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Petrogenesis of mare basalts: A record of lunar volcanism*

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Abstract—Returned rock and soil samples from our nearest planetary neighbor have provided the basis for much of our understanding of the origin and evolution of the Moon. Of particular importance are the mare basalts, which have revealed considerable information about lunar volcanism and the nature of the mantle, as well as post-magma-generation processes. This paper is a critical review of the petrogenetic models for the generation of mare basalts formulated over the last twenty years.

We have used all available mare basalt analyses to define a six-fold classification scheme using TiO_2 contents as the primary division (i.e., <1 wt% = very low-Ti or VLT; 1–6 wt% = low-Ti; >6 wt% = high-Ti). A secondary division is made using Al_2O_3 contents (i.e., <11 wt% = low-Al; >11 wt% = high-Al), and a tertiary division is defined using K contents (i.e., <2000 ppm = low-K; >2000 ppm = high-K). Such divisions and subdivisions yield a classification containing twelve categories, of which six are accounted for by the existing Apollo and Luna collections. Therefore, we present our discussions in the form of six mare basalt rock types: (1) high-Ti/low-Al/low-K (referred to as "high-Ti/low-K"); (2) high-Ti/low-Al/high-K (referred to as "high-Ti/high-K"); (3) low-Ti/low-Al/low-K (referred to as "low-Ti"); (4) low-Ti/high-Al/low-K (referred to as "high-alumina"); (5) low-Ti/high-Al/high-K (referred to as "VHK"); and (6) VLT/low-Al/low-K basalts (referred to as "VLT").

A variety of post-magma-generation processes have been invoked, such as fractional crystallization, either alone or combined with wallrock assimilation, to explain the compositional ranges of the various mare basalt suites. In order to evaluate these proposed petrogenetic processes, this review is by rock type and is non-site specific, but for each rock type, reference to particular lunar sample return missions is brought forth. This permits a comparison of similarities and differences of broadly similar rock types correlated with geography on the Moon, which, in turn, allows a more thorough petrogenetic evaluation.

High-Ti mare basalts (i.e., high-Ti/low-Al/low-K) are found at Apollo 11 and Apollo 17 sites; however, the A-11 basalts contain lower TiO_2 abundances, a considerably larger range in trace-element contents, and the only occurrence of high-Ti/high-K mare basalts. Fractional crystallization and source heterogeneity within each site are the keys to understanding the petrogenesis of the high-Ti basalts. Low-Ti basalts (including both low-Al/K and high-Al/K varieties) are found at Apollo 12, 14, and 15, and Luna 16 sites. The low-Ti basalts exhibit a wide range of major- and trace-element compositions and require source heterogeneity, fractional crystallization, and some assimilation. The high-alumina mare basalts (i.e., low-Ti/high-Al/low-K) are found at Apollo 14 and Luna 16 sites and exhibit a wide range of major- and trace-element compositions. However, in these examples, source heterogeneity is not a major factor. Indeed, fractional crystallization coupled with KREEP assimilation, particularly for the Apollo 14 variants, can explain the compositional ranges of these high-alumina basalts. The VHK mare basalts (i.e., low-Ti/high-Al/high-K) have been sampled only at the Apollo 14 locale and are products of a parental high-alumina magma assimilating lunar granite. Very low-Ti (VLT) mare basalts (i.e., VLT/low-Al/low-K) are found at Apollo 17 and Luna 24 sites. Fractional crystallization has had a major influence upon the range in VLT compositions, but Luna 24 VLT basalts have been derived from a source slightly different in composition from that for Apollo 17 VLT varieties. For example, the Luna 24 VLT basalts generally exhibit positive Eu anomalies, a unique property for mare basalts, which almost always have negative Eu anomalies.

The concept of a lunar magma ocean (LMO) is generally accepted, and source modelling of all basalts invokes a "mafic LMO cumulate source." This is the only unifying model for mare basalt petrogenesis, but the semantics and logistics of it are and will be debated for many years. For example, major convective overturn of the LMO appears plausible, but whether this occurred on a local- or planet-wide scale to produce source heterogeneity remains to be determined.

INTRODUCTION

MISSIONS TO THE MOON FROM 1969 to 1976, both manned and unmanned, returned valuable samples which provided the first opportunity to study in detail the origin and evolution of a celestial body other than Earth. These missions were generally scattered about the nearside lunar equator, with the US Apollo 12 and 14 landings in the "west," Apollo 11 and 16 in the "east," and the unmanned Soviet Luna missions (16, 20, and 24) on the eastern extremity of the Moon (Fig. 1). The Apollo 15 and 17 missions landed north of the lunar equator, in the Appenine Highlands and Mare Serenitatis, respectively (Fig. 1).

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FIG. 1. Map of lunar sample return sites—American and Soviet.

Critical in studying mare basalts is an appreciation of early lunar evolution. The crust of the Moon is feldspathic, and it has been postulated (e.g., Wood et al., 1970; Smith et al., 1970; Warren and Wasson, 1979; Taylor, 1982; Warren, 1985) that this crust formed by the flotation of plagioclase-rich cumulates in a lunar magma ocean (LMO). This concept requires that the Moon was initially molten (due primarily to energy from initial accretion) to a depth of 200–400 km or more (in either one huge or several smaller "magmaspheres"; Shirley and Wasson, 1981) and underwent fractional crystallization with the dense mafic cumulates sinking and forming the lunar upper mantle. Early mafic cumulates from the LMO would be olivine and orthopyroxene followed by late-stage clinopyroxene and ilmenite (e.g., Taylor and Jakes, 1974; Walker et al., 1975; Hughes et al., 1989). The separation of plagioclase from these mafic cumulates requires that the latter had resultant negative Eu anomalies. Because almost every mare basalt analyzed to date has a negative Eu anomaly, it seems likely that the basalts are derived from such mafic cumulates. Therefore, analysis of the basaltic samples returned from the Moon is particularly significant as they should yield detailed mineralogic and chemical information about their source regions in the lunar interior (i.e., the lunar mantle). As such, the significance of these mare basalts far outweighs their relative crustal volume ($\sim$1%) and surficial extent ($\sim$17%).

Mare basalts are higher in total iron and Fe/Mg and generally lower in Al_2O_3 (<15 wt%) and viscosity relative to terrestrial basalts, but they exhibit a wide range in TiO_2 contents (<1 wt% to $\sim$15 wt%). They also exhibit a wide range of igneous textures, from vitrophyric through sub-ophitic/ophitic to poikilitic. Occasionally, certain textures are correlated with a particular chemical suite. For example, most Apollo 11 high-Ti/high-K basalts have an intersertal texture, whereas the Apollo 11 high-Ti/low-K mare basalts exhibit ophitic to sub-ophitic textures (e.g., Beaty and Albee, 1978; Beaty et al., 1979a,b). However, as texture is dependent upon cooling rates (i.e., nucleation and growth), it is common to find a variety of igneous textures within any given suite. One of the most significant uses of textural analysis is in recognizing parental magmas, which usually exhibit vitrophyric (glassy) textures (e.g., Rhodes et al., 1976; Beaty et al., 1979a,b; Warner et al., 1979). Detailed summaries of mare basalt textures can be found in Papike et al. (1976), Papike and Vaniman (1978), and Taylor et al. (1991). Papike et al. (1976) and Papike and Vaniman (1978) demonstrated the correlation between major-element chemistry and modal mineralogy (Table 1). As expected, the high-Ti basalts contain the largest modal abundances of opaques (mostly ilmenite and spinels) and the VLT basalts the least, whereas high-alumina basalts are dominated by feldspar (>40 modal %).

Essential to understanding mare basalt petrogenesis is the unravelling of post-magma-generation processes in order to define a primary magma (i.e., unmodified mantle melt). Therefore, the suites of basalts from each landing site require

Table 1: Summary of average modal data for the mare basalts (modified after PAPIKE and VANIMAN, 1978).

	OPAQUES	PYROXENE	FELDSPAR	OLIVINE
A-17 high-Ti	24.4	47.7	23.4	4.6
A-11 high-K/high-Ti	20.6	57.5	21.7	0.1
A-17 low-K/high-Ti	15.1	51.6	33.3	---
A-11 low-K/high-Ti	14.6	50.9	32.2	2.3
A-12 ilmenite	9.3	61.1	25.9	3.6
A-12 pigeonite	9.1	68.4	21.1	1.4
A-12 olivine	7.1	53.5	19.2	20.2
A-15 olivine-normative	5.5	63.3	24.1	7.0
A-15 quartz-normative	3.7	62.5	33.8	---
Luna 16 high-alumina	7.1	51.5	41.2	0.1
A-14 high-alumina	9.0	37.1	43.9	10.0
A-14 VHK/high-alumina	8.0	33.2	46.8	12.0
Luna 24 "VLT"	1.8	48.6	39.1	10.4
A-17 "VLT"	1.0	61.7	31.9	5.4

Apollo 14 data from Neal et al. (1988a, 1989).

careful examination in order to define their evolution and highlight the processes which have produced the final compositions found. It seemed initially that the most likely candidates for parental melts were the glasses produced from fire-fountaining of volcanic vents. Some twenty-five distinct groups of volcanic glasses have been defined from existing lunar sample collections (DELANO, 1986). DELANO and LIVI (1981) identified two arrays of volcanic glasses, using the most primitive members from each of the twenty-five groups. These authors concluded that each array seems to have been produced from a chemically isolated, cumulate system present at a specific depth within the Moon. SHEARER et al. (1991) concluded that each of these two systems was composed of both primitive and evolved cumulate components, whereas GALBREATH et al. (1990) concluded that three cumulate components (a differentiated cumulate, trapped residual liquid, and extremely Fe-rich components) were required. Perhaps the most intriguing, but perplexing, problem is establishing a genetic link between mare basalts and these volcanic glasses (LONGHI, 1987, 1990; SHEARER et al., 1991; DELANO, 1990). Although many attempts have been made to link the crystalline basalts with the volcanic glasses (e.g., VANIMAN and PAPIKE, 1977; TAYLOR et al., 1978), no relationship has been unequivocally established to date. Therefore, the glasses will not be considered further in this review.

