 1985). However, this requires temporal relations among the early stages of intense solar activity, lunar accretion, and magma ocean formation. Short-lived radioactive heating through the decay of ^{26}Al ($t^{1/2}$ = 730 000 years) would also result in complete lunar melting under conditions of rapid accretion. Nonetheless, a fundamental problem with ^{26}Al as a heat source for lunar melting is that the magma ocean was formed probably 20–100 million years after the event that generated the ^{26}Al. Therefore, ^{26}Al was extinct and could not have provided heat following the accretion of the Moon.

If heat derived by accretion is the most likely cause of lunar melting, the conditions under which it occurred is key to the existence, size, and character of the magma ocean. With accretion involving relatively small projectiles (Safronov 1978; Wetherill 1980, 1981; Ransford 1982), heat would have been lost when subsequent impacts ejected material across the lunar surface (Wetherill 1976; Warren 1985). Slow accretion (>10^8 years) would have resulted in the inefficient storage of heat in the lunar mantle. Under these circumstances, the Moon would have experienced limited melting inconsistent with lunar differentiation through magma ocean processes. On the other hand, substantial degrees of melting would have occurred if the projectile radii during accretion were larger than 40 km (Warren 1985) and if accretion occurred over a short period of time (10^6–10^8 years). Therefore, rapid accretion is required to produce whole-moon melting, similar to models proposed by Wood (1972, 1975), Taylor and Jakes (1974), and Longhi (1977, 1980). Slightly slower accretion rates and smaller projectiles would have produced a partially molten "magmifer" alluded to by Shirley (1983) and Warren (1985). These differences are illustrated in Figure 5.

Magma ocean size, dynamics, and crystallization. The term magma ocean is vague in its description of early melting and differentiation of the Moon. There is no consensus about the depth of the magma ocean or the volume of undifferentiated lunar mantle that was melted. Estimates range from whole-Moon melting (Binder 1976, 1982; Runcorn 1977) to melting

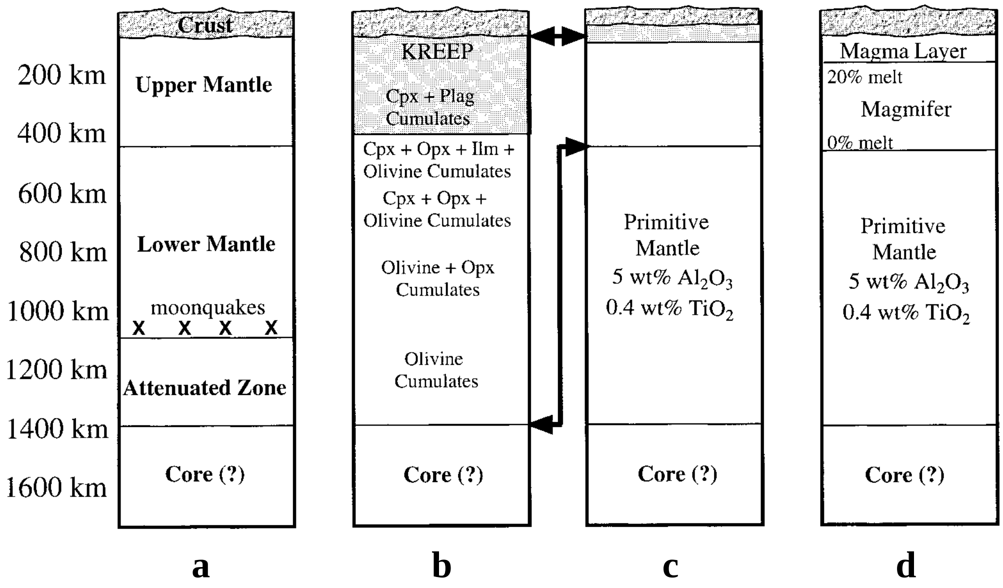


FIGURE 5. Different types of lunar magma oceans (LMO). (**a**) Geophysical structure of the Moon. (**b**) Deep LMO that consumes all of the primitive lunar mantle. (**c**) Shallow LMO that preserves the primitive lunar mantle at depth. (**d**) Magmifer processing of the primitive lunar mantle.

of only the upper 20 km (Kopal 1977). Several lines of evidence have been used to ascertain the extent of early lunar melting. First, the ineffectiveness of large impacts (e.g., Imbrium) to sample ultramafic cumulates that are complementary to the ferroan anorthosites (Warren 1985) and the rarity of ultramafic cumulates throughout the crust (Liebermann and Ringwood 1976) set a minimum depth for the magma ocean at 50 km. Second, mass-balance calculations have been made to estimate the amount of lunar melting that is required to produce the ferroan anorthositic upper crust that was formed from the magma ocean (Gast 1972; Hodges and Kushiro 1974; Wood 1972, 1975; Walker et al. 1975; Warren 1985). Assuming that the ferroan anorthosites make up 50% of the pristine upper crust, the bulk Moon contains 2.8% Al and if 100% of the Al in the magma ocean was extracted into the crust, then at least 36% of the Moon (depth of ~250 km) had to be melted (Warren 1985). Mass-balance calculations using incompatible-element enrichments in the crust imply a magma ocean depth of about 500 km (Taylor 1978). Third, Solomon and Chaiken (1976) estimated from compressional structures in the lunar crust that the Moon's radius has changed by <1 km over the last 3.8 Ga. Because compressional stresses are proportional to the degree of primordial melting, Solomon and Chaiken (1976) and Solomon (1980) concluded that the magma ocean was between 200 and 300 (±100) km deep. However, Binder (1982) argued that the change in the Moon radius was, in fact, >1 km and that this change indicated total lunar melting. Fourth, there is abundant evidence to suggest that mare basalts were produced by melting of a nonchondritic, differentiated lunar mantle. Estimates of the depth of origin for these basalts (250–1000 km) indicate that differentiated mantle existed to these depths (Longhi 1992). Finally, a deep magma ocean (>500 km) is consistent with geophysical models of the lunar interior (Taylor and Jakes 1974). Seismic data have been interpreted as indicating either incipient melting below 1000 km (Toksoz 1979) or a lower Mg' [=Mg/(Mg + Fe)] for the mantle below 1000 km (Goins et al. 1979). The latter result suggests that the 1000 km seismic boundary may be the base of the magma ocean cumulate pile that accumulated on either a convective, partially molten mantle or a primitive lunar mantle (Hollister 1975; Longhi 1981).

Many models for the thermal and chemical evolution of the lunar magma ocean are based on the assumption that the outer portion of the Moon was totally molten (Wood et al. 1970; Wood 1972, 1975; Hubbard and Minear 1975; Walker et al. 1975; Taylor and Jakes 1974; Longhi 1977; Solomon and Longhi 1977; Minear and Fletcher 1978). The dynamics of a totally molten magma ocean have been explored by numerous investigators (Solomon and Longhi 1977; Longhi 1977, 1981; Tonks and Melosh 1990; Spera 1992). Because of the expected high Rayleigh number of the magma ocean, convection would be turbulent (Spera 1992). As shown by Spera (1992), there are three basic flow regimes in such turbulent systems (Fig. 6). Adjacent to the base and top of the magma ocean, thin thermomechanical convective boundaries exist. These layers are equivalent to an early developing crust and a basal cumulate zone. Both are bounded by a viscous sublayer. Magma convection velocities within this sublayer are on the order of 10^{-2} m/s. The separation of crystals from melt is restricted to this flow regime near the floor region or along the base of the crust. The interior is dominated by inertial forces. Large convective velocities (on the order of 10 m/s) occur in this nearly isothermal region. Settling of crystals in this zone is precluded by the high rate of convection. Cooling rates were extraordinarily high during early stages of magma ocean crystallization because of the significant radiative heat loss through the discontinuous and thin protocrust. As the crust grew, the radiative heat loss decreased substantially. As the magma ocean continued to cool, the thickness of the boundary and viscous sublayers continued to grow at the expense of the interior region. The protocrust and cumulate zone continue to grow until they eventually merged at the base of the crust.

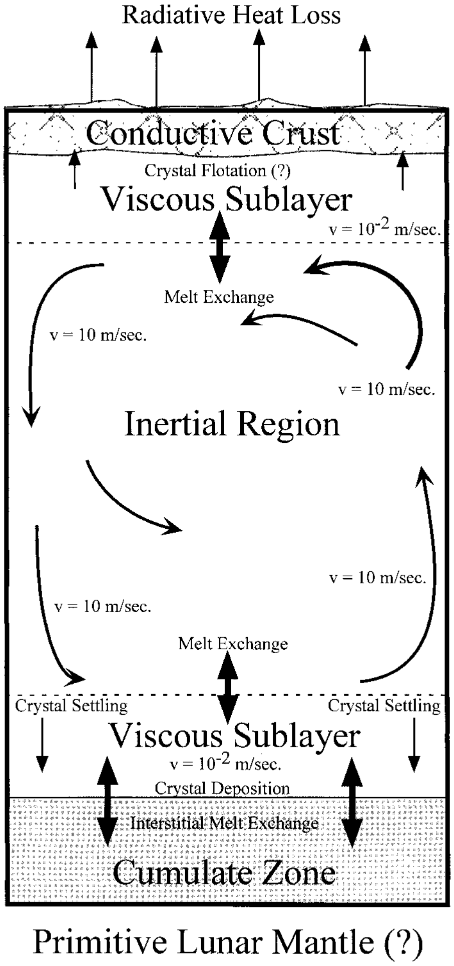


FIGURE 6. Possible structure of a large (>400 km) dynamic LMO (modified after Spera 1992).

The crystallization history of the magma ocean was summarized by Schnetzler and Philpotts (1971), Taylor and Jakes (1974), Taylor (1982), Longhi (1977, 1981), and Snyder et al. (1992). Schematic diagrams for the cumulate pile produced by magma ocean crystallization are presented in Figures 5 and 7. The sequence of crystallization is highly dependent on magma ocean bulk composition and the pressure and flow regimes under which crystallization occurred and is therefore difficult to pre-

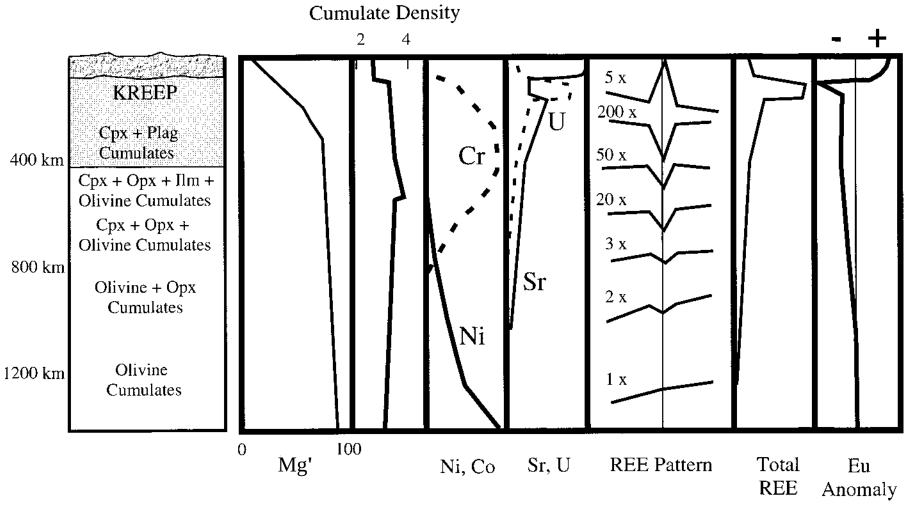


FIGURE 7. Mineral and chemical characteristics of the LMO cumulate pile (from Taylor 1982).

dict. In a dynamically simple magma ocean, the crystallization sequence advocated by most of these models is olivine → orthopyroxene ± olivine → olivine → clinopyroxene ± plagioclase → clinopyroxene + plagioclase → clinopyroxene + plagioclase + ilmenite. On the basis of estimated bulk compositions of the magma ocean (Taylor and Bence 1975; Ringwood and Kesson 1976; Buck and Toksöz 1980; Warren 1986; Hughes et al. 1988, 1989), olivine crystallizes first until the olivine-orthopyroxene boundary line is reached. The extent of the olivine cumulate assemblage in the magma ocean cumulate pile has been estimated to be between 30 and 40%. The effect of the olivine-orthopyroxene field boundary on cumulate assemblages depends on both the pressure of crystallization and the efficiency of crystal accumulation (equilibrium vs. fractional crystallization). As pressure increases, this boundary moves toward the olivine apex of the olivine-anorthite-SiO_2 pseudoternary, which has two effects. First, this shift effectively decreases the volume of the monomineralic olivine assemblage in the cumulate pile. Second, the olivine-orthopyroxene boundary becomes a cotectic surface, resulting in coprecipitation of olivine + orthopyroxene regardless of the efficiency of crystal separation. Within the context of the flow regimes proposed by Spera (1992), olivine and orthopyroxene precipitated from magmas at fairly high-pressure near the base of the cumulate zone. At the upper magma ocean boundary, significant radiative heat loss resulted in rapid olivine and orthopyroxene crystallization. These early crystallization products of the upper boundary layer should sink and be incorporated into the inertial inner region of the magma ocean. It appears unlikely that these early phases from the upper boundary would have been incorporated into the basal cumulate layer. The appearance of low-Ca (pigeonitic) and high-Ca clinopyroxene followed the precipitation of orthopyroxene and olivine. The exact sequence of the appearance of these pyroxenes is compositionally dependent. The relationship among clinopyroxenes and plagioclase in the magma ocean crystallization sequence is discussed in detail by Longhi (1980). In the bulk composition used by Snyder et al. (1992), high-Ca clinopyroxene crystallized after plagioclase.