KREEP basalts form a small, yet significant, portion of the returned pristine basalt suite and comprise a distinct group on the basis of their high incompatible-element abundances. KREEP basalts are found at practically every site from which lunar samples have been returned, either as basaltic impact melts (e.g., SHIH, 1977; MCKAY et al., 1978, 1979; TAYLOR, 1982; RYDER and SPUDIS, 1987) or as pristine basalts of volcanic origin (e.g., MEYER, 1972; BASU and BOWER, 1976; DOWTY et al., 1976; RYDER, 1976, 1987; SALPAS et al., 1987). The intricacies of KREEP basalt formation are beyond the scope of this review, but they represent basalts contaminated with "urKREEP" (urKREEP represents the final incompatible-element-rich dregs after the crystallization of the LMO; WARREN and WASSON, 1979). As such, KREEP basalts do not represent pure urKREEP (NEAL and TAYLOR, 1989) and, conversely, KREEP basalts cannot represent primary melts from the lunar mantle—both components have contaminated each other. Therefore, neither the source nor melt compositions for either urKREEP or the parent magmas of KREEP basalts magma can be determined and because of this, KREEP basalts will not be discussed further.

Many unresolved problems remain in the studies of the petrology, geochemistry, and petrogenesis of mare basalts. Some of the difficulties in interpreting lunar petrology and geochemistry have resulted from the "splitting" rather than "lumping" of samples and data, as well as over-interpretations based on limited samplings. Indeed, the Soviet Luna samples are soils and as such do not contain sizeable rock fragments; almost all basaltic samples weigh less than 1 g, most only 100 mg. In this paper, we will review and evaluate mare basalts, their classification, sources, and overall petrogenesis. For brevity, we have not reviewed all models and hypotheses, but have attempted to present what we consider to be the most plausible and realistic ones.

CLASSIFICATION OF MARE BASALTS

Initial studies of the lunar samples were conducted immediately following the Moon mission. Therefore, many of the investigations of the mare basalts were and continue to be site specific. As such, a myriad of different classifications have been entrenched in the literature, each relevant to only one particular landing site, producing at least twenty-one different basaltic types (Table 2). However, although the nomenclature is different, the compositions of several of these basaltic types are surprisingly similar (e.g., RYDER, 1991a).

Another problem which complicates the classification and study of lunar samples is that of representative sampling. In our quest for an increased data base, samples on the order of a few milligrams have been analyzed in order to obtain "whole-rock" chemistry (e.g., the Soviet Luna samples, which are basically soils). HASKIN et al. (1977) and LINDSTROM and HASKIN (1978, 1981) demonstrated the problems involved in obtaining a "representative sample," and LINDSTROM and HASKIN (1978) suggested that variations found in most groups of mare basalts can be attributed to crystallization of a single, undifferentiated flow. Indeed, RHODES et al. (1976) noted that several Apollo 17 coarse-grained samples did not fall upon already defined trends for Apollo

Table 2: Types of Mare Basalts Present at Each Returned-Sample Site.

Site	No. of Types	Names
Apollo 11	5	A,B1,B2,B3,D
Apollo 12	4	Olivine, Pigeonite Feldspathic, Ilmenite
Apollo 14	2	High-Alumina, VHK
Apollo 15	2	Quartz-Normative Olivine-Normative
Apollo 17	6	A,B1,B2,C,D (U?) VLT
Luna 16	1	High-alumina/Feldspathic
Luna 24	1	VLT/Ferrobasalt

17 high-Ti basalts. These authors attributed the scatter to unrepresentative sampling or "short-range unmixing" and set such samples apart as Class U. As another example, HASKIN et al. (1977), LINDSTROM and HASKIN (1978), and SCHURAYTZ and RYDER (1988, 1991) demonstrated that much of the scatter of data for low-Ti Apollo 15 olivine-normative basalts was due to unrepresentative sampling, although a small degree of olivine fractional crystallization was still required to explain the data spread. Although these authors attempted to resolve the question of unrepresentative sampling, at least in the case of the Apollo 15 olivine-normative basalts, it is generally minimized when considering fine-grained and vitrophyric samples. Representative analyses of coarse-grained basalts require a much larger sample size, which is determined by grain size (SCHURAYTZ and RYDER, 1988, 1991). Therefore, unrepresentative sampling has, on occasion, also clouded the classification of mare basalts (e.g., Apollo 17 Type U).

PAPIKE et al. (1976) initially subdivided the mare basalts into two groups on the basis of TiO_2 contents, and subsequently PAPIKE and VANIMAN (1978) defined three groups of mare basalts: (1) very low-Ti (VLT) basalts (<1 wt% TiO_2); (2) low-Ti basalts (1–5 wt% TiO_2); and (3) high-Ti basalts (9–14 wt% TiO_2). They also noted some intermediate-Ti basalts (i.e., 5–9 wt% TiO_2) primarily from the Apollo 14 site, particularly from breccia 14063 (RIDLEY, 1975). We have plotted all available (>500) mare basalt analyses and placed them within this original classification (Fig. 2). The high-Ti basalts are found at the Apollo 11 and 17 landing sites, low-Ti basalts at Apollo 12, 14, and 15 and Luna 16 landing sites, and VLT basalts at the Apollo 17 and Luna 24 locales. Generally, mare basalts appear to exhibit fractional crystallization trends in that the high-Ti basalts exhibit a decrease of Ti contents with decreasing MG# [100∗molar (Mg/(Mg + Fe))], due to early crystallizing ilmenite/armalcolite and olivine; low-Ti basalts exhibit a slight increase in Ti with decreasing MG#; VLT basalts do not exhibit a striking trend, but generally those samples with the lowest MG# contain the highest Ti contents. These trends often overlap the boundaries set by PAPIKE et al. (1976), largely because of the availability of new data.

While this classification by Ti contents is not perfect, it does allow a convenient framework for describing the mare basalts. Furthermore, such a classification allows the grouping of basalts of similar composition, which would otherwise be split apart if each site was individually described. We propose that the primary criterion for the classification of mare basalts be based on TiO_2 contents, but modifications of this classification are required. The first modification is that the lower boundary of the high-Ti basalts be lowered to 6 wt% and the upper limit of the low-Ti basalts be raised to 6 wt% TiO_2, thus eliminating the questionable intermediate-Ti region.

While TiO_2 content allows a general classification of mare basalts, consideration of other major elements demonstrates that TiO_2 is not all inclusive. For example, Apollo 14 and Luna 16 mare basalts are generally low-Ti, but they also are rich in alumina and have been described as "high-alumina" mare basalts (e.g., WARNER et al., 1980; VINOGRADOV, 1971; Fig. 3a,b). We concur with this designation for Apollo 14 and Luna 16 mare basalts as high-alumina, as they almost always contain > 11 wt% Al_2O_3 (Fig. 3a,b). Therefore, we will use Al_2O_3 contents as a secondary criterion for the classification of mare basalts. Although some Luna 24 VLT basalts also contain >11 wt% Al_2O_3, they fall at the end of a general fractional crystallization trend (Fig. 3a) and will be described as strictly VLT. We will not include those few Apollo 12 basalts containing >11 wt% Al_2O_3 (Fig. 3a,b) in the high-alumina division as it has been suggested that these are either related by fractional crystallization to others from that site containing <11 wt% Al_2O_3 (NYQUIST et al., 1979; NEAL, 1991) or are unrepresentative samples of basalts containing <11 wt% Al_2O_3 (NEAL et al., 1992). In general, the majority of other mare basalts contain <11 wt% Al_2O_3.

A tertiary subdivision of the low-Ti basalts (Fig. 4) from Apollo 14 will be implemented on the basis of K contents. This will include the Apollo 14 very-high-potassium (VHK) basalts (e.g., SHERVAIS et al., 1985b), as well as the Apollo 11 high-K group of high-Ti basalts (i.e., all samples that contain >2000 ppm K). Note that on the basis of this classification, the tridymite ferrobasalt of SHERVAIS et al. (1985a) falls within the VHK field and, therefore, is described as such.

In summary, we will use TiO_2 contents as the primary criterion for the classification of mare basalts (divisions at 1 wt% and 6 wt% dividing the basalts into VLT, low-Ti, and high-Ti), Al_2O_3 contents as the secondary criterion (division at 11 wt% dividing mare basalts into low- and high-alumina), and K contents as the tertiary criterion (division at 2000 ppm K dividing the basalts into low- and high-K groups). As can be seen from Fig. 5, this classification has the potential to highlight twelve mare basalt groups, of which only six are represented in the current sample base. Whether the remaining six will be sampled, or whether this classification will be superseded by future missions, remains to be seen.

Based upon the above chemical divisions, we have organized our further discussions below into six mare basalt rock-types: (1) high-Ti/low-Al/low-K (referred to as "high-Ti/low-K"); (2) high-Ti/low-Al/high-K (referred to as "high-Ti/high-K"); (3) low-Ti/low-Al/low-K (referred to as "low-Ti"); (4) low-Ti/high-Al/low-K (referred to as "high-alumina"); (5) low-Ti/high-Al/high-K (referred to as "VHK"); and (6) VLT/low-Al/low-K basalts (referred to as "VLT").

An observation which is consistent with the classification of mare basalts defined above concerns the radiometric ages of the VLT, low-, and high-Ti (including high-Ti/high-K

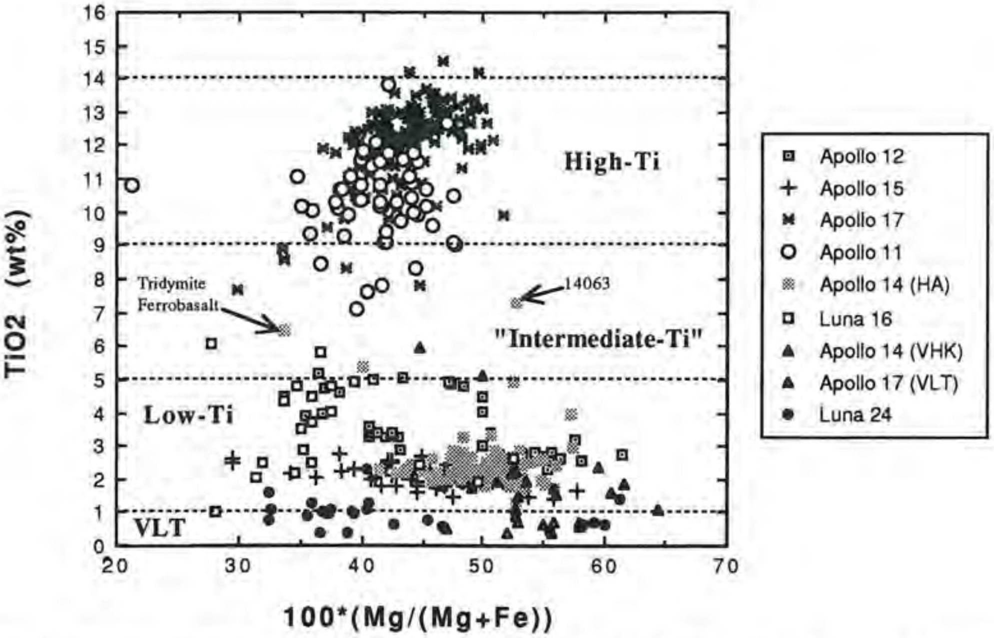


FIG. 2. TiO_2 (wt%) vs. MG# (molar) for five hundred lunar basalts, Apollo 17 orange glass (HUGHES et al., 1989), and Apollo 15 green glass (DELANO and LIVI, 1981). Boundaries are those defined by PAPIKE and VANIMAN (1978) for mare basalt classification on the basis of TiO_2 abundance. Data were taken from the following sources: Apollo 11—BEATY et al. (1979a); MA et al. (1980); RHODES and BLANCHARD (1980); Apollo 12—BEATY et al. (1979c); BRUNFELT et al. (1971); COMPSTON et al. (1971); CUTTITTA et al. (1971); ENGEL et al. (1971); HASKIN et al. (1971); HUBBARD and GAST (1971); KHARKAR and TUREKIAN (1971); KUSHIRO and HARAMURA (1971); MAXWELL and WIIK (1971); MORRISON et al. (1971); NYQUIST et al. (1979); RHODES et al. (1977); SCHNETZLER and PHILLPOTTS (1971); SCOON (1971); TAYLOR et al. (1971); VOBECKY et al. (1971); WAKITA and SCHMITT (1971); WAKITA et al. (1971); WÄNKE et al. (1971); Apollo 14 (high-alumina)—DICKINSON et al. (1985); NEAL et al. (1988a, 1989); RIDLEY (1975); SHERVAIS et al. (1985a); WARNER et al. (1980); Apollo 14 (VHK)—GOODRICH et al. (1986); NEAL et al. (1988b, 1989); SHERVAIS et al. (1985b); WARREN et al. (1986); Apollo 15—CHAPPELL and GREEN (1973); DOWTY et al. (1973); MA et al. (1978a); RHODES and HUBBARD (1973); RYDER and STEELE (1988); VETTER et al. (1988); Apollo 17 (high-Ti)—BRUNFELT et al. (1974); MURALI et al. (1977); NEAL et al. (1990a,b); RHODES et al. (1976); RYDER (1990); SHIH et al. (1975); WARNER et al. (1975a,b, 1979); Apollo 17 (VLT)—TAYLOR et al. (1978); VANIMAN and PAPIKE (1977); WENTWORTH et al. (1979); Luna 16—ALBEE et al. (1972); GRIEVE et al. (1972); HELMKE and HASKIN (1972); KEIL et al. (1972); KURAT et al. (1976); MA et al. (1979); REID and JAKES (1974); VINOGRADOV (1971); Luna 24—HASKIN (1978); LAUL et al. (1978); MA et al. (1978b); RYDER and MARVIN (1978); TAYLOR et al. (1978).

varieties), high-alumina, and VHK basalts (Fig. 6, modified after NYQUIST and SHIH, 1992). With a larger data base, we are able to see correlations between basalt chemistry and age, contrary to the earlier observation of PAPIKE and VANIMAN (1978). The high-Ti basalts from the Apollo 11 and 17 landing sites are generally older (3.6–3.9 Ga) than the low-Ti basalts from the Apollo 12 and 15 sites and the VLT basalts from Mare Crisium (3.2–3.4 Ga; e.g., TAYLOR et al., 1983; DASCH et al., 1987). The Apollo 17 VLT basalts have not been dated because of their small sample size. The high-alumina basalts from Apollo 14, and to a lesser extent the VHK's, yield the oldest ages (3.8–4.3 Ga). However, the high-alumina basalts from Mare Fecunditatis (Luna 16) yield similar crystallization ages to the low-Ti and VLT basalts (3.2–3.4 Ga). This suggests that low-Ti, high-alumina, basaltic volcanism spanned approximately 1.3 Ga and may be the most persistent of the returned basaltic types. Furthermore, crater degradation studies (e.g., BOYCE, 1976) suggest that basalts younger than 3.0 Ga also are present on the lunar surface and many of these appear to be of the high-Ti varieties.

MODELS FOR MARE BASALT GENERATION

Upon return of the first lunar samples, it was recognized that lunar basaltic melts have lower viscosities (higher Fe/Mg ratios) than their terrestrial counterparts, accounting for the lack of relief to the maria ("seas"). The general circular form of the maria was caused by flooding of large ringed basins by such low-viscosity basaltic magmas. A clear distinction exists between the timing of formation of multiringed basins, excavated by meteorite impact, and the maria formed by lava flooding (e.g., TAYLOR, 1982). For example, the Imbrium basin was formed by impact, and Mare Imbrium is the sea of lava which later filled this basin. The height to which lavas can rise is governed by the density difference between melt and the overlying column of rock (TAYLOR, 1982). Therefore, there is a discrepancy between the amount of extruded basalt on the nearside with that on the farside (BVSP, 1981), with mare basalts being more common on the nearside of the Moon, probably in response to a thinner crust (60–70 km on the nearside vs. 80–90 km on the farside).