The appearance of plagioclase in the crystallization sequence is extremely important to our fundamental understanding of the development and evolution of the early lunar crust. In simple dynamic models, the lunar crust is thought to have been formed by plagioclase crystallization and flotation after substantial amounts of the magma ocean had crystallized. Using the initial bulk composition suggested by Warren (1986) in which the Al_2O_3 was equal to 7 wt%, Snyder et al. (1992) calculated that plagioclase would be a liquidus phase after 57% of the magma ocean crystallized. Using estimated bulk compositions containing <5 wt% Al_2O_3, Snyder et al. (1992) estimated that plagioclase would appear in the crystallization sequence after 70–80% crystallization. The ability of plagioclase to float and accumulate in "rockbergs" at the surface also depends on the Mg' of the magma ocean. Flotation of anorthositic rocks (plagioclase + minor mafic minerals) may have been impossible until the "magmasphere" became more Fe-rich after substantial crystallization of mafic silicates (Warren 1985). High degrees of crystallization prior to plagioclase flotation are also implied by the trace element characteristics of the ferroan anorthosites (Palme et al. 1984; Papike et al. 1997). On the basis of these models, there should be a large portion of the cumulate pile that has either no Eu anomaly or only a minor negative one (Taylor 1982; Shearer and Papike 1989; Snyder et al. 1992).

In more complex models for magma ocean dynamics (Spera 1992), it could be expected that plagioclase was a liquidus phase during a substantial period of crystallization (Longhi 1980). Because of rapid heat loss of the upper magma ocean boundary, plagioclase might have become a liquidus phase earlier in the upper boundary than in the lower boundary layer. This may have resulted in formation of a protocrust prior to 60% magma ocean crystallization. In addition, mafic components that "saw" this early plagioclase crystallization may have imposed a negative Eu anomaly on early basal cumulates (olivine + orthopyroxene) by sinking into the central zone of the magma ocean.

Ilmenite-bearing cumulates were precipitated after 90%

crystallization of olivine+pyroxene+plagioclase. Ilmenite made up between 3–12% of the cumulate minerals. Incompatible trace elements excluded from the crystal structures of olivine, pyroxene, and plagioclase were concentrated in the residual melt. These residual liquids were enriched in K, REE, and P ("KREEP"), Th, U, Zr, Hf, and Nb. The signature of this KREEP component was later incorporated into the feldspathic highland crust through remobilization, assimilation, and mixing.

At low accretion rates, whole-Moon melting or complete melting of a large portion of the Moon is unlikely. Shirley (1983) and Warren (1985) proposed a shallow, partially molten magma ocean model to account for early lunar differentiation under this accretional scenario. In the model advanced by Shirley (1983), the lunar crust sits upon and continually grows from a fully molten layer that is estimated to be ~10 km thick. This thin molten magma chamber occurs at the top of a 300–400 km thick partially molten layer. The top of the partially molten layer [referred to as a "magmifer" by Shirley (1983) and Warren (1985)] contained 20% melt, which decreased to 0% melt at its base. The fully molten chamber was continually fed new batches of magma from the underlying, convecting magmifer. Crystallization of these batches of magma would result in crystallization and accumulation of mafic cumulates near the base of the fully molten layer. The denser mafic cumulates would eventually sink to the base of the magmifier and become isolated from the magma system. Plagioclase would be segregated through flotation and be added to the evolving lunar crust. Initial crustal formation occurred over an upwelling part of the early lunar convection system and spread laterally. The system evolved until the entire Moon was surfaced with anorthositic crust and a magmifier. In this manner, with limited degrees of partial melting, a large portion of the primitive lunar mantle may have been processed in a conveyer belt type mechanism.

The duration over which the magma ocean crystallized is uncertain. Part of the problem is interpreting the precise age of the ferroan anorthositic crust. The anorthosites are among the oldest lunar rocks. But because of low abundances of Rb and REE, intense meteoritic bombardment and prolonged subsolidus cooling, their Rb-Sr and Sm-Nd internal isochron ages might not reflect actual crystallization ages and are difficult to interpret. Low ^{87}Sr/^{86}Sr ratios for ferroan anorthosites approach those of primitive meteorites, implying that they have been separated from any significant concentrations of Rb for a length of time comparable with the age of the Moon (Nyquist and Shih 1992). Model ages for nonpristine rocks using U-Pb isotopic systematics indicate a major differentiation event occurred at about 4.47–4.56 Ga (Carlson and Lugmair 1988; Hanan and Tilton 1987; Premo and Tatsumoto 1991). Young ages for some ferroan anorthosites (Borg et al. 1998) may be explained by prolonged subsolidus cooling, but their derivation from light rare-earth element (LREE)-depleted sources (Borg et al. 1998) is still open to interpretation.

The idea that the KREEP component in the lunar crust was the residuum from the crystallization of the magma ocean has played a key role in estimates of the duration of lunar differentiation (Papanastassiou et al. 1970; Tera and Wasserburg 1974; Lugmair and Carlson 1978). The average model age for KREEP is 4.42 ± 0.07 Ga for Sm/Rb = 2.3 (Nyquist and Shih 1992). The interpretation of this model age is open to speculation in that it may represent a true crystallization age, a cooling or disturbance age, or a time of mixing. It represents the maximum duration over which the magma ocean crystallized. Model ages for the mare basalts are interpreted as indicating that their source region (i.e., mafic cumulates from magma ocean crystallization) was isotopically closed by 4.4 Ga.

In addition to the age of the rocks presumably produced during the early differentiation of the Moon, the duration of magma ocean crystallization has been estimated from thermal modeling (Solomon and Longhi 1977; Longhi 1980) and the isotopic systematics of short-lived isotopes (Shearer and Newsom 1998). Solomon and Longhi (1977) and Longhi (1980) suggested that the time for the completion of the crystallization of the magma ocean was 100–200 million years. Their thermal model is independent of initial magma ocean depth, but assumes an initial crustal thickness of 20 km (Solomon and Longhi 1977). Based on W-Hf isotopic systematics (Lee et al. 1997), Shearer and Newsom (1998) suggested that the duration for magma ocean crystallization was less than 40 million years. This estimate was based on the observation that the lunar mantle contained nonchondritic W isotopic reservoirs (Lee et al. 1997), which were produced during the last-stages of lunar magma ocean (LMO) crystallization by the fractionation of Hf from W (caused by precipitation of ilmenite and Ca-rich clinopyroxene) and the decay of ^{182}Hf. ^{182}Hf has a half life of ~9 million years and would have decayed totally soon after the Moon accreted. This shorter duration of LMO crystallization indicates that the Moon may have lost heat more quickly than assumed by Solomon and Longhi (1977). The shorter crystallization time of the magma ocean implies that the early lunar crust was relatively unstable, and that high meteorite influx was a more important factor controlling early lunar heat loss than was conduction through a thick lunar crust.

Mantle sources produced for subsequent lunar magmatism. An appealing aspect of these magma ocean models is that they provide processed mantle assemblages that are appropriate sources for subsequent periods of lunar magmatism. Different cumulate rock types have been identified as potential sources for a wide range of Mg suite parental magmas and mare basalts. For example, the highly fractionated KREEP component in the cumulate pile is thought to have been remobilized and emplaced into the lunar crust. Mg suite magmas are also thought to have acquired a KREEP signature through assimilation or mixing. The late-stage, ilmenite-bearing cumulates have been considered as possible sources for the high-Ti mare basalts (Taylor and Jakes 1974; Taylor 1982; Snyder et al. 1992), whereas early olivine-orthopyroxene dominated cumulates have been proposed as the mantle source for the very-low Ti picritic magmas (Taylor and Jakes 1974; Taylor 1982).

Chemical signatures within the magma ocean cumulate stratigraphy are illustrated in Figure 7. For example, early cumulates should be enriched in elements such as Ni and perhaps Co, and have rare-earth element (REE) patterns with virtually no Eu anomaly (Shearer and Papike 1989). Later cumulates should develop a REE pattern with negative Eu anomalies, be systematically enriched in incompatible elements and be depleted in elements such as Ni.

STAGE 2: DISRUPTION OF LUNAR MAGMA OCEAN CUMULATES

Several potential problems have been identified with the magma ocean crystallization sequence. First, based on the above cumulate stratigraphy, the very-low-Ti basalts should be derived from source regions below those of the high-Ti basalts. The source for the high-Ti basalts should be particularly shallow because ilmenite-bearing cumulates were deposited after more than 90% crystallization of the magma ocean. Using the extreme model of whole-moon melting, this would put the first ilmenite-bearing cumulates at depths of <175 km. Yet, high-pressure experiments on picritic mare glasses of all compositions demonstrate that they represent partial melting of sources at depths >400 km (Delano 1980, 1986; Longhi 1992). Second, although there is a systematic relationship between REE concentration and the magnitude of the Eu anomaly in mare basalts (and all lunar basalts have a negative Eu anomaly), there should be a substantial portion of the cumulate pile that did not experience plagioclase removal. Third, although the sources for the mare basalts had experienced prior plagioclase removal, many other characteristics appear not to be consistent with this source being the sinking cumulate fraction complementary to the floating plagioclase fraction. In contrast to the parent magma that produced the basal cumulate, the ferroan anorthosite parent magma was more evolved as indicated by its lower Mg' and Ni content, but less evolved as indicated by its low abundance of clinopyroxene, flat REE pattern, and low incompatible-element abundances (Ryder 1991). Fourth, the Mg' in the upper portions of the cumulate pile should be lower than that in early cumulates. In particular, the Mg' of the ilmenite-bearing cumulates should be substantially lower. Yet, very low-, low-, and high-Ti mare basalts (early to late cumulate sources) have similar Mg' (Ringwood and Kesson 1976). Fifth, a KREEP component appears to have been incorporated into very primitive mare basalts. Geochemical modeling indicates that this KREEP component was incorporated into the basalts at their sources (>400 km) rather than through assimilation (Shearer et al. 1991).

In response to these observations and interpretations, LMO proponents have introduced several modifications to the basic magma ocean layer-cake cumulate stratigraphy. In this cumulate pile, there is an upward progressive enrichment in FeO and thus in the density of cumulates (Fig. 7). The cumulate pile should be gravitationally unstable resulting in overturning of the cumulate pile. There are varying scales over which catastrophic overturning may occur. Snyder et al. (1992) suggested that cumulate mixing occurred over a very local scale, on the order of tens or hundreds of meters. This scale of overturning will not satisfy some of the objections to the layer cake cumulate model. Ringwood and Kesson (1976) suggested that the cumulate pile was partially disrupted by the sinking of dense pods of Fe- and Ti-rich cumulates into the earlier cumulate pile over a period of several hundred million years. Numerous researchers have suggested that the gravitational instability resulted in cumulate pile overturn on a global scale (Dowty 1975; Herbert 1980; Spera 1992; Shearer et al. 1990, 1991; Ryder 1991; Shearer and Papike 1993; Hess and Parmentier 1995). Models of large-scale overturn of the cumulate pile indicate that hypersolidus to subsolidus convection tended to mix magma ocean crystallization products over a period of 200 Ma (Spera 1992). Ryder (1991) suggested that overturning of the cumulate pile may have taken a shorter amount of time, perhaps aided by lubrication of residual melt products. Hess and Parmentier (1995) suggested that the dense ilmenite-bearing cumulates may have even formed the small lunar core.