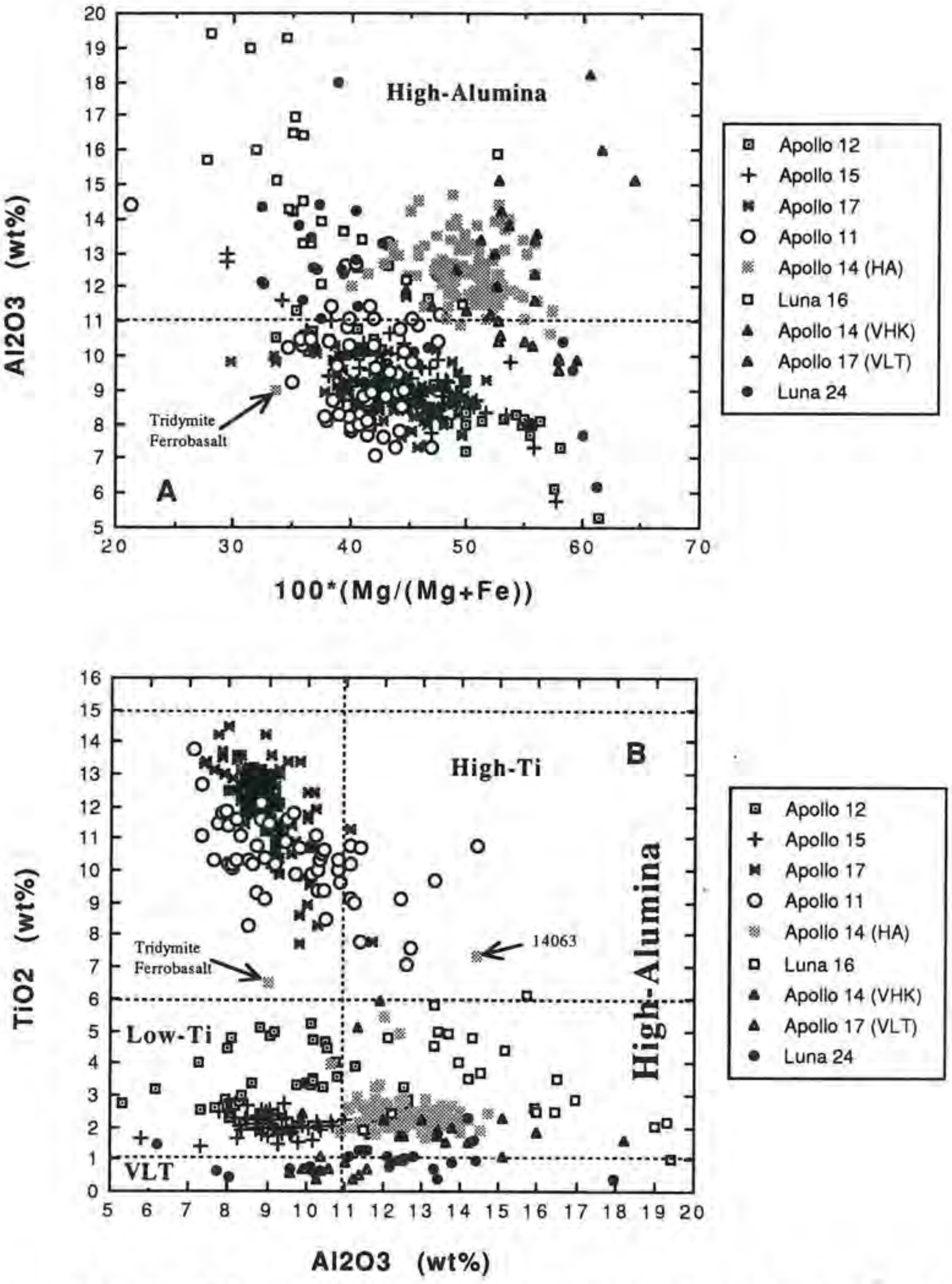


FIG. 3. Al_2O_3 (wt%) vs. MG# (molar) and TiO_2 (wt%) demonstrating the subdivision of mare basalts into high-alumina varieties. Data sources as in Fig. 2.

An early question in mare basalt genesis was whether basalt generation was a direct response to giant impactors (e.g., BIRCK and ALLÉGRE, 1974) or a result of unrelated heating of the interior. The ages of impacts that formed the mare are >3.9 Ga, whereas radiometric ages of mare basalts demonstrate an almost continuous distribution from ~4.3 Ga to 3.0 Ga, suggesting that melt generation in response to the occasional giant impactor was not significant (TAYLOR, 1982).

General models have been developed for the generation of the mare basalts from a single, primitive source, by variable degrees of partial melting. Others suggest the lunar interior is layered, and the differences in composition between various mare basalts is due to origins at different depths within the Moon. These general models may be divided into four groups: (1) primitive source, (2) assimilation, (3) cumulate "layer-cake," and (4) cumulate "convective-overturn" models.

Primitive Source (1)

The primitive source model proposes that mare basalt source regions were composed of primitive, unfractionated

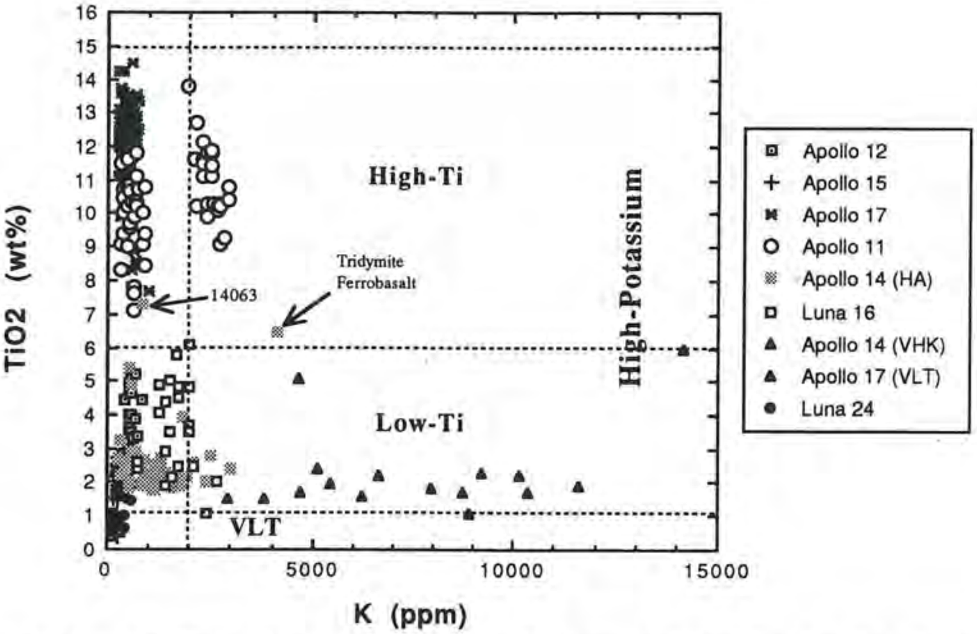


FIG. 4. K (ppm) vs. TiO_2 (wt%) demonstrating the subdivision of mare basalts into high-K varieties (data sources as in Fig. 2).

lunar material of bulk Moon composition at the time of magma genesis (RINGWOOD and ESSENE, 1970; GREEN et al., 1971). The range in chemistry is attributed to variable degrees of partial melting of this homogeneous source, but such a model has the following problems: (1) the wide variety and composition of high- and low-Ti basalts cannot be accounted for; (2) the ever-present negative Eu anomaly in practically all mare basalts requires prior plagioclase separation from the source; (3) primitive source models cannot account for the early differentiation event recorded by radiogenic isotopes; and (4) high-Ti basalts contain practically no nickel, but low-Ti basalts are relatively Ni-rich (TAYLOR, 1982), thereby indicating that the high-Ti basalts carry no memory of, or contact with, a primitive lunar nickel content.

Assimilation / Hybridization (2)

The assimilation or hybridization source models include the idea that magmas from the primitive lunar interior

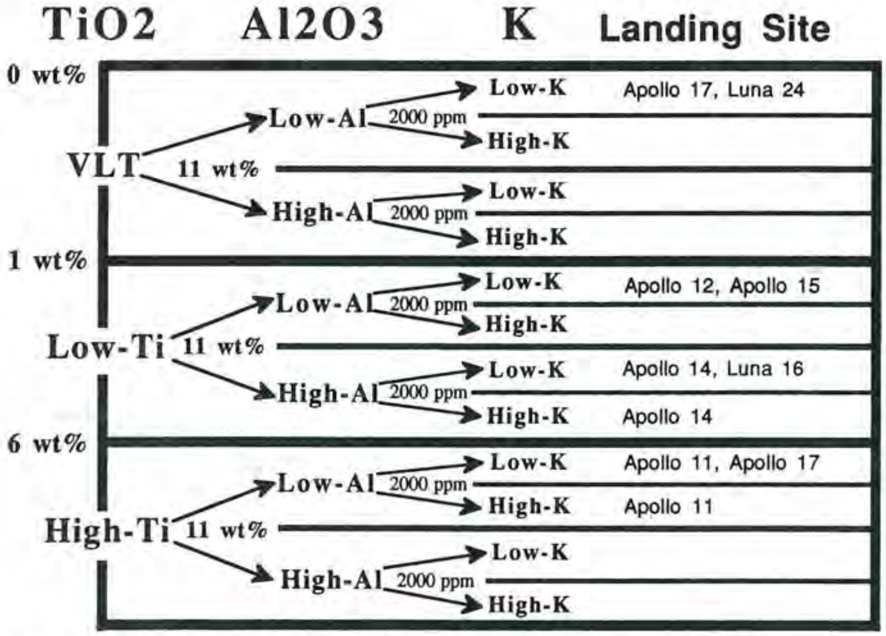


FIG. 5. The proposed classification for mare basalts based primarily upon TiO_2 contents, with a secondary subdivision on the basis of Al_2O_3, and a tertiary subdivision using K contents.