Cumulate overturn not only resolves some of the basic problems with the magma ocean model, but it also provides a possible heat source for initiating melting in the deep magma ocean cumulate pile (olivine ± orthopyroxene). The transport of the evolved cumulates to the deep lunar mantle will redistribute incompatible, heat-producing elements such as U, Th, K, REE, and Rb (Shearer et al. 1991; Shearer and Papike 1993; Hess and Parmentier 1995)

STAGE 3. POST-MAGMA OCEAN HIGHLAND MAGMATISM: MG SUITE, ALKALI SUITE, AND GRANITIC MAGMATISM

After the early anorthositic lunar crust formed and consolidated, it was intruded episodically by slightly younger of magmas. The products of this younger magmatism are referred to as the Mg suite and the alkali suite. The Mg suite consists primarily of troctolites and norites, with lesser amounts of gabbros and dunites. This suite contains rocks with contrasting primitive and evolved magmatic chemical signatures. For example, rocks of the Mg suite commonly contain very magnesian mafic minerals with Mg' >90 coupled with high REE and low Ni abundances (Warren and Wasson 1977; Norman and Ryder 1979; Ryder 1991; Papike et al. 1996; Shearer and Floss 1999). The parent magmas appear to have been saturated with plagioclase and at least one mafic silicate. It was recognized fairly early in studies of the lunar highland rocks that the ferroan anorthosites and the Mg suite defined different compositional fields in plots of An content of plagioclase vs. Mg' of mafic phases (Fig. 1).

As with the ferroan anorthosites, the petrologic history of the Mg suite is obscured to various degrees because of the effect of intense meteoritic bombardment and prolonged subsolidus cooling. Isotopic data indicate that Mg suite magmatism immediately followed the generation of the ferroan anorthositic lunar crust and extended over a period of 400 my (Carlson and Lugmair 1981a, 1981b; Carlson 1982; Dasch et al. 1989). Initial Sr isotopic compositions of the Mg suite plutonic rocks are higher than those of ferroan anorthosites (Carlson and Lugmair 1981a, 1981b). The Mg suite appears to mark the transition between magmatism associated with the magma ocean and post-magma ocean serial magmatism. This transition period may have occurred as early as 30 Ma (Shearer and Newsom 1998) to as late as 200 Ma after LMO formation (Solomon and Longhi 1977; Longhi 1980). Some of the younger ages for the Mg suite may not be crystallization ages, but are more closely related to subsolidus closure (Carlson and Lugmair 1981a, 1981b; Carlson 1982).

Early studies of the lunar highlands also identified rocks with anomalously high alkali element contents (Hubbard et al. 1971) and highly evolved mineral and chemical signatures (Brown et al. 1972). These alkali-rich lunar rocks are generally defined as containing greater than 0.1% K_2O and 0.3% Na_2O (Brown et al. 1972; Hubbard et al. 1971; Warren and Wasson 1980). Although this is low compared with terrestrial magmatic rocks, the bulk Moon is highly depleted in alkali and volatile elements (Taylor 1982). The alkali suite is also characterized by elevated incompatible lithophile elements. This distinctive suite of highland rocks consists of anorthosites, gabbros, monzodiorites, quartz monzodiorites and a range of siliceous varieties referred to as granites, rhyolites, and felsites (Hubbard et al. 1971; Brown et al. 1972; Warren and Wasson 1980; Snyder et al. 1995); collectively, rocks of the alkali suite appear to make up a much smaller portion of the lunar crust than the ferroan anorthosites and Mg suite. The period of alkali suite magmatism extended from ~4.4 Ga to ~3.9 Ga. On the basis of ages, mineral compositions, and relative incompatible-element contents, Warren and Wasson (1980) demonstrated that these rock types were unrelated to either the ferroan anorthosites or the mare basalts. The alkali suite might represent evolved differentiates of highland Mg suite magmatism or a separate episode of magmatism that is the plutonic equivalent of KREEP basaltic magmatism (Warren and Wasson 1980; Warren et al. 1981; Hunter and Taylor 1983; Longhi 1988; Snyder et al. 1995). The more evolved members of this suite, such as the granites, rhyolites, and felsites, may be products of either extensive differentiation (Marvin et al. 1991) or liquid immiscibility (Neal and Taylor 1989)

Models for the generation of the Mg suite

Numerous petrogenetic models have been proposed to account for the contrasting primitive (e.g., high Mg') and evolved (e.g., saturated with plagioclase, high concentrations of incompatible trace elements, low concentrations of Ni) characteristics of the Mg suite. Models for the petrogenesis of the Mg suite include: (1) impact origin (Taylor et al. 1993; Hess 1994) (Fig. 8a); (2) products of magma ocean crystallization (Wood 1975; Longhi and Boudreau 1979; McCallum 1983) (Fig. 8b); (3) remelting and remobilization of late magma ocean cumulates and/or KREEP infiltrated lower crust (Hess et al. 1978; Hess 1989) (Fig. 8c); (4) Mg-rich magmas derived from lower portions of the cumulate pile that were enriched in Al and incompatible trace elements through assimilation of KREEP or anorthositic crust or both (Warren and Wasson 1980; Longhi 1981; James and Flohr 1983; Warren 1986; Ryder 1991; Papike et al. 1994, 1996) (Fig. 8d); and (5) decompressional melting of deep, hybrid mixed cumulate sources (Fig. 8e).

An impact origin for the Mg suite has been proposed to explain the chemical paradox of primitive and evolved chemical signatures in the same rocks (Wanke and Dreibus 1986; Taylor et al. 1993; Hess 1994). The model proposed by Taylor et al. (1993) combined late accretion impactors (~bulk Moon) as a source for the primitive component with magma ocean crystallization products (anorthosite, KREEP) as a source for the evolved component (Fig. 8a). The impact of this material into the Moon during the end of magma ocean crystallization would have mixed this primitive material with remelted ferroan anorthosite and residual KREEP liquid. The resulting magmas pooled beneath the ferroan anorthositic crust and subsequently intruded the crust. There are several difficulties with this model. The trace element fingerprints for impactors, such as elevated siderophile element abundances, are not found in the Mg suite. For example, the Cr/Ni ratios for the highland Mg suite show a typical lunar value (Cr/Ni >5) in contrast with primitive cosmic abundances (Cr/Ni = 0.25). There is also a mass-balance problem with this process. The ratio of the mass of the impactor to the mass of the impact melt is too small (O'Keefe and Ahrens 1977) to make a substantial contribution to the high Mg' observed in the Mg suite. Formation of the Mg suite by impact also circumvented the problem of a heat source capable of producing large volumes of primitive, high-Mg magmas following magma ocean crystallization (Taylor et al. 1993). How-

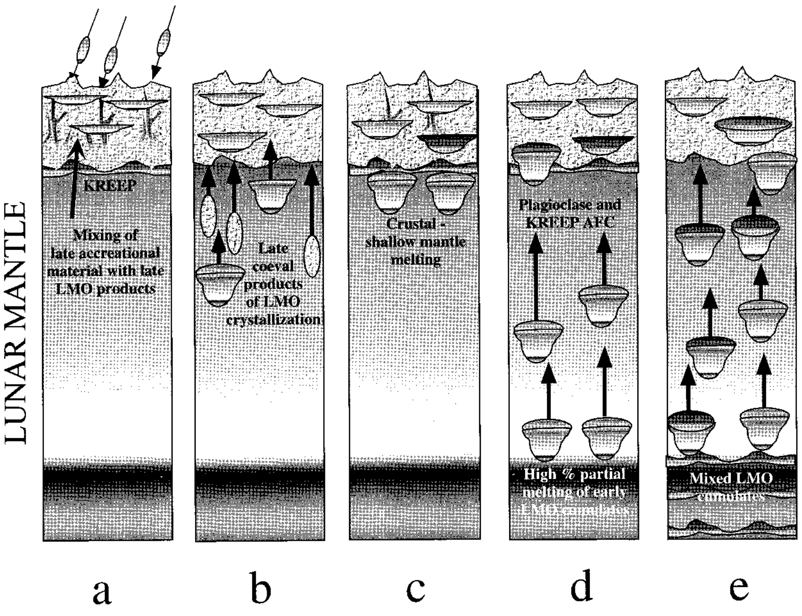


FIGURE 8. Models for the generation of the Mg suite plutonic rock types. (**a**) Late-accretionary components mixing with late-stage LMO components. (**b**) Products of LMO crystallization that were coeval with the ferroan anorthosites. (**c**) Shallow crustal melting of LMO cumulates. (**d**) High degrees of partial melting of early LMO cumulates followed by the assimilation of KREEP and plagioclase. (e) Melting of a mixed cumulate pile.

ever, several studies have identified viable processes that could have triggered melting in the deep lunar mantle (Spera 1992; Ryder 1991; Warren and Kallemeyn 1993; Shearer and Papike 1993).

Hess (1994) explored the possibility that the Mg suite was generated by impact melting of the plagioclase-rich lunar crust and olivine-rich cumulates of the magma ocean. This eliminates the mass-balance problem in the model of Taylor et al. (1993). Hess postulated that large impact melt sheets that were superheated and sufficiently insulated could cool slowly and differentiate to produce the troctolite-norite-gabbro sequence observed in the Mg suite. Variable incorporation of KREEP and crustal components during shock melting could explain the variation in the evolved component in the Mg suite plutonic rocks. The siderophile element signature of the impactor would have been significantly diluted during these processes. If an impact origin for the Mg suite is correct, one would not expect that the ferroan anorthosite and Mg suite chemical trends to be distinctly different (Fig. 1). The observed trends cannot be rationalized in terms of mixing of crustal and magma ocean cumulate components. Hess (1994) also pointed out that impact melts involving a substantial upper mantle component should have olivine on the liquidus and olivine-rich mineral assemblages are expected. However, dunites are not well represented in the Mg suite. Finally, if a substantial upper mantle component was incorporated into surface impact melt sheets, it should be expected that upper mantle lithologies would have been excavated and incorporated into the lunar regolith. No such samples have been found in the lunar regolith.

Wood (1975), Longhi and Boudreau (1979) and Raedeke and McCallum (1980) proposed models in which the Mg suite was produced during magma ocean crystallization and evolution (Fig. 8b). Wood (1975) suggested that the Mg suite and ferroan anorthosites were simply contemporaneous products of crystal accumulation and trapped melts. In this scenario, the Mg suite intrusions consist of cumulus olivine/pyroxene plus plagioclase, whereas the ferroan anorthosites consist of cumulus plagioclase with mafic crystals produced from intercumulus melts. In addition to the differential incorporation of cumulates and trapped melts, Longhi and Boudreau (1979) proposed that the cumulus minerals in both rock types were produced by different styles of crystallization. The plagioclase in the ferroan anorthosites were products of equilibrium crystallization. The cumulate mafic silicates of the Mg suite precipitated at approximately the same time, but under conditions of fractional crystallization. Raedeke and McCallum (1980) demonstrated that minerals from the banded zone in the Stillwater Complex showed two fractionation trends remarkably similar to the $Mg/(Mg+Fe)_{\text{mafic mineral}}$ vs. $An_{\text{plagioclase}}$ fractionation exhibited by the lunar highland plutonic rocks (see Fig. 1). They attributed the bimodality of the Stillwater rocks to differences in the style of crystallization (fractional crystallization accompanied by crystal accumulation vs. equilibrium crystallization of trapped intercumulus liquid in a plagioclase-rich crystal mush). A contemporaneous relationship between members of these suites is also suggested by the overlapping ages for some of the Mg suite rocks with the ferroan-anorthosite suite (Shih et al. 1993).

These models are not consistent with all observations made for the Mg suite. First, the ages for many Mg suite rocks clearly postdate the ferroan anorthosites that were derived from magma ocean crystallization (Papanastassiou and Wasserburg 1975; Carlson and Lugmair 1981a, 1981b; Nyquist and Shih 1992; Shih et al. 1993). Some of the apparent overlap of ages between the ferroan anorthosites and the Mg suite is the combined result of an early transition between magma ocean and post-magma ocean serial magmatism (Shearer and Newsom 1998) and an inaccurate interpretation of the crystallization ages for these early crustal rocks (Papike et al. 1998). Second, detailed geochemical characteristics of the two magmatic suites are not consistent with a near contemporaneous origin of the two magmatic suites from the magma ocean. For example, the Mg suite parental magmas has high initial Sr, Sm, and Eu/Al, but low Sc/Sm and Ti/Sm relative to the melts parental to the ferroan anorthosites (Norman and Ryder 1979; Raedeke and McCallum 1980; Warren and Wasson 1980; James and Flohr 1983; Warren 1986).