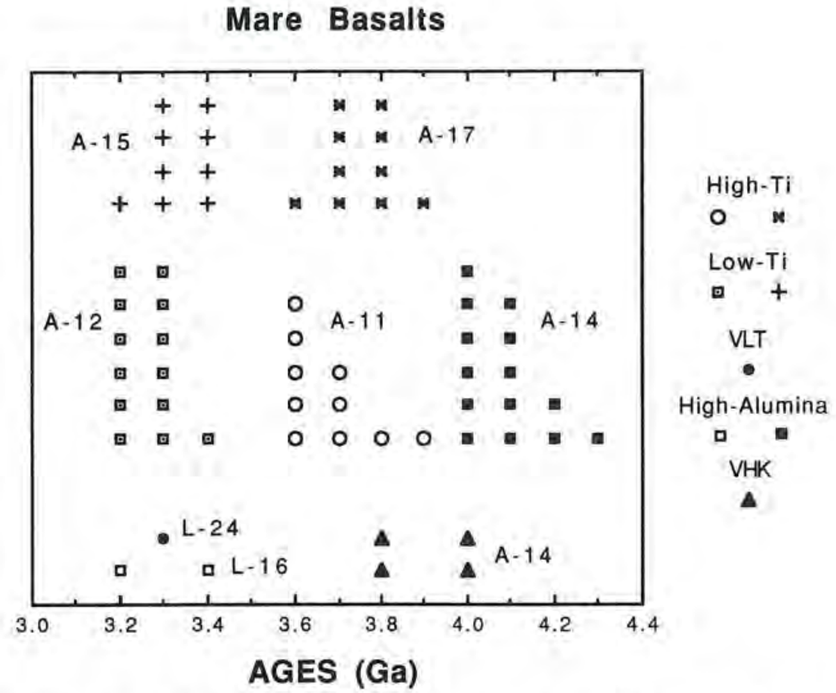


FIG. 6. Histogram of mare basalt radiometric ages (after NYQUIST and SHIH, 1990) demonstrating a general correlation between composition with age. See text for discussion.

selectively assimilated the late-stage lunar cumulates from the LMO, producing hybrid mare basalts (HUBBARD and MINEAR, 1976; RINGWOOD and KESSON, 1976). The high-Ti basalts were interpreted by KESSON (1975) as hybrid liquids produced by the assimilation of a sub-crustal Ti-rich residuum by a parent magma of low-Ti composition. RINGWOOD and KESSON (1976) suggested a "dynamic assimilation" model for mare basalt petrogenesis, assuming LMO initiation shortly after Moon formation. In this model, pods (5–20 km wide) of Fe-, Ti-, and incompatible-element-rich, late-stage, LMO residual melt resided beneath the lunar crust, but above the early cumulates. The largest pods continued to differentiate and formed ilmenite-rich pyroxenites of sufficient density to sink into the underlying early cumulates. The temperature of the interior rose because of decay of radioactive elements, and as the ilmenite-rich pyroxenites consisted of relatively low-melting-point components, continued sinking caused melting of the contaminated areas of the interior. A few hundred Ma later, continued radioactive heating resulted in partial melting of a zone in the primordial interior near the base of the differentiated sequence. Smaller, less efficiently evolved ilmenite-pyroxenite pods sank more slowly through the differentiated sequence, but upon reaching the hot primordial interior above the partially molten zone, these pods sank rapidly without achieving thermal equilibrium with their surroundings. The infall of these relatively cool pods into the partially molten zone was accompanied by complex assimilative interactions, locally producing hybrid magmas and, perhaps, initiating diapirism. The high-alumina mare basalts were considered by RINGWOOD and KESSON (1976) to have been produced at the crust-mantle interface, when parental magmas from the deep interior assimilated limited amounts of plagioclase-rich crustal material. The dynamic assimilation model was expanded upon by KESSON and RINGWOOD (1976), inasmuch as they felt that **all** mare basalts appeared to be hybrids.

TAYLOR (1982) pointed out inconsistencies with this assimilation model in that the high-alumina basalts have unusually low Ni contents (<2 ppm), which is in contradiction with the dynamic assimilation model. Also, the low Ni abundances in the high-Ti basalts cannot be produced by contamination or assimilation, and there is a good inverse correlation between incompatible-element abundances in mare basalts with those in highland rocks; the dynamic assimilation model predicts a positive correlation. Furthermore, this model does not take into account possible contamination of the magma en route to the surface.

Cumulate "Layer-Cake" (3)

The cumulate model requires that the mare source regions are mafic cumulates from the crystallization of a lunar magma ocean (LMO). The cumulate concept was proposed by SMITH et al. (1970) and PHILLPOTTS and SCHNETZLER (1970), developed by TAYLOR and JÂKES (1974), and explains the complementary geochemical nature of the lunar highlands and mare basalt source regions (e.g., SHIH et al., 1975; SHIH and SCHONFELD, 1976). In the TAYLOR and JÂKES (1974) model, source regions of the low-Ti mare basalts are composed of olivine-orthopyroxene cumulates (deeper, earlier LMO differentiates), whereas the proposed source regions for the high-Ti mare basalts are composed of ilmenite and clinopyroxene (shallower, later LMO differentiates). This requires a "layered" sequence with olivine and orthopyroxene (low-Ti source) at lower levels and clinopyroxene and ilmenite at higher levels (high-Ti source). Prior extraction of plagioclase from the source regions is facilitated by density separation to form the plagioclase-rich lunar crust or highlands. The major drawback in this cumulate remelting model is the difficulty in generating the same MG# for the high-Ti basalts, which should be derived from late-stage cumulates, as are found in the low-Ti basalts, which should be derived from early cumulates (e.g., Fig. 2). Furthermore, geochemical modelling demonstrates that plagioclase did not fractionate from the LMO until 50–70% of it had crystallized. Therefore, the mafic cumulates fractionated prior to this would not exhibit significant negative Eu anomalies, certainly not of the magnitude required to generate those observed in mare basalts. However, an attractive feature of this model is that it can explain the two-stage evolution preserved by radiogenic isotopes.

BINDER (1982, 1985) modified the cumulate model of mare basalt petrogenesis to include assimilation. In this model, the source region is composed of density-graded rhythmic bands of olivine, pyroxene, plagioclase, ilmenite, and chromite which settled out of a convecting magma (i.e., the LMO). Compositions range from the VLT source (0.2% ilmenite) to the high-Ti basalt source (9% ilmenite). Heat from radioactive decay caused melting within these compositional bands. BINDER (1982) suggested that melt viscosity, which decreases rapidly at about 30 ± 5% partial melting, constrained the degree of partial melting of the source regions to about 30 ± 2%. These partial melts rose to the crust-mantle boundary, where they pooled in storage chambers. The stored magma resided in these chambers for varying time periods, fractionating 0–30% olivine or ~30% olivine plus 0–20% pyroxene. Unfractionated magma erupted to the surface to produce the pyroclastic glasses, magma which lost increasing amounts of olivine ± pyroxene in the production of the Luna 16, high-Ti, low-Ti, Luna 24 basalts. The magma storage chambers were situated in a zone where urKREEP formed during the initial lunar differentiation, and urKREEP residuals were assimilated by the stored magma. The assimilation of these residuals produced magmas of differing LREE depletion and Eu abundance, as well as a wide range in REE abundances.

There are several problems with the BINDER (1982, 1985) model of mare basalt petrogenesis. For example, it requires a relationship between the volcanic glasses and the mare basalts (which has yet to be proven) and does not account for the fact that high-alumina Luna 16 basalts are of the same age as the low-Ti mare basalts (Fig. 6). Furthermore, this model does not take into account the fact that mare volcanism lasted at least 1.3 Ga and probably longer. The BINDER (1982, 1985) model, if it is to account for the relationships between ages and compositions of mare basalts (Fig. 6), requires several magma pulses which undergo varying degrees of post-magma-generation evolution (including assimilation and fractional crystallization). The Apollo 14 high-alumina basalts appear to be the oldest and most evolved of the mare basalts. The high-Ti mare basalts are younger and less evolved, while the low-Ti mare basalts are even younger and the least evolved. In the Binder model, when all types are present, it might be possible to explain the sequence by calling on a developing conduit to become enlarged and armored with time, thereby allowing less post-magma-generation evolution of the rising melts with time. However, these basalts are found in different areas on the Moon, and there is no evidence at each site for a progression from high-alumina to high-Ti, to low-Ti, to VLT. Furthermore, as noted above, the high-alumina basalts from Luna 16 appear to be the same age as the low-Ti mare basalts. Such inconsistencies suggest that the overall model of BINDER (1982, 1985) for mare basalt petrogenesis requires considerable refinement.

Cumulate "Convective-Overturn" (4)

This model requires the sinking of the denser late cumulates and convective overturn of the cumulate pile immediately after LMO crystallization (e.g., DOWTY, 1975; DELANO, 1980; HUGHES et al., 1988, 1989; HESS, 1991; SPERA, 1991). RYDER (1991b) demonstrated that in contrast to the mare basalt source parent magma, the ferroan anorthosite parent magma (i.e., the magma which gave rise to the plagioclase-rich lunar crust) was more evolved on the basis of lower Mg/Fe and Ni abundances, but it was less evolved with regard to clinopyroxene constituents, flat (chondritic) REE profile, and incompatible-element abundances. As noted above, the ferroan anorthositic crust would not have crystallized from the LMO until 50–70% solidification, and the cumulate sources crystallized prior to this would not exhibit negative Eu anomalies. RYDER (1991b) agreed with HUGHES et al. (1988, 1989) in that global overturn of a hot, gravitationally unstable mantle immediately followed the solidification of the LMO. Inefficient mixing of primitive and evolved cumulates would, therefore, account for the dichotomy of geochemical signatures exhibited by the lunar crust and the mantle sources of the mare basalts. SNYDER et al. (1992) have proposed a modification of this global overturn scenario, in that convective overturn was on a more localized scale. This model can generate all mare basaltic types, except sources of the high-Al varieties, which require incomplete plagioclase separation during crustal formation.

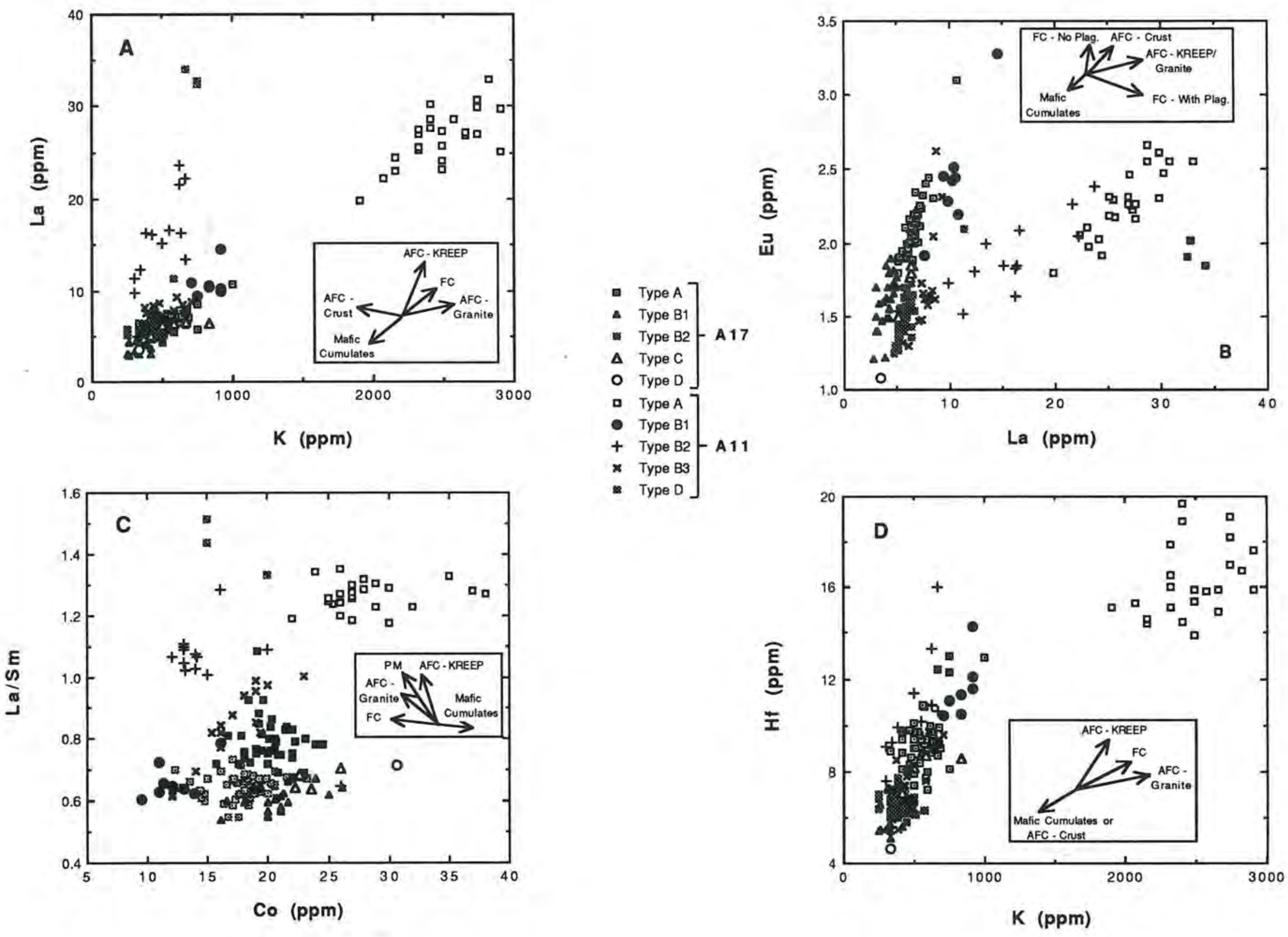
HIGH-Ti/LOW-K MARE BASALTS AND HIGH-Ti/HIGH-K MARE BASALTS

Introduction

The classification of mare basalts on the basis of Ti, Al, and K contents proposed above is a departure from prior conventions. Previous workers have always included the high-K variants of high-Ti mare basalts with the low-K varieties. As such, many of the conclusions regarding the high-Ti/low-K and high-Ti/high-K mare basalts are intricately interwoven. Therefore, in order to eliminate considerable repetition, we will discuss the high- and low-K varieties of high-Ti mare basalts in one section and emphasize that both varieties are low-Al (Fig. 5). We suggest that future studies on high-Ti basalts should consider the high-K variants as a genetically separate class.

High-Ti mare basalts were returned mainly by the first and last manned Apollo missions. The Apollo 11 lunar module landed in the southwestern part of Mare Tranquilitatis, whereas the Apollo 17 lunar module landed in the Taurus-Littrow region at the southeastern margin of the Mare Serenitatis basin (Fig. 1).

At Apollo 11, five groups of mare basalts containing > 6 wt% TiO_2 have been defined (Fig. 7a): Type A = high K,


A
La (ppm)
K (ppm)
AFC - KREEP
FC
AFC - Crust
AFC - Granite
Mafic Cumulates
B
Eu (ppm)
La (ppm)
FC - No Plag.
AFC - Crust
AFC - KREEP/ Granite
Mafic Cumulates
FC - With Plag.
C
La/Sm
Co (ppm)
PM
AFC - KREEP
AFC - Granite
FC
Mafic Cumulates
D
Hf (ppm)
K (ppm)
AFC - KREEP
FC
AFC - Granite
Mafic Cumulates or AFC - Crust
Type A
Type B1
Type B2
Type C
Type D
A17
Type A
Type B1
Type B2
Type B3
Type D
A11

high La; Type B1 = intermediate K, low La; Type B2 = low K, intermediate La; Type B3 = low K, low La; and Type D = low K, high La (Beaty and Albee, 1978; Beaty et al., 1979a,b; Ma et al., 1980; Rhodes and Blanchard, 1980). Beaty and Albee (1978, 1980) and Beaty et al. (1979a,b) demonstrated that the three subdivisions of Type B basalts are real, based upon textures, mineral chemistries, and ages. Crystallization ages were reported by Geiss et al. (1977), Papanastassiou et al. (1977), and Guggisberg et al. (1979): Type A = 3.60 Ga; Type B1 = 3.72 Ga; Type B2 = 3.90 Ga; Type B3 = 3.76 Ga; Type D basalts remain undated. Seventy-three Apollo 11 high-Ti basalts have been chemically analyzed, and seventy of these fit into the classification outlined above. The remaining rocks are not classified because they cannot be distinguished between Types B2 and B3 (Beaty et al., 1979a,b). Of the seventy classified samples, thirty-three are of Type A, three are of Type B1, ten are of Type B2, twenty-one are of Type B3, and three are of Type D. These groups have been considered representative of at least five separate igneous events (Beaty and Albee, 1978, 1980; Beaty et al., 1979a; Rhodes and Blanchard, 1980). The Type A basalts are classified as high-Ti/high-K and will be discussed in more detail below.

The Apollo 17 high-Ti basalts are all low-K and have been divided into a three-fold classification on the basis of texture and into five groups on the basis of whole-rock chemistry. The petrographic divisions have been derived independently by three different studies (Papike et al., 1974; Brown et al., 1975; Warner et al., 1975a), and the differences between the groups defined by each petrographic study are identical. These are Type IA = olivine-porphyritic ilmenite basalt, usually quench textured and olivine-rich; Type IB = plagioclase-poikilitic ilmenite basalt, slower-cooled (coarse-grained) and olivine-poor; Type II = Apollo 11 low-K-type (olivine-free), low-Mg basalt.

On the basis of whole-rock chemistry alone, three types of Apollo 17 mare basalts (A, B, and C) were recognized by Rhodes et al. (1976), and these groupings were reinforced by the work of Warner et al. (1979). It is this chemical scheme of classification that is most commonly referred to in the literature, and although it is alphabetical, as for the Apollo 11 variants, it is evident that corresponding groups between these two sites are not identical. Apollo 17 Type A basalts contain 50–60% higher incompatible trace-element abundances than do the Type B basalts, although they possess similar major-element compositions. Neal et al. (1990a) demonstrated that the Type B Apollo 17 basalts did not comprise one single group, but two. When La is plotted against Eu (Fig. 7b), two groups are clearly defined—Types B1 and B2. Type C basalts contain higher MgO and Cr_2O_3 contents than the Type A and B basalts, although they contain REE contents similar to the Type A basalts (Neal et al., 1990b). Also, Type C basalts contain olivines of the highest Fo content (Fo_{80}) found in Apollo 17 high-Ti basalts, but only five examples have been documented (Rhodes et al., 1976; Warner et al., 1979; Neal et al., 1990b). Ryder (1988, 1990) reported the chemistry and petrography of a sample which may be a new variant of Apollo 17 high-Ti basalt (70091,2161) and which he classified as Type D. This basalt has lower REE and a deeper Eu anomaly than the Type B1 and B2 basalts, but is more MgO- and Cr_2O_3-rich, and CaO- and TiO_2-poor than other Apollo 17 basalts. It is questionable if one sample alone should be incorporated as a new basaltic type.