Because of these difficulties, numerous models have been proposed that advocate a post-magma ocean origin for the Mg suite that involves the melting of magma ocean cumulates. These models call upon either the melting of shallow, evolved magma ocean cumulates or melting of deep, magma ocean cumulates with incorporation of KREEP or crustal component through assimilation or mixing. Hess et al. (1978), Hess (1989), Snyder et al. (1995), Papike (1996) and Papike et al. (1994, 1996, 1997) explored the possibility that the Mg suite and temporally associated highland basaltic magmas (i.e., KREEP basalts, high alumina basalts) originated by partial melting of magma ocean cumulates at a shallow depth (Fig. 8c). Although they did not draw a genetic connection between the Mg suite and KREEP basalts, Hess et al. (1978) suggested that the KREEP-like highland basalts were generated by partial melting of magma ocean cumulates that crystallized soon after extensive ilmenite crystallization, but prior to the formation of KREEP (between 95 and 99% crystallization of the magma ocean). These magmas may have assimilated KREEP. Alternatively, Hess (1989) suggested that the KREEP-rich magmas were a product of partial melting of a lower lunar crust that had been altered metasomatically by KREEP. Pressure release melting and remobilization of these rock types may be related to catastrophic impacts on the lunar surface (Snyder et al. 1995; Papike 1996; Papike et al. 1994, 1996, 1997). Although derivation of the parental KREEP basaltic magma from very-late magma ocean cumulates may be consistent with some of the reconstructed trace-element melt compositions from individual minerals (Papike 1996; Papike et al. 1994, 1996, 1997), the chemical signatures common in the more-primitive Mg suite [olivine-bearing, Mg', Ni, (Na/Na+Ca)] cannot be accounted for by this model.

Models that require the Mg suite to represent crystallization products of high Mg' magmas generated in the deep lunar mantle were developed to resolve some of the problems with a shallow cumulate source (James 1980; Warren and Wasson 1980; Longhi 1981, 1982; Morse 1982; Ryder 1991; Smith 1982; Hunter and Taylor 1983; James and Flohr 1983; Shirley 1983; Shervais et al. 1984; Warren 1986; Hess 1994.) (Fig. 8d). Is the lunar mantle capable of producing primary magmas with

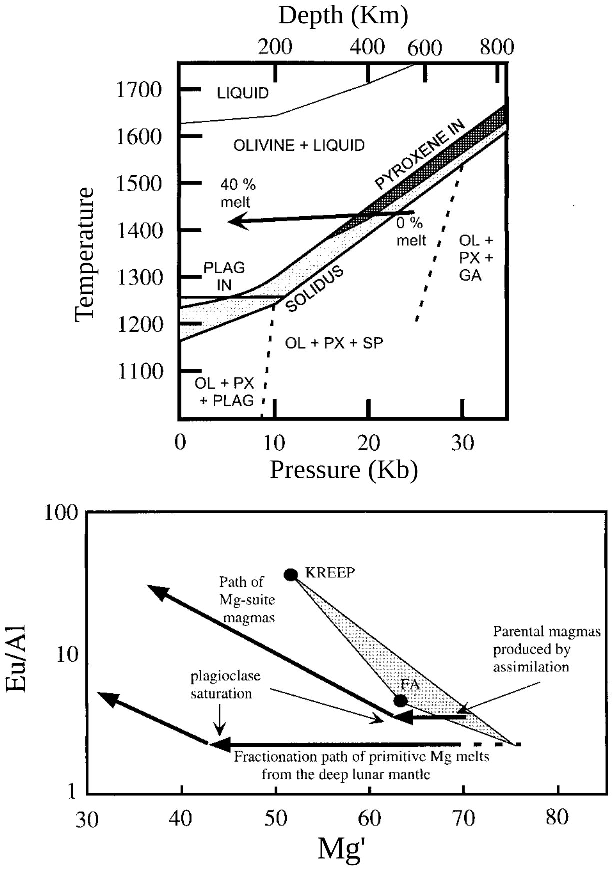


FIGURE 9. Diagrams illustrating the petrogenesis of the Mg suite parental magmas. (**a**) *P-T* diagram from Ringwood (1975) illustrating the effect of decompressional melting on producing Mg suite parental melts through high degrees of partial melting of deep magma ocean cumulates. (**b**) Mg' vs. Eu/Al illustrating the need for assimilation of KREEP or ferroan anorthosite (FAN) to produce the Mg suite magmas. Fractional crystallization of a simple magma ocean cumulate will not have Eu/Al or plagioclase saturation for the appropriate Mg'. An assimilation path will produce appropriate Eu/Al and plagioclase saturation at high Mg' (Warren 1983; 1986). Without assimilation of FAN, a hybrid source is required.

Mg' >74? Do these magmas have Al contents that are appropriate for plagioclase saturation at Mg' >74? Can assimilation processes increase the Al and incompatible-element content in these primitive magmas without dramatically lowering the Mg'? Hess (1994) demonstrated that magmas with Mg' appropriate for Mg suite parent magmas could be generated by melting of early magma ocean cumulates. A magma ocean with an Mg' value equivalent to the bulk Moon (80 to 84; Jones and Delano 1989; O'Neill 1991) upon crystallization at high pressures would produce early cumulates of olivine having Mg' greater than 91. Subsequent melting of these cumulates would produce magmas with Mg' equivalent to that of the parental Mg suite magmas. Because of the pressure dependence of the FeO-MgO exchange equilibrium between olivine and basaltic melt, crystallization of these high-pressure melts near the lunar surface would result in liquidus olivine that is slightly more magnesian than residual olivine in the mantle source. Higher Mg' values for the melt may result from the reduction of small amounts of FeO to Fe in the source (Hess 1994).

Melting of the early magma ocean cumulates initially could have been triggered by either radioactive decay (Hess 1994) and/or cumulate overturning (Hess and Parmentier (1994). The deep lunar mantle materials would have been less dense than the overlying cumulates and tend to move upward and be subjected to pressure-release melting. The pressure release from 400 km to 100 km would have been on the order of 15 kbars. This resulted in relatively high degrees of melting (>30%; Ringwood 1976; Herbert 1980) (Fig. 9a). Partial melting of early magma ocean cumulates could have produced primitive melts with high Mg' but, as shown in Figure 9b, these magmas would not have the same geochemical characteristics as the highland Mg suite. For example, these primitive magmas would not possess the high incompatible-element enrichments, fractionated Eu/Al and Na/(Na+Ca), and plagioclase as a liquidus phase until the Mg' of the melt was <42. Two types of processes have been proposed to resolve this problem: assimilation of evolved crystallization products of the magma ocean and melting of a hybrid cumulate sources.

In assimilation models, Warren (1986) calculated that if these high-Mg magmas assimilated ferroan anorthosite and KREEP, they would have reached plagioclase saturation at values of Mg' appropriate for Mg suite magmas. Such magmas also would have inherited a fractionated incompatible-element signature (high REE, fractionated Eu/Al). Hess (1994) explored the thermal and chemical implications of anorthosite melting and plagioclase dissolution by high-Mg basaltic magmas. In his analysis of anorthosite melting and mixing as a mechanism to drive high Mg' magmas to plagioclase saturation, he concluded that the resulting crystallization of olivine and the mixing of relatively Mg-poor cotectic melts produced from the anorthosites would lower the Mg' of the hybrid melt below that expected for the Mg suite parental magmas. In addition, diffusion rates for Al_2O_3 in basaltic melts are extremely slow (Finnila et al. 1994) and indicate that the time scales to dissolve even a small amount of plagioclase are of the same order as the characteristic times of solidification of a large magma body. Similar thermal constraints are less severe for the shallow melting-assimilation of a KREEP component into a primitive Mg suite magma. Although assimilation of KREEP does not dramatically drive the Mg-rich magmas to plagioclase saturation, this mechanism may account for the evolved trace-element signatures in the Mg suite. However, mixing of viscous melts of KREEP composition with more-fluid Mg-rich magmas could be prohibitive (Finnila et al. 1994). In addition, melt compositions for the Mg suite norites that were calculated from pyroxene trace element data have a KREEP component equivalent to or slightly higher than KREEP (Papike et al. 1997). Simple mass-balance calculations indicate that it is impossible for a primitive Mg suite magma to assimilate such an abundant amount of KREEP (Shearer and Floss 1999).

As an alternative to assimilation at shallow mantle levels, is it possible that the KREEP component was added to the deep mantle source for the Mg suite (Fig. 8e)? Hess (1994) proposed that the source for the Mg suite may be a hybrid mantle consisting of early magma ocean cumulates (dunite) and a bulk Moon component that may be either primitive lunar mantle or an early quenched magma ocean rind. Polybaric fractional melting of a 50–50 mixture would produce melts with appropriate Mg' and reasonable Al_2O_3. However, these mixing components would not produce some of the incompatible-element signatures exhibited by the Mg suite. Shearer et al. (1991) suggested a cumulate overturn mechanism to transport a KREEP component to the deep lunar mantle to explain the evolved KREEPy signature imprinted on selected picritic glasses associated with mare basaltic magmatism. A similar process may have produced the KREEP signature in the Mg suite.

A potential pitfall of this model is the Ni content of the olivine in Mg suite rocks. Although the Mg' of olivine ($Fo_{92–88}$) in the Mg suite is high and suggests a primitive magma derived from deep within the LMO cumulate pile, the Ni content of the olivine determined by microprobe (Ryder 1983) ranges from 10 to 320 ppm. This is fairly low for a primitive lunar basalt. In comparison, ion microprobe analyses of olivine cores ($Fo_{75–72}$) from Apollo 12 olivine basalts yield Ni concentrations of 450 to 515 ppm (Papike et al. 1999). The calculated olivine composition in equilibrium with mare basalts more primitive than the Apollo 12 olivine basalts (i.e., Apollo 15 green glass) is approximately Fo_{85} and Ni ~ 1500 ppm. This poses a dilemma for interpretation of the high Mg' of the Mg suite and hints at the involvement of a metal phase either during melting of the source or during evolution of the basaltic magma (Papike et al. 1997).

Models for the generation of the alkali suite

The alkali suite may be part of a continuum of crystallization products of parental basaltic magmas similar to the KREEP basalts (Fig. 10a) from the Apollo 15 landing site (Ryder 1976; Warren and Wasson 1980; James 1980; Hunter and Taylor 1983; James et al. 1987; Hess 1989; Marvin et al. 1991; Snyder et al. 1995); the origin of KREEP basalts follows in the discussion of pre-mare basaltic volcanism. Using a KREEP basalt (from the Apollo 15 collection site) as a starting parental magma, Snyder et al. (1995) demonstrated that fractional crystallization and the accumulation of mineral phases and trapped KREEP-like residual liquid (2–15%) could produce the range of mineral and rock compositions observed in the highland Mg and alkali suites. The sequence of crystallization and accumulation that they proposed is Mg suite (0 to 43% crystallization) → alkali anorthosites, alkali norites (43 to 74% crystallization) → alkali gabbros, alkali norites (74 to 90% crystallization) → quartz monzodiorites (90 to 99.8% crystallization) → granites. The absence of Mg suite rocks with ages that correspond to the younger episodes of KREEP basaltic magmatism is possibly a result of a lack of deep sampling because of a decrease in impact flux after 3.9 Ga. The lower Mg' of KREEP basalts (~61–66) relative to the Mg' that would be expected for parental magmas for the Mg suite cumulates (mafic minerals with Mg' ~ 90) is potentially a result of sampling (Irving 1977).

Alternatively, the Mg suite and alkali suite may represent contemporaneous, yet separate episodes of basaltic magmatism (Warren and Wasson 1980; James 1980; James et al. 1987). There is some compositional evidence to suggest genetically distinct highland rock types. For example, James et al. (1987) subdivided many of these highland rock types into various groups on the basis of their mineral chemistry and mineral associations. Whether these subdivisions are artificial or petrologically significant is debatable. Within this scenario, difference between these two suites may be attributed to the depth of initial melting prior to assimilation. For example, Mg suite magmatism would be a product of deep-mantle melting followed by KREEP assimilation just below the lunar crust, whereas the alkali suite magmatism would involve initial melting at shallower mantle levels followed by assimilation (Fig. 10b).