Petrogenesis of High-Ti/Low-K Mare Basalts

The range in Apollo 11 and 17 high-Ti basalt compositions can be generated only by a combination of source heterogeneity and post-magma-generation evolution. This is emphasized by the range in compositions and the intra-group diversity noted at each high-Ti site. For example, while both Apollo 11 and the majority of Apollo 17 mare basalts are described as being high-Ti and have essentially the same composition (Figs. 2–4), the Apollo 11 high-Ti basalts generally contain lower Ti abundances than those from Apollo 17 (Fig. 2). Major elements plotted against MG# for the high-Ti basalts tend to exhibit general correlations which can be described as fractional crystallization trends. For example, as MG# decreases, Al_2O_3 contents increase in both Apollo 11 and 17 high-Ti basalts (Fig. 3a). This suggests that plagioclase was not a major part of a fractionating assemblage for either suite of basalts, at least not until late in the crystallization sequence.

Beaty and Albee (1978), Beaty et al. (1979a,b), and Rhodes and Blanchard (1980) demonstrated that the various Apollo 11 rock types represent different basaltic units and cannot be related by surface or near-surface crystal-liquid fractionation. This is demonstrated by plotting Co against the La/Sm ratio (Fig. 7c). The range in La/Sm ratio at similar Co abundances of Apollo 11 Types B1, B2, B3, and D mare basalts could not have been generated by fractional crystallization of observed phenocryst phases, although Beaty and Albee (1978) and Beaty et al. (1979a) demonstrated that within-group variations were produced by crystal fractionation. For example, the intra-group variation demonstrated by the Type B2 Apollo 11 basalts can be generated by ilmenite fractional crystallization, and the Type B3's are related by olivine fractional crystallization. This contrast in fractionating minerals may be related to different sizes, morphologies, and settling rates of olivine and ilmenite in the two rock types (Beaty et al., 1979a).

As with the Apollo 11 high-Ti basalts, the compositional differences between the A, B, C, and D basaltic groups at

FIG. 7. Comparison of Apollo 11 and Apollo 17 mare basalt chemistry: (a) K (ppm) vs. La (ppm) demonstrating the original subdivision of the Apollo 11 high-Ti mare basalts into five groups and the restricted composition of the Apollo 17 varieties; (b) La (ppm) vs. Eu (ppm) demonstrating the division of Apollo 17 high-Ti mare basalts into A, B1, B2, and D groups on the basis of La and Eu abundance—note the similarity of Type C basalts (defined primarily using major elements MgO and Cr2O3) to Type A's and the more variable composition of the Apollo 11 basalts; (c) La/Sm vs. Co (ppm) indicating that the various groups of Apollo 11 and 17 basalts cannot be related by fractional crystallization; (d) K (ppm) vs. Hf (ppm) demonstrating relative constant Hf abundances in the Apollo 17 basalts relative to those of Apollo 11. Hypothetical evolution paths are also shown: FC = Fractional Crystallization; AFC – KREEP = Assimilation and Fractional Crystallization with KREEP as the assimilant; PM = Partial Melting.

Apollo 17 cannot be reconciled by near-surface fractional crystallization of liquidus phases. Therefore, these basaltic types require distinct source regions (SHIH et al., 1975; RHODES et al., 1976; WARNER et al., 1979; RYDER, 1988, 1990). However, SNYDER et al. (1990a) have demonstrated the similarity between Apollo 11 Type B3 and Apollo 17 Type C basalts both in major and trace elements, which suggests that they were derived from a similar (if not the same) source. Earlier work has emphasized the processes of partial melting and near-surface fractional crystallization in the petrogenesis of these basalts (e.g., LONGHI et al., 1974; SHIH et al., 1975; RHODES et al., 1976; WARNER et al., 1979). WARNER et al. (1979) proposed that due to variations in the La/Sm ratio (Fig. 7c), all Type B basalts could be modelled by varying the degree of partial melting of a single source. However, relative to the Type A Apollo 17 basalts, the Type B's exhibit little variation in Hf abundances (NEAL et al., 1990b; Fig. 7d). If various degrees of partial melting of a common source were responsible for generating the La/Sm variation in the Type B basalts, the Hf abundances also should vary accordingly. The need for addressing the range in La/Sm ratio in the Type B Apollo 17 basalts has been largely overcome by the splitting of Type B basalts into B1 and B2 (NEAL et al., 1990a). This has dramatically reduced the range in La/Sm ratios (Fig. 7c).

Generation of Apollo 17 Type A, high-Ti basalts was originally modelled by fractional crystallization of observed phenocryst phases (e.g., RHODES et al., 1976; WARNER et al., 1979). This was modified by NEAL et al. (1990a) because of the observation that armalcolite is mantled by ilmenite and olivine by pyroxene, which demonstrates the evolution of the fractionating assemblage. NEAL et al. (1990a) modelled the Type A, B1, and B2 basalts by progressive closed-system fractional crystallization, varying the proportions of olivine, Cr-spinel, armalcolite, ilmenite, and pyroxene. The fractionating phases were changed as fractional crystallization proceeded in accordance with petrographic observations. This demonstrated that nearly all of the Apollo 17 basalt compositions could be generated by up to 40–50% fractional crystallization of these liquidus phases. Each type required a separate source, and isotopic studies demonstrated open-system evolution of Type B2 basalts (PACES et al., 1991).

The trace elements in the high-Ti mare basalts exhibit more diversity than the major elements, mainly in the Apollo 11 variants (Fig. 7). For example, all Apollo 17 high-Ti basalts show little range in K and La (Fig. 7a) and are similar to the Apollo 11 Type B3's. The remaining varieties of Apollo 11 mare basalts contain higher K and/or La. The Apollo 11 high-Ti basalts also contain higher Hf abundances (Fig. 7d) and higher La/Yb and La/Sm ratios (Fig. 7c) than their Apollo 17 counterparts, indicating a more LREE-enriched REE profile. Such observations demonstrate the chemical heterogeneity, at least in the trace elements, within the high-Ti/low-Al/low-K mare basalt suite.

Petrogenesis of High-Ti/High-K Mare Basalts

The Apollo 11 Type A mare basalts comprise the high-Ti/high-K category of our classification. These basalts stand apart from the bulk of other high-Ti basalts on the basis of K abundances (Fig. 4), as well as elevated Hf (e.g., Fig. 7c) and Ba contents. While the elevated K contents could be considered the fractionated counterpart of Apollo 11 Type B1 basalts on the basis of K vs. La (Fig. 7a), this is not the case when elemental ratios (Fig. 7c) or isotopes are considered. Immediate parallels can be made with the only other group of high-K mare basalts, the VHK basalts from Apollo 14 (low-Ti/high-Al/high-K; Fig. 5). The Apollo 14 VHK basalts contain elevated K, Rb, and Ba, and generally Hf abundances relative to their high-alumina counterparts (e.g., SHERVAIS et al., 1985b). Models proposed for VHK basalt petrogenesis favor the assimilation of a lunar granitic component by a high-alumina magma (e.g., SHERVAIS et al., 1985b; NEAL et al., 1989). Such a process accounts for the elevated K, Rb, Ba, and in some cases, Hf abundances of the Apollo 14 VHK basalts.

Can assimilation of lunar granite account for the elevated K contents of Apollo 11 Type A basalts? One of the distinctive features of Apollo 14 VHK basalts is their elevated K and La abundances, which are intermediate between low-K mare basalts (of all types) and lunar granite. The Apollo 11 Type A mare basalts have similar K/La ratios to low-K mare basalts (Fig. 8) rather than the VHK basalts from Apollo 14, which strongly argues against an origin involving granite assimilation. It is evident from Fig. 8 that the Apollo 11 high-Ti/high-K Type A basalts exhibit a fractional crystallization trend extending towards KREEP, which, unlike granite, is enriched in both K and La. Therefore, assimilation of a KREEP component could account for the composition of these high-Ti/high-K mare basalts.

Although KREEP contains elevated REE abundances, it exhibits a LREE-enriched profile. If KREEP assimilation produced the elevated K contents, a substantial proportion (>5%) would have to be included, thereby producing a LREE-enriched profile for these basalts. Unfortunately, all Apollo 11 Type A basalts exhibit relatively LREE-depleted profiles, arguing against KREEP assimilation.

We consider the composition of high-Ti/high-K mare basalts to be the product of source composition. Several authors suggest that high-Ti/low-K and high-Ti/high-K mare basalts can be produced from the same source by varying the degree of partial melting (e.g., DUNCAN et al., 1974, 1976; SHIH et al., 1975; WALKER et al., 1975—see the following text). However, as noted by PACES et al. (1991), high-Ti mare basalt sources also may have been subject to metasomatic processes. Such metasomatism followed by partial melting may have produced the Apollo 11 Type A mare basalts, and if the metasomatic fluid was KREEPy, it would explain the trend towards KREEP shown by these basalts in Fig. 8.