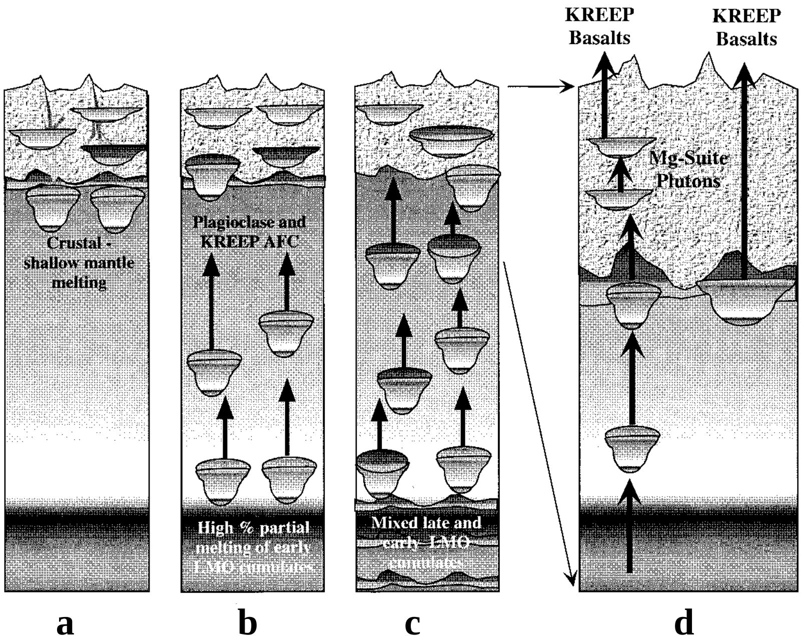


FIGURE 10. Possible models illustrating the relationship between Mg suite magmatism and alkali suite magmatism.

Materials of granitic composition occur as immiscible glasses within the mesostasis of mare basalts, in melt inclusions, as glasses in selective lunar soils, and as crystalline clasts in breccias. In all petrologic models, the granite, rhyolite, and felsite clast are considered to be products of extensive fractional crystallization of alkali suite parental magma. Marvin et al. (1991) proposed that these evolved granitic rock types were produced through the crystallization and removal of phosphates and zircon from parental magmas equivalent to the quartz monzodiorite of the highland alkali-suite. This fractional crystallization mechanism accounted for the enrichment of REE, Zr, and Hf in the quartz monzodiorites relative to the more evolved and lunar granites. It has been also proposed that the lunar granites are a product of liquid immisciblity following extensive fractional crystallization of a KREEP basalt. Experimental studies by Hess et al. (1975) illustrated that after 90–95% of crystallization of various lunar basaltic magmas, spherules of granitic magma exsolved from the residual ferrobasaltic liquid. Chemical signatures of these lunar granites have been attributed to immiscibility (Quick et al. 1977; Taylor et al. 1980; Warren et al. 1983; Neal and Taylor 1989). On the other hand, Jolliff (1991) and Jolliff et al. (1993) argued that many of these chemical signatures were a result of phosphate crystallization prior to silicate liquid immiscibility.

STAGE 4: PRE-BASIN VOLCANISM—VOLCANISM ASSOCIATED WITH EARLY REMELTING OF MAGMA OCEAN CUMULATES

Introduction

Although the petrologic record has been obscured by the early catastrophic impact history of the Moon, there is abundant evidence of pre-3.9 Ga non-mare basaltic volcanism. Most of this record is retained in clasts from highland soils and breccias (Dickinson et al. 1985; Shervais et al. 1985; Neal et al. 1988; 1989a, 1989b) or identified through remote sensing (Schultz and Spudis 1979; Bell and Hawke 1984; Hawke et al. 1990). Two types of pre-mare volcanism have been identified thus far: KREEP basaltic volcanism and high-alumina basaltic volcanism. Unlike younger episodes of basaltic magmatism (mare basaltic magmatism), the pre-mare KREEP basalts have relatively high concentrations of Al_2O_3 and incompatible trace elements. The pre-mare, high-alumina basalts are mare basalt-like in composition and texture. They have a variety of trace-element characteristics ranging from LREE-depleted to LREE-enriched (Shervais et al. 1985; Dickinson et al. 1985). Numerous lines of evidence such as igneous textures, lack of iron-metal particles, low siderophile trace element abundances, and non-meteoritic siderophile element ratios indicate that clasts representing both pre-mare basalt types are volcanic and not impact derived (James 1980). Distribution and abundance of both rock types are difficult to quantify. However, numerous studies imply that large volumes of KREEP and high-alumina basaltic magmas “cryptomaria” are present (Metzger and Parker 1979; Davis and Spudis 1985, 1987; Head and Wilson 1992). Head and Wilson (1992) have suggested that perhaps up to a third of the erupted basalts at the lunar surface was this type of cryptomaria volcanism.

The KREEP basalts thus far sampled have crystallization ages of 3.85–4.10 Ga (Nyquist and Shih 1992). Several models considered above imply that older, highland rock types (Mg suite and alkali suite) are preserved early remnants of KREEP basaltic magmatism. The ε_{Nd} values of KREEP basalts and highland Mg suite rock types are consistent with a cogenetic relationship (Nyquist and Shih 1992). If true, generation, emplacement, and eruption of KREEP basaltic magmas may have extended for a much longer period of time (~600 m.y.). KREEP basaltic volcanism is thought to be primarily pre-basin, pre-mare, highland volcanism. However, some interpretations of basin formation ages (Deutsch and Stoffler 1987; Stadermann et al. 1991) imply that later KREEP eruptive episodes may have been triggered by basin formation. Nyquist and Shih (1992) speculated that the melting episodes that generated both the KREEP basalts and the high-alumina basalts were triggered by the early impact of a large bolide.

The high-alumina basalt clasts from the Apollo 14 site have ages that range from 4.0 to 4.3 Ga (Shih et al. 1992; Nyquist and Shih 1992), some 200 m.y. older than the oldest known KREEP basalt. However, the presence of high-alumina clasts in breccias from the Fra Mauro Formation suggests that their provenance was similar to the KREEP basalts. With similar provenance and the possibility that KREEP basalts are parental to highland plutonic suites, the KREEP and high-alumina magmatism and volcanism may have been contemporaneous (Nyquist and Shih 1992).

Generation of pre-mare KREEP basalts

Several models have been developed that explain the generation of KREEP basaltic magmas either through assimilation of KREEP by Mg-rich parental magmas or by melting of a hybridized or non-hybridized lunar mantle or crust (Fig. 11a–c). The relationship of the KREEP basalts to Mg suite plutonism is still uncertain (Fig. 11d). Models by Taylor (1975), Hess et al. (1978) and Hess (1989) require the source for the KREEP basalts to be shallow. They suggested that the KREEP basalts were generated by partial melting of the magma ocean cumulates that crystallized soon after extensive ilmenite crystallization, but prior to the formation of KREEP (between 95 and 99% crystallization of the magma ocean). Magmas produced by the melting of this cumulate source then assimilated KREEP. Hess (1989) later modified this model to suggest the KREEP basalts could have been a product of partial melting of a lower lunar crust that was metasomatized by KREEP (Fig. 11a).

Warren (1988) modeled the KREEP basalts as a product of assimilation and fractional crystallization involving KREEP and “Mg-rich” parental magmas (Fig. 11b). This is similar to processes suggested for magmas parental to the various highland plutonic suites. It would be expected that the assimilation of KREEP by different batches of Mg-rich parental magmas would produce random I_{Sr} and ε_{Nd} values. However, Nyquist and Shih (1992) argued that the I_{Sr} and ε_{Nd} values are not consistent with assimilation processes because they varied progressively with time.

Ryder (1976) argued that the KREEP basalts were products of partial melting of distinct source regions and that assimilation was not a process prominent in their origin. Such a model requires the KREEP trace-element signature to be a character-

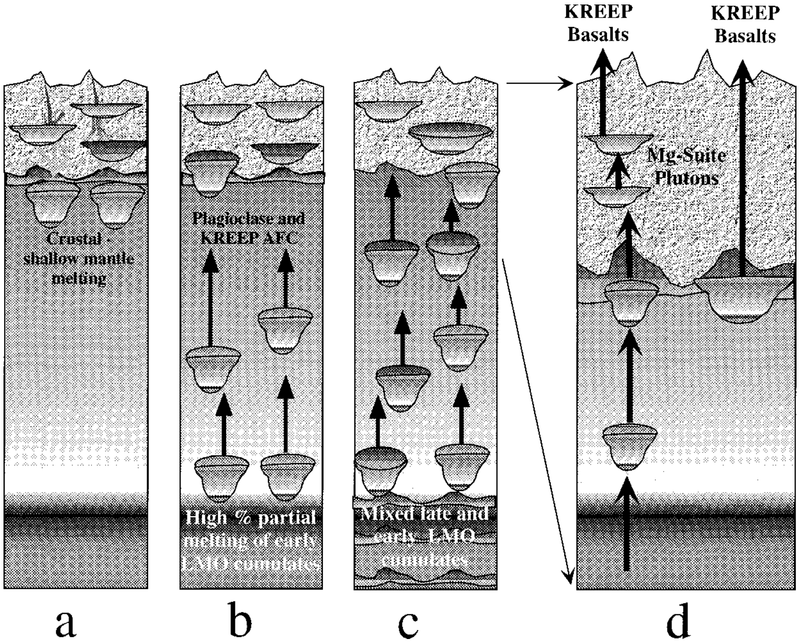


FIGURE 11. Models for the generation of KREEP basalts. (**a**) Shallow melting. (**b**) Deep melting with KREEP assimilation. (**c**) Deep melting of a mixed cumulate pile. (**d**) Possible relationships between KREEP basalts and Mg suite plutonic rock types.

istic of the source and implies a hybridized mantle. Nyquist and Shih (1992) proposed that the observed variations in I_{sr} and ε_{Nd} values were compatible with the systematic melting of a hybridized lunar mantle to produce KREEP basalts (Fig. 11c).

Generation of pre-mare, high-alumina basaltic magmatism

Models for the generation of the high-alumina basalt group range from melting of hybridized and nonhybridized mantle sources (Reid and Jakes 1974; Taylor and Jakes 1974; Binder 1982, 1985; Kurat et al. 1976; Ma et al. 1979; Dickinson et al. 1989; Hughes et al. 1988, 1989; Shervais and Vetter 1990) to assimilation of an evolved LMO component (KREEP, lunar granite rock types) by mantle-derived basaltic magmas (Dickinson et al. 1985; Shervais et al. 1985; Neal et al. 1988, 1989a, 1989b). These various models are similar to those proposed for the KREEP basalts (Fig. 11), but involve different sources and assimilants. All of these models are compromised by the small size of many of the high-alumina basalt fragments which may introduce some of the chemical variations due to their non-representative nature.

Early studies, based primarily on high-alumina basalts returned by the Luna 16 mission placed the source for the high-alumina basalts just below the crust at depths of 40–100 km (Reid and Jakes 1974; Taylor and Jakes 1974; Binder 1976; Kurat et al. 1976). In these studies, the source region was thought to be a late-stage magma ocean cumulate rich in clinopyroxene and plagioclase (Reid and Jakes 1974; Taylor and Jakes 1974; Binder 1976; Ma et al. 1979) and was LREE-enriched (Ma et al. 1979). The wide range in incompatible trace-element abundances documented in later studies of high-alumina basalt clasts from the Apollo 14 site was shown to be inconsistent with these models (Shervais et al. 1985; Dickinson et al. 1985; Dickinson et al. 1989; Neal et al. 1988; 1989a, 1989b; Hughes et al. 1990; Shervais and Vetter 1990). Rather, the variability was ascribed to different degrees of partial melting and KREEP assimilation (Dickinson et al. 1985; Shervais et al. 1985; Neal et al. 1988, 1989a, 1989b). The best-fit model by Neal et al. (1988; 1989a, 1989b) and Neal and Taylor (1990, 1992) employed cyclical assimilation of KREEP by a LREE-depleted olivine basalt. They concluded that assimilation of up to 15% KREEP and 70% fractional crystallization would produce the array of compositions observed in a suite of high-alumina basalts from the Apollo 14 site. Nyquist and Shih (1992) modified this model by suggesting that there were multiple batches of olivine basaltic magmas produced by different degrees of partial melting and that the parental olivine basalt assimilated KREEP prior to extensive fractional crystallization. Although assimilation-fractional crystallization (AFC) models are consistent with major and trace element data, they are not compatible with isotopic data. AFC processes involving basaltic magmas and KREEP ought to yield a variety of initial isotopic compositions, which are not observed (Nyquist and Shih 1992).

As an alternative to AFC, Hughes et al. (1990) suggested that the incompatible trace-element variability could be attributed to partial melting of a hybridized source consisting of a mixture of early and late-stage magma ocean cumulates. This model appears to agree with the Ge abundances of the high-alumina basalts (Dickinson et al. 1989) and the variations in initial isotopic compositions (Nyquist and Shih 1992). Whereas Hughes et al. (1990) suggested that the mechanism for mantle mixing and source hybridization was the gravitational destabilization of the cumulate pile, Nyquist and Shih (1992) suggested that the mechanism for cumulate mixing was the impact of a large bolide. Nyquist and Shih (1992) based their interpretation on the seemingly cyclical and localized nature of high-alumina basaltic magmatism (Dasch et al. 1987; Neal and Taylor 1992; Nyquist and Shih 1992). These two cumulate mixing models are not only different in the mechanism of mixing, but they also differ with respect to the depth of the mantle cumulate source. The gravitational destabilization of the cumulate pile implies a deep (>400 km) mantle source for the high-alumina magmas, whereas impact mixing implies a shallow source (<100 km).

STAGE 5: MARE BASALTS—LATE REMELTING OF MAGMA OCEAN CUMULATES

Distribution and composition of the mare basalts

Mare basalts are exposed over 17% of the lunar surface and are thought to make up ~1% of the lunar crust. They primarily fill multi-ringed basins and irregular depressions on the Earth-facing hemisphere of the Moon. This distribution has been attributed to the thicker highland crust on the far side of the Moon and the gravitational attraction of the Earth. The thickness of these flood basalts range from 0.5 km to 1.3 km in irregular basins to 4.5 km in the central portions of younger basins (Bratt et al. 1985). The general style of volcanic activity was the eruption of large volumes of magma from relatively deep sources (not shallow crustal reservoirs) with very high effusion rates (Head and Wilson 1992). Evidence of lunar fire fountaining is preserved in the form of spherical glass beads (Reid et al. 1973; Meyer et al. 1975; Delano 1979, 1980, 1986; Shearer and Papike 1993) and dark mantling deposits (Wilhelms and McCauley 1971; Head 1974; Head and Wilson 1992).

Within the context of terrestrial basalt classification (i.e., basalt tetrahedron: Yoder and Tilley 1962; Yoder 1976), lunar basalts range from quartz to olivine normative. The volcanic glass beads are more Mg-rich than the crystalline mare basalts and are all olivine normative. Nepheline normative lunar basalts have not yet been identified.

The mare basalts exhibit numerous distinct mineralogical and chemical characteristics: (1) a spectacular range in TiO_2; (2) mineralogical (Fe metal) and chemical (reduced valance states for Fe, Ti, Cr) signatures reflecting extremely low oxygen fugacities (~ IW buffer); (3) depletions in alkali, volatile, and siderophile elements; (4) absence of water and hydrous phases; and (5) a negative Eu anomaly in all crystalline mare basalts and picritic glasses (BVSP 1981; Taylor 1982; Newsom 1986; Shearer and Papike 1993; Papike et al. 1998).

The compositional diversity of the lunar mare basalts is partially illustrated in a plot of TiO_2 vs. Mg' (Fig. 12). As an approximation, the lunar basalts form three distinct chemical clusters: a high-TiO_2 cluster (TiO_2 = 9–14 wt%), a low- to intermediate-TiO_2 cluster (TiO_2 = 1–5 wt%), and a very low-TiO_2 cluster (<1 wt%). In general, from the very low TiO_2 to the high TiO_2 mare basalts there is an increase in the abundance of incompatible-elements (REE, Zr, Ba, Nb), an increase in the depth of the negative Eu anomaly, and a decrease in the abundance of compatible elements (e.g., Ni). The variability in Mg' (Fig. 12) predominantly reflects the effect of olivine and pyroxene fractionation within individual TiO_2 suites of mare basalts. The scarcity of basalts in the TiO_2 range of 5 to 9% is a result of sampling. Remote sensing studies indicate that only one third of mare basalts are represented in our sample suite (Pieters 1978, 1990). In addition, while the lunar sample suite suggests a bimodal abundance of low- and high-TiO_2 basalts, remote sensing measurements show that the TiO_2 contents of mare basalts do not form a bimodal distribution and that basalts with intermediate TiO_2 contents are more abundant than are those with high TiO_2 contents (Giguere et al. 1999).

Using major- and trace-element characteristics of picritic glass beads, Delano (1986) defined 25 glass groups. Volcanic glass beads exhibit a wide range in TiO_2 (0.2–17 wt%), yet are consistently higher in MgO and Mg' in comparison with crystalline mare basalts having similar TiO_2 contents (Fig. 12). The higher Mg' of the picritic glasses indicate that they are the best candidates for primary mare basaltic magmas and that most of the crystalline mare basalts experienced fractional crystallization. The picritic glasses also imply that a range of primary melts is responsible for the diversity of mare basalts and that the sources for the basaltic magmas are fractionated (e.g., negative Eu anomaly).

Eruptive history of the mare basalts

The radiometric ages for sampled mare basalts range from 4.0 to 3.16 Ga (as summarized by Nyquist and Shih 1992). Schaber (1973) estimated the ages of three flows to be 3.00 ± 0.04, 2.70 ± 0.03, and 2.50 ± 0.03 Ga on the basis of crater-counting methods. Crater-density statistics indicate that the youngest basaltic units on the Moon may be between 0.9 and 2.0 Ga (Schultz and Spudis 1983). The eruptive flux was not constant, and a majority of mare basalts were emplaced between 3.8 and 3.2 Ga. The 3.8 Ga age corresponds roughly with the rapid falloff of meteorite bombardment and with the excavation of the Imbrium basin (Fig. 13). Head and Wilson (1992, 1997) suggested that the peak eruptive flux of mare basalts at 3.8 Ga may have been a continuation of "cryptomare" volcanism (KREEP basalts, high-alumina basalts). Some eruptions during the peak eruptive flux may have lasted on the or-

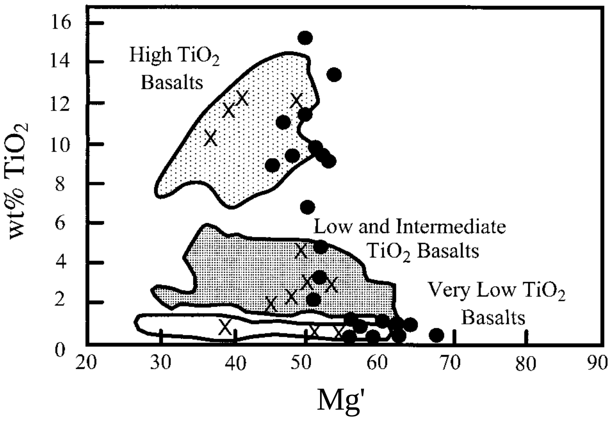


FIGURE 12. Compositional variations observed in mare basalts illustrated in a plot of Mg' vs. TiO_2. Fields for mare basalts are from Neal and Taylor (1992). Picritic glasses are shown in filled circles. Crystalline mare basalts that approximate liquid compositions are represented by X.

der of one year and may have emplaced 10^3 km^3 of basaltic lava (Head and Wilson 1992, 1997). Following this extensive eruptive pulse, eruptions were of low volume and episodic, and account for <5% of the total volume of mare basalts (Schaber 1973; Schultz and Spudis 1983; Head and Wilson 1992, 1997). Samples returned by Apollo and Luna missions present an ambiguous picture concerning the relation between magma composition and time (Papike et al. 1976). Head and Wilson (1992, 1997) concluded that, although various magma types were being erupted in nearside mare basins during the period 3.8 to 3.2 Ga, the early and intermediate phases of eruption were dominated by high-Ti basalts. Later periods of eruption within a basin are predominately low-Ti basalts.

Models for the origin of mare basalts

All models for the generation of mare basalts concur that fractional crystallization was an important process in producing diverse basalt compositions. The models differ in their concept of the role that early lunar differentiation played in the formation of the mantle sources for the basalts, the subsequent evolution of these sources, and the conditions under which melting occurred. Fundamental source models are (1) primitive source models, (2) magma ocean cumulate source models, and (3) serial magmatism source models (Fig. 14). Based on our earlier discussions, the source for the mare basalts is clearly differentiated on the basis of its nonchondritic composition. Therefore, a primitive source for the mare basalts is not appropriate.

To account for this differentiated signature of the mare basalts, the magma ocean cumulate source model was advanced

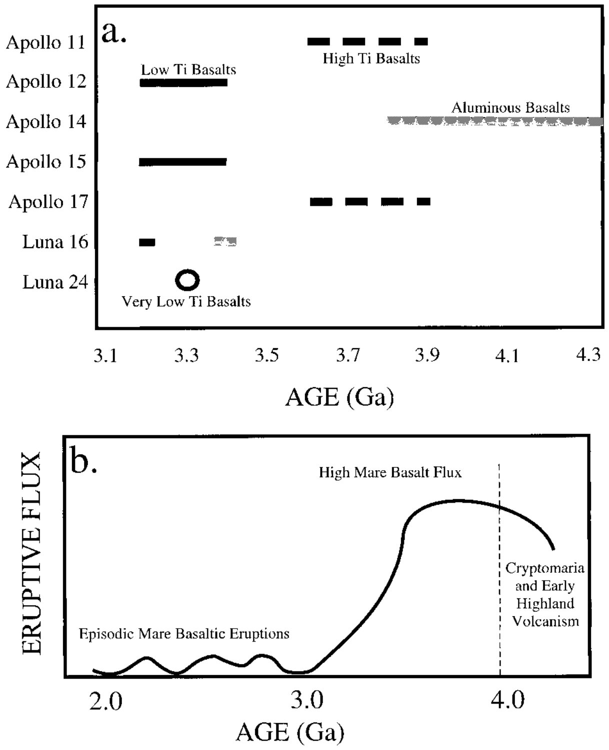


FIGURE 13. (**a**) Ages of mare basalts from various sampling sites (from Nyquist and Shih 1992). (**b**) Estimate of mare basalt eruptive flux (from Head and Wilson 1992, 1997).

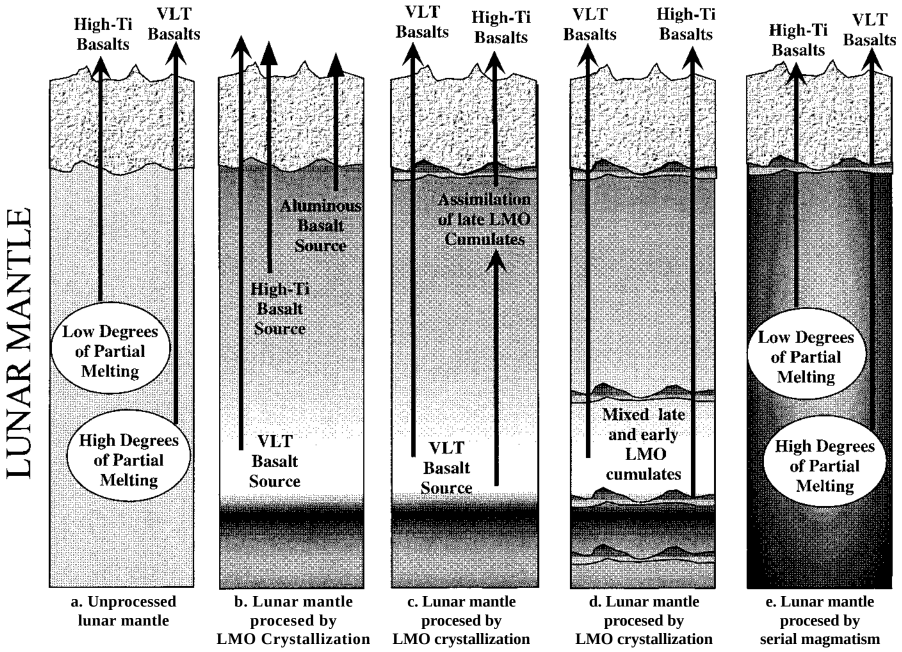


FIGURE 14. Models for the generation of lunar mare basalts. (**a**) Melting of a primitive unprocessed lunar mantle. (**b**) Melting of the LMO cumulate pile. (**c**) Melting of deep LMO cumulates producing VLT basaltic magmas. The high-Ti basalts are produced by the assimilation of late-stage LMO cumulates by the VLT basaltic magmas. (**d**) Melting of a disrupted LMO cumulate pile. (**e**) Melting of a lunar mantle that was processed by serial magmatism.

relatively early in the Apollo program and substantially matured through many other studies (Wood 1972, 1975; Taylor and Jakes 1974; Solomon and Longhi 1977; Longhi 1977; Warren 1985, 1986). In the Taylor and Jakes model (1974), early differentiation of the Moon through extensive melting resulted in the formation a layered mantle composed of cumulates. The source of the low-Ti mare basalts was composed of relatively early formed ultramafic cumulates (predominantly olivine and low-Ca orthopyroxene). Late-stage cumulates, which contain olivine, pyroxene, and ilmenite, were proposed as a source for the high-Ti basalts. Plagioclase flotation during crystallization of the LMO imposed a negative Eu anomaly on the crystallizing mafic cumulates and produced a complementary positive Eu anomaly in the feldspathic lunar crust. In addition to explaining the negative Eu anomaly in the mare basalts, it also placed into a single conceptual model the observations that the source for the mare basalts was differentiated, and that the sources for the various mare basalts were isotopically, chemically, and mineralogically distinct.

A model in which the source for mare basalts is analogous to a terrestrial layered intrusion has major problems. First, the multiple-saturation pressures for both high-Ti (late cumulate source) and low-Ti basalts (early cumulate source) overlap indicating that the basalts were derived from similar depths in the mantle (Delano 1979, 1980, 1986). Second, there is a lack of significant differences in Mg' and Cr concentrations between the most primitive of the high- and low-Ti basalts. Third, magmas derived from early cumulates may possess evolved, late-stage cumulate signatures (KREEP) (Shearer et al. 1990, 1991**)**. Numerous modifications to the simple layered intrusion analogy have been made in response to these objections. These models invoke either assimilation of evolved cumulates by magmas produced deep within the cumulate pile, or partial melting of hybridized mantle sources created by the redistribution of the cumulate pile.

Hubbard and Minear (1975) and Ringwood (1975) called upon assimilative models to account for some of these discrepancies in the layered cumulate source model. In these models, low-Ti basaltic liquids assimilate to varying degrees the high-Ti cumulates in the shallow portion of cumulate pile. The wide compositional spectrum of mare basalts was a result of varying degrees of assimilation, the compositional diversity in high-Ti cumulates, and composition of the low-Ti magmas. Several studies have shown that the assimilative models have problems explaining the thermal requirements of assimilating large amounts of high-Ti cumulate material, overlapping multiple saturation pressures for both high-Ti and low-Ti basalts, and accounting for variations in abundances of the trace elements Sc, Ti, Zr, Nb, Ce, and Ni (Ringwood and Kesson 1976; Papike et al. 1976; Taylor 1982; Shearer et al. 1996b). However, recent experimental studies have breathed new life into the assimilation model (Wagner and Grove 1995). These assimilation experiments involving very low-Ti basaltic melts and pyroxene-ilmenite assimilants question the thermal arguments against assimilation made by Ringwood and Kesson (1976). They demonstrated that the dissolution rate of ilmenite is faster than diopside and concluded that ilmenite is preferentially assimilated into the melt. This observation thereby resolves at least some of the chemical arguments that were made against assimilation. Several arguments against this model still remain. First, the mass of assimilation not only depends on the relative rate of dissolution, but also the diopside/ilmenite ratio in the cumulate. Second, ilmenite in the ilmenite-cumulate horizon should have highly frationated Nb/Ce (2000) and Nb/Zr (50) (Shearer et al. 1996b). Assimilation of ilmenite by low-Ti magmas should dramatically increase those values in the resulting melt. This is not the case. The high-Ti picritic glasses have only a slight increase in these ratios compared to the low-Ti picritic glasses (Shearer et al. 1996b; Shearer and Papike 1999). Finally, Beard et al. (1998) concluded that the difference in eHf vs. eNd trends for high- and low-Ti mare basalts is not consistent with assimilation of high-Ti cumulates to produce the complete array of Ti observed in mare basalts.

As an alternative to assimilating high-Ti cumulates at shallow mantle levels, Ringwood and Kesson (1976) argued that the dense, late-stage Ti- and Fe-enriched cumulates sank into deeper portions of the cumulate pile. Exchange of material, different degrees of partial melting, and different domains of partial melting of a heterogeneous source produced a wide spectrum of chemically distinct basaltic magmas (Hughes et al. 1988, 1989; Ryder 1991; Shearer et al. 1990, 1991, 1996a, 1996b; Shearer and Papike 1993; Hess and Parmentier 1996; Shearer and Newsom 1998). This mechanism resulted in the transport of the more evolved components of the LMO cumulate pile (KREEP, high-Ti cumulates) to zones of melting and provided a heat source (decay of incompatible, radioactive elements) to initiate melting at depth.

Spera (1992) and Hess and Parmentier (1995) have modeled large-scale overturn of the cumulate pile. Snyder et al. (1992) suggested a much more localized overturn of the cumulate pile. Although localized overturning of magma ocean cumulates may account for some of the chemical variability observed in distinct batches of magma (Shearer et al. 1996a), catastrophic overturn is required to account for numerous chemical and mineralogical signatures (e.g., KREEP component in primitive magmas and high multiple-saturation pressures for the primitive high-Ti basaltic magmas).

As an alternative to a mare basalt source that was processed by magma ocean formation and crystallization, Walker (1983) suggested that the source region for the mare basalts was produced during early lunar differentiation by serial magmatism. In this model, mafic cumulates from large layered intrusions were separated from their anorthositic cap and transported by convection into the deep lunar mantle where they were mixed with undifferentiated mantle and formed the mare basalt source region. As a result, the contrasting Eu anomalies observed in the ferroan anorthosite crust and mare basalts are inconsequential.

Walker (1983) and Haskin (1989) suggested that the bulk lunar crust does not exhibit a positive Eu anomaly predicted by the cumulate model. In addition, Walker (1983) suggested the negative Eu anomaly exhibited by mare basalts was not a characteristic of the lunar mantle. Rather, it reflected the mixing of primitive magmas with basaltic magmas that had experienced plagioclase fractionation. These conclusions are suspect. First, the bulk lunar crust as estimated by Walker (1983) and Haskin (1989) includes igneous rocks produced during and following magma ocean crystallization. Therefore, even if the bulk lunar crust does not exhibit a positive Eu anomaly, the products of LMO crystallization (ferroan anorthosites) most certainly do. Second, primary magmas represented by the picritic volcanic glasses all have negative Eu anomalies. There is no evidence that these magmas mixed with more evolved ones that had experienced plagioclase crystallization.

The very low-Ti basalts (VLT) (Vaniman and Papike 1977) are represented by small fragments of crystalline basalts and a suite of picritic glasses. The crystalline basalts show a wide range of compositions that have been attributed to extensive olivine fractionation from slightly different parental magmas (Taylor et al. 1978; Norman et al. 1978; Coish and Taylor 1978). Members of the VLT picritic glass suite have been suggested as being parental to the crystalline VLT basalts (Vaniman and Papike 1977; Ryder and Marvin 1978; Grove and Vaniman 1978; Taylor et al. 1977, 1978). However, it has been demonstrated through experimental studies (Grove and Vaniman 1978), major-element modeling (Longhi 1987), and trace-element modeling (Ma et al. 1978; Shearer et al. 1990; Shearer and Papike 1993) that these picritic glasses do not directly represent parental compositions for the crystalline VLT.

Most of the petrologic models for the crystalline VLT basalts require small degrees of partial melting (1–10%) of a mantle source consisting of early magma ocean cumulates dominated by olivine and orthopyroxene (Grove and Vaniman 1978; Ryder and Marvin 1978; Wentworth et al. 1979; Nyquist et al. 1977, 1978; Ma et al. 1978; Neal and Taylor 1992). Variable amounts of clinopyroxene and plagioclase might be present in the cumulate and residua assemblages (Nyquist et al. 1977, 1978; Ma et al. 1978). Experimental studies indicate that the pressure of multiple saturation, which has been equated to the depth of melting, was 5 kbars (100 km; Grove and Vaniman 1978). As these crystalline mare basalts are not primary magmas, source modeling and experimental estimates of depth of origin are significantly in error. Modeling of fractionated VLT magmas underestimates the Mg' and olivine content of the source and overestimates incompatible-element abundances and the importance of plagioclase and clinopyroxene in the source. In addition, the experimentally determined multiple-saturation pressure gives a minimum depth of melting. Using VLT picritic glasses that approach primary melt compositions is a more reasonable approach to evaluating source regions.

Delano (1986) defined 12 groups of VLT picritic glasses that he concluded approached primary melt compositions. Four sets of experiments on VLT picritic glass compositions indicate multiple-saturation pressures of between 17 and 22 kbars. This is equivalent to a minimum depth of melting of 370 km. Longhi (1992) proposed that these magmas may have been produced by polybaric melting and therefore this multiple saturation depth was only an average depth of melting. Melting may have initiated at depths as great as 1000 km (Longhi 1992). An alternative to the interpretation that multiple saturation depth was the depth of melting has been proposed by Binder (1982, 1985), who pointed out that at high degrees of melting the multiple saturation depth was meaningless because the melts would only be saturated with olivine. Therefore, these picritic magmas could potentially be generated from shallow mantle depths. Subsequent work by Hess (1993) and Hess and Finnila (1997) demonstrated that all the picritic glasses were multiply saturated with olivine and orthopyroxene and consequently were produced by low degrees of partial melting in the deep lunar mantle.

Trace-element studies of the VLT picritic glasses provide evidence supporting the concept that the sources for these picritic magmas were mixed magma ocean cumulates. Hughes et al. (1988, 1989) showed that selected VLT glass compositions were produced by small degrees of partial melting of a mixed cumulate source consisting of early magma ocean cumulates and a small proportion of late magma ocean cumulates. Shearer et al. (1989, 1990, 1991, 1994, 1996a, 1996b) and Shearer and Papike (1993) further confirmed the mixed source models. Both large- and small-scale mixing appeared to be important processes for generating VLT sources. They documented that extremely late-stage products of magma ocean crystallization (KREEP) were transported into the deep lunar mantle through cumulate overturn and incorporated into the source for some of the VLT magmas.

Several groups of low-Ti mare basalts have been recognized: olivine basalts, pigeonite basalts, ilmenite basalts, feldspathic basalts, and quartz normative basalts (Rhodes, 1972; James and Wright 1972; Rhodes and Hubbard 1973; Chappel and Green 1973; Walker et al. 1976; Papike et al. 1976; Rhodes et al. 1977; Baldridge et al. 1979). There appears to be no fractional crystallization process that relates these different petrologic types of basalts (Rhodes et al. 1977; Neal and Taylor 1992). Dynamic melting has been suggested as a possible mechanism for generating quartz-normative and olivine-normative basalts from the same source (Vetter et al. 1988; Vetter and Shervais 1989). However, Neal and Taylor (1992) suggested that these basalts were generated from two different sources. Closed-system fractional crystallization in a thick lava flow or shallow magma chamber is the main process responsible for intragroup mineral and chemical variation (Rhodes 1972; Rhodes and Hubbard 1973; Chappel and Green 1973; Rhodes et al. 1977; Walker et al. 1976).

Early studies concluded that low-Ti basalts were generated from a mantle source that approached a whole-Moon composition (Duncan et al. 1976). However, subsequent studies demonstrated that these basalts were also derived by small to moderate degrees of melting of magma ocean cumulates. Experimentally determined pressures of multiple saturation for the low-Ti basalts range from <5 kbars to >25 kbars (<100 km to >400 km; Green et al. 1971a, 1971b; Hodges and Kushiro 1972; Kesson and Lindsley 1974; Kesson 1975; Walker et al. 1976, 1977). Again, because these basalts deviate to varying degrees from primary basaltic compositions, these multiple saturation depths should be considered minimum depths of melting and segregation. Cumulate assemblages that were melted to produce the low-Ti basalts consisted of olivine + orthopyroxene + clinopyroxene ± plagioclase according to Ma et al. (1976, 1978) and Shih and Schonfeld (1976). Nyquist et al. (1977, 1979) and Neal and Taylor (1992) discounted the presence of plagioclase in the sources of mare basalts on the basis of isotopic and trace-element characteristics, and the probability that the source had formed prior to the onset of plagioclase crystallization during solidification of the LMO. The latter argument has limited validity in that cumulate mixing possibly occurred. Nyquist et al. (1977, 1979) suggested that the magma ocean was LREE-enriched at the time that the cumulate source regions for the mare basalts formed. This REE signature may be attributed to either substantial orthopyroxene (LREE/HREE <<1) crystallization prior to crystallization of this source or by mixing of early and late cumulate rock types (Hughes et al. 1988, 1989). The fractionation of Nb from Ce and Ti from Nd in the primitive, low Ti-basalts indicate ilmenite accumulation in magma ocean cumulates prior to its stability as a liquidus phase (Longhi 1992; Shearer et al. 1996b). This implies that the sources for these basalts were mixtures of late, ilmenite-bearing cumulates with early, ilmenite-free cumulates (Longhi 1992; Shearer et al. 1996b).

Low-Ti picritic glasses are rare within the lunar collection. Delano (1986) distinguished only three low-Ti picritic glasses. High-pressure experimental studies have not be done on these picritic glasses. High-pressure experiments on low-Ti basalt compositions that have Mg' similar to the picritic glasses give multiple-saturation pressures of 20–25 kbars (Green et al. 1971b). It also appears that the low-Ti picritic glasses are not related to the known low-Ti crystalline mare basalts by fractional crystallization (Longhi 1987; Shearer et al. 1991; Shearer and Papike 1993), although it is expected that the parental magmas for the low-Ti basalts are similar. Trace-element studies by Hughes et al. (1988, 1989), Shearer et al. (1990, 1991), and Shearer and Papike (1993) are consistent with melting of mixed cumulate sources for the different types of picritic glasses.

The compositional range exhibited by the crystalline high-Ti mare basalts is the result of extensive fractional crystallization (40–50%) of olivine ± spinel ± pyroxene ± armalcolite ± ilmenite from different batches of Ti-rich magmas. For example, Beaty and Albee (1978), Beaty et al. (1979a, 1979b), and Rhodes and Blanchard (1980) interpreted the range in La/Sm and K at similar Co abundance for a series of Apollo 11 high-Ti basalts as indicating that the basalts represented different flows that were unrelated by surface or near-surface fractional crystallization. There are similar chemical fingerprints in high-Ti basalts from other sites that are interpreted as indicating the generation of different batches of magma from heterogeneous Ti-rich sources. These different signatures have been attributed to melting of heterogeneous sources (Paces et al. 1991), and to varying degrees of partial melting (Duncan et al. 1974, 1976; Shih et al. 1975; Walker et al. 1975). Hubbard and Minear (1975) and Ringwood (1975) invoked assimilative models for the generation of the high-Ti basalts. These models called upon the assimilation of ilmenite-bearing magma ocean cumulates by low-Ti basaltic magmas. Neal and Taylor (1992) further explored the possibility of assimilation (lunar granite, KREEP) and discounted the process.

Most of the models for the generation of primary high-Ti basalts require the melting of source rock types consisting of olivine + clinopyroxene + ilmenite ± plagioclase ± orthopyroxene (Duncan et al. 1974, 1976; Longhi et al. 1974; Shih et al. 1975; Walker et al. 1975; Nyquist et al. 1975, 1976; Drake and Consolmagno 1976; Delano 1986; Hughes et al. 1988, 1989; Snyder et al. 1990, Shearer et al. 1991, 1996b; Shearer and Papike 1993). This assemblage represents late-stage cumulates from the LMO and, in simple cumulate melting scenarios (Taylor and Jakes 1974; Walker et al. 1975; Taylor 1982; Snyder et al. 1992), these melts should have been generated at shallow depths. On the other hand, cumulate overturn and mixing could result in the generation of Ti-rich primary magmas deep in the cumulate pile. Although high-pressure experiments on the crystalline high-Ti basalts indicate melting and segregation depths of <100 km–220 km (O'Hara et al. 1970; Ringwood and Essene 1970; Longhi et al. 1974; Kesson and Lindsley 1974), the multiple-saturation pressure of the picritic glasses indicates that high-Ti basalts were generated at depths >400 km (Green et al. 1975; Walker et al. 1975; Delano 1980). Trace-element modeling of these picritic glasses also implies that a mixed cumulate source is, at this time, the most appropriate explanation (Delano 1986; Hughes et al. 1988, 1989; Snyder et al. 1990; Shearer et al. 1991, 1996b; Shearer and Papike 1993).

In the above discussion, we suggest that many of the crystalline mare basalts were a product of shallow fractional crystallization of more primitive basaltic magmas. What then are the parental magmas for the mare basalts? Two possible alternatives are that they are similar to the picritic glasses and were derived from initial melting in the deep lunar mantle (>400 km) or that mare basalts represent the crystallization of more Fe-rich, olivine-poor primary magmas that were produced at shallower depths. We propose that the former conclusion is correct for several reasons. First, the crystalline mare basalts and picritic mare glasses define the same ranges in TiO_2 and overlap in age. This is interpreted as indicating that they were produced by melting of similar mantle assemblages under comparable conditions. Second, crystalline mare basalts have all the chemical attributes (lower Ni, Mg', higher Al) associated with substantial olivine fractionation. Although Longhi (1992) and Shearer and Papike (1993) demonstrated that the picritic mare glasses and crystalline mare basalts, in most cases, were not related by shallow fractional crystallization, calculated liquid lines of descent are parallel. Therefore, the preservation of primitive magmas as picritic mare glasses is a function of magma transport and eruptive mechanism rather than dramatically difference source regions. Third, the generation of mare basalts from a shallow, subcrustal source between 3.9 and 3.0 Ga is not reconcilable with the thermal nature of the upper lunar mantle. The liquidus temperatures of mare basalts exceed 1200 °C at depths corresponding to about 0.5 GPa (Longhi 1992). No mechanism exists to reheat the shallow lunar mantle an additional 200 to 400 °C 600 Ma after the formation of the Moon and to maintain this temperature for 700 Ma. Furthermore, the existence and preservation of mascons in multiringed basins that formed at about 3.9 Ga indicate a relatively cool (1000 °C) and rigid upper mantle.

SYNOPSIS

Over the past three decades, lunar magmatism has been studied by numerous researchers. We have attempted to summarize the history of lunar magmatism and models proposed for the generation of lunar magmas. This review illustrates the continuous nature of lunar magmatism and the profound influence of early differentiation and catastrophic bombardment on later lunar mantle dynamics, and magmatism. We grouped magmatism into multiple stages of activity based on sampled rock types and summarized some of the more complete models for each stage (Fig. 15).

Stage 1: Lunar differentiation and associated magmatism (Fig. 15a).

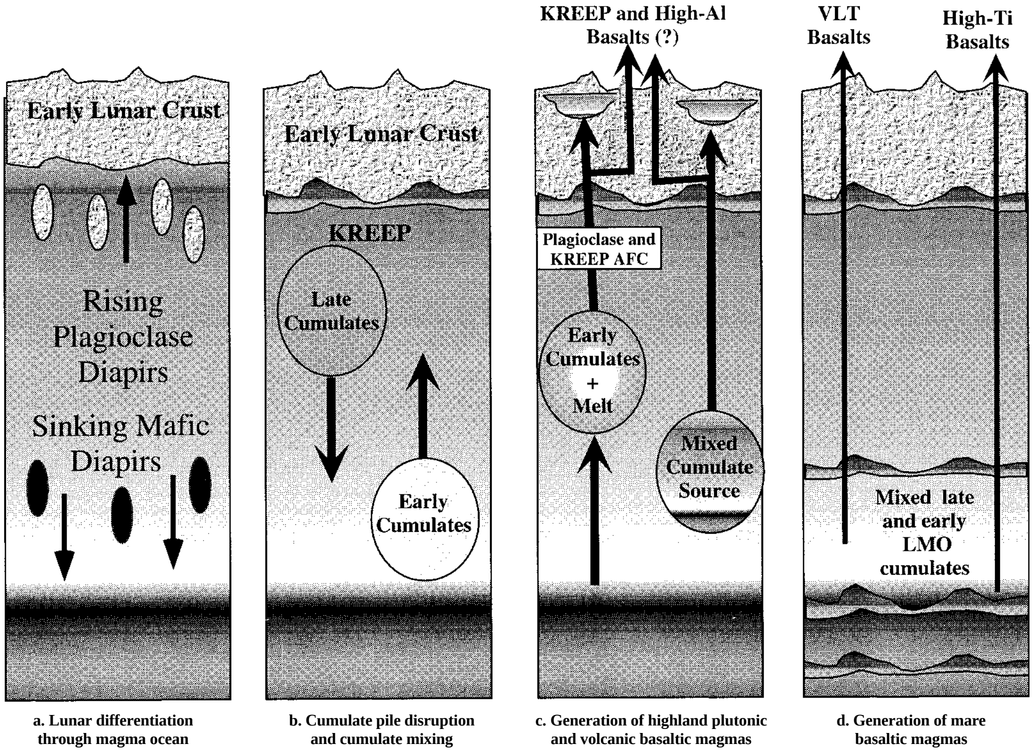
The generation of a LMO that was responsible for producing an anorthositic crust and a differentiated mantle is an elegant explanation of the available lunar data. The extent of lunar melting and mantle processing strongly depends on the melting mechanisms. For example, high accretion rates may favor a more-extensive magma ocean, whereas lower accretion rates may result in a magma ocean complex as proposed by Shirley (1983). Estimates for the time over which the magma ocean crystallized range from tens (Shearer and Newsom 1998) to hundreds of millions of years (Solomon and Longhi 1977).

Stage 2: Disruption of lunar magma ocean cumulates (Fig. 15b)

On the basis of the mare basalt data (chemical and experimental) and dynamic modeling of an idealized LMO cumulate pile, it appears that soon after the crystallization of most of the LMO, the cumulate pile experienced gravitationally driven disruption. This resulted in the transport of late cumulates into the deep lunar mantle and mixing of magma ocean cumulates on a variety of scales. The transport of late, incompatible-element-enriched (U, Th, Rb, K) cumulates may have initiated melting in the more-depleted cumulates (olivine + orthopyroxene dominant) and contributed evolved signatures to these deeper sources (e.g., negative Eu anomaly). Upwelling of the less-dense, more-magnesian, early cumulates may have initiated decompressional melting.

Stage 3: Post-magma ocean highland magmatism (Fig. 15c)

It has been demonstrated clearly that the highlands Mg suite and alkali suite are chemically distinct from the ferroan anorthositic crust. Yet, the duration of magmatism of these two highland plutonic suites may have overlapped the formation of the ferroan anorthositic crust and continued substantially after the termination of magma ocean crystallization. Whereas the ferroan anorthositic crust was probably produced during the



a. Lunar differentiation through magma ocean
b. Cumulate pile disruption and cumulate mixing
c. Generation of highland plutonic and volcanic basaltic magmas
d. Generation of mare basaltic magmas

FIGURE 15. Summary models for the evolution of the lunar mantle and magmatism. (**a**) Early lunar crust formation and lunar mantle differentiation through magma ocean development and crystallization. (**b**) Disruption of cumulate products of magma ocean crystallization due to gravitational destabilization. (**c**) Generation of parental highland basaltic magmas through pressure release melting. The high Mg' and incompatible-element signature of the Mg suite rock types are a result of either assimilation of KREEP and early crust by the parental magmas followed by fractional crystallization or partial melting of hybrid cumulate sources and assimilation of KREEP. The alkali suite and highland volcanics may be related to Mg suite magmas through fractional crystallization and crystal accumulation or are simply contemporaneous magmas produced through partial melting of distinctively different sources. (**d**) Generation of mare basalts through small to moderate degrees of partial melting of mineralogically chemically varied hybrid cumulates. Melting to produce these magmas was initiated in the deep lunar mantle.

crystallization of the magma ocean, these slightly younger highland rock types were a product of melting of the cumulate products of the magma ocean. The melting that produced the magnesian parental magmas for these suites may have been initiated by decompressional melting of early magma ocean cumulates during cumulate pile destabilization. An aluminous component and evolved KREEP signature could have been incorporated into these primitive basaltic magmas through melting of a hybridized mantle source and through assimilation near the base of the lunar crust, respectively. The petrogenetic relationship between the Mg suite and alkali suite is still debated. The alkali suite could represent the differentiation products of Mg suite parental magmas. On the other hand, the two suites may represent separate, but contemporaneous episodes of basaltic magmatism.

Stage 4: Pre-basin volcanism (Fig. 15c)

Sample and remote sensing data indicate that early lunar volcanism was contemporaneous with periods of highlands plutonism and catastrophic bombardment of the lunar surface. Because of our limited sampling, these basaltic magmas are represented by KREEP basalts and high-alumina basalts. Their relationship to the contemporaneous plutonism is not clear. Numerous researchers have suggested that the KREEP basalts may be parental to both the Mg suite and alkali suite.

Stage 5: Mare basalts, late remelting of Magma ocean cumulates (Fig. 15d)

During and following the late stages of catastrophic bombardment, and perhaps an extension of thermal events that initiated pre-basin volcanism, basin-associated eruptions of mare basalts occurred. These basalts exhibit a wide range of compositions that resulted from near-surface fractionation of an extremely wide range of primary basaltic magmas. The crystalline mare basalts provide abundant information concerning near surface processes, but it is the volcanic, picritic glasses that provide us with more reliable information concerning mantle sources. Most likely, the high-Ti to very low-Ti basalts were produced by small to moderate degrees of partial melting of hybrid cumulate sources initiated in the deep lunar mantle (greater than 400 km). Chemical and isotopic signatures reflect basaltic magma generation from a differentiated and hybridized source. The wide range of compositions exhibited by the mare basalts compared to earlier episodes of basaltic magmatism may reflect the thermal regime in the lunar mantle that limited the extent of partial melting and melt-source homogenization.

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