Source Modelling

Source models developed for high-Ti basalts (both low- and high-K) have all utilized the general cumulate model of mare basalt petrogenesis (see previous discussion). These source models can be subdivided into those which require high-Ti mare basalts to be derived from (1) early-stage LMO cumulates; (2) late-stage LMO cumulates; and (3) hybridized sources containing both early- and late-stage LMO cumulates.

Early-stage LMO cumulate source models (1)

The first attempt at pinpointing a source for the high-Ti mare basalts was made by RINGWOOD and ESSENE (1970).

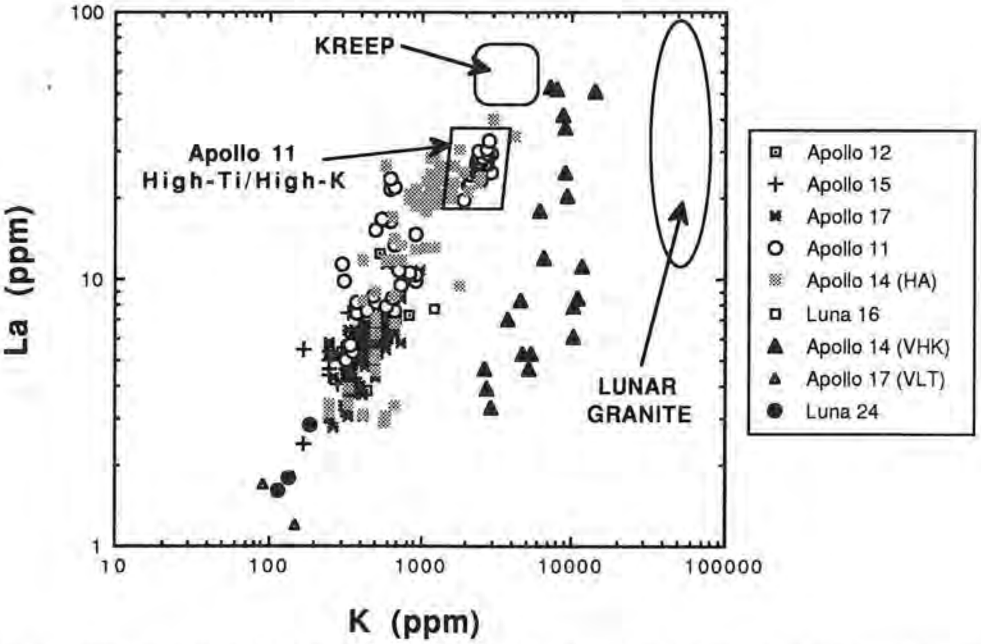


FIG. 8. K (ppm) vs. La (ppm) for all mare basalts. The high-Ti/low-Al/high-K varieties do not appear to contain a granitic component, unlike the low-Ti/high-Al/high-K (VHK) varieties.

These authors used experimental petrology and analyzed mineral phases from Apollo 11 basalts to suggest that the basalts were derived by a 3% partial melt of a source comprised largely of orthopyroxene and pigeonite plus some olivine. These melts were derived from depths between 200 and 400 km. However, it is apparent from Fig. 7 that each magma type seems to require a distinct source region, as most igneous processes do not significantly fractionate K from La nor La from Sm.

Late-stage LMO cumulate source models (2)

More recent models require melting of an ilmenite-rich cumulate from which low-K melts could be produced by larger degrees of partial melting than with the high-K suites (DUNCAN et al., 1974, 1976; SHIH et al., 1975; WALKER et al., 1975). There is some controversy as to whether ilmenite is left as a residual phase or is eliminated from the source during partial melting. DUNCAN et al. (1974, 1976) concluded that ilmenite was exhausted from the high-Ti/low-K basalt source, yet was a residual for the high-Ti/high-K basalts. In the source models of SHIH et al. (1975) and MA et al. (1976), ilmenite is a required residual phase, whereas the high-pressure experimental work of GREEN et al. (1975) indicated that ilmenite was exhausted. GREEN et al. (1975) further suggested that high-Ti mare basalts originated from depths between 200 and 400 km. A problem with such depths of origin was identified by initial experimental work on returned samples of mare basalts (e.g., O'HARA et al., 1970; RINGWOOD and ESSENE, 1970; BIGGAR et al., 1971; GREEN et al., 1971; LONGHI et al., 1974; WALKER et al., 1975). These authors demonstrated that at pressures > 12 kbar and temperatures > 1100°C, high-Ti basalts are transformed to eclogite (garnet + clinopyroxene). Partial melting of eclogite would not be able to produce the REE profiles observed in the mare basalts. Therefore, a source other than eclogite (i.e., without garnet) is required, and depths would have to be shallower.

WALKER et al. (1975) concluded that the source region(s) consisted of late-stage LMO cumulates sandwiched between the feldspathic crust and the complementary mafic cumulate. This source was comprised of olivine, pyroxene, and ilmenite, situated in the outer 150 km of the Moon. WALKER et al. (1975) further concluded that the differences between the Apollo 11 high-K and the Apollo 11 and 17 low-K suites were due to varying degrees of partial melting at different depths. Their source region does not contain plagioclase, and compositional differences arise chiefly from the depths of magma generation.

The variance in the proposed depths of the source regions may be explained by the fact that the samples used in such experiments were not endmember or primary magmas. As has been demonstrated by BEATY et al. (1979a), MA et al. (1980), RHODES et al. (1976), and NEAL et al. (1990a), the variation within each group of high-Ti basalts is produced by fractional crystallization of liquidus phases, and, as such, the melt composition will no longer be a true representation of the source region. Therefore, experiments at high pressure upon such fractionated samples will give misleading information regarding both source depth and mineralogy. The one unifying aspect of petrologically and experimentally formulated models is that high-Ti mare basalts are derived from a source composed of olivine, clinopyroxene, Fe-Ti oxide ± plagioclase ± orthopyroxene (DUNCAN et al., 1974, 1976; LONGHI et al., 1974; SHIH et al., 1975; GREEN et al., 1975; WALKER et al., 1975; NYQUIST et al., 1975, 1976; DRAKE and CONSOLMAGNO, 1976; SNYDER et al., 1990a). However, while these models require clinopyroxene as a residual phase, the isotopic study by PACES et al. (1991) concluded that clinopyroxene in the source must be exhausted, at least in the generation of the Apollo 17 A, B1, and C basalts. For these basalts, source heterogeneities appear to have been produced at ~4.1 Ga by a variable influx of alkalis via halogen-rich metasomatic fluids or KREEP magma (KREEP magmatism occurred at the Apollo 17 site ~4.0–4.1 Ga; COMPSTON et al., 1975; SHIH et al., 1990). The REEs remained unaffected either because no REEs were added during the formation of the source heterogeneity, or Sm-Nd fractionation during partial melting obliterated all evidence. Extension of this type of model to other types of high-Ti basalts is in progress.

Hybridized source models (3)

HUGHES et al. (1989) applied a hybridization model to the petrogenesis of high-Ti basalts. In this model, the sources for each basalt type are composed of an early and a late LMO component, with a proportion of trapped LMO residual liquid (<3.5%). The trapped residual liquid is KREEPy in composition and is considered to be produced after 99% crystallization of the LMO (i.e., urKREEP). All components are mixed by a late-stage convective-overturn event within the LMO cumulate pile. Using this procedure, Apollo 11 high-K (Type A) basalt compositions cannot be explained completely, although appropriate models are obtained for Apollo 11 low- and intermediate-K compositions (Types B3 and B1, respectively). Intermediate- and high-K trace-element patterns can be adequately modelled as magmas initially having a low-K signature, but which were contaminated by ~10% of a KREEPy component and which have experienced variable amounts of olivine fractional crystallization (HUGHES et al., 1989). Source depths are considered by HUGHES et al. (1989) to be <400 km, and these sources would be olivine dominated, but at these depths eclogite (see preceding text) would exist which could not produce the observed trace-element compositions. For Apollo 17 high-Ti basalts, HUGHES et al. (1989) used an average composition in their modelling and concluded that it could be produced by 15% partial melting of a source, again dominated by olivine, with subordinate augite, plagioclase, ilmenite, apatite, and trapped liquid. The percentage partial melt would exhaust the late-stage clinopyroxene, ilmenite, and plagioclase cumulates plus the trapped liquid.

Problems also exist with the scale of the convective overturn in the hybridization model of HUGHES et al. (1989). SPERA (1991) stated that LMO-wide overturn was possible, but SNYDER et al. (1992) considered that the overturn may have been only localized in extent. Another problem with this hybridization model is in the choice of the Apollo 11 basaltic compositions upon which modelling was undertaken; those used did not take post-magma-generation processes into consideration when choosing a parental magma composition. For example, the Apollo 11 intermediate-K basalt quoted by HUGHES et al. (1989) does not fall in any of the groupings defined by BEATY et al. (1979a,b; Fig. 7a). In fact, it plots between Types B1 and A. Furthermore, as is noted above, all Apollo 11 magmatic types have experienced post-magma-generation evolution through fractional crystallization. Likewise with the Apollo 17 high-Ti basalts, an average composition was used which does not take into consideration the fractional crystallization experienced by each high-Ti magma type at this site. However, the conclusion that the percentage of partial melting that is required would exhaust the late stage clinopyroxene, ilmenite, and plagioclase cumulates plus the trapped liquid in the Apollo 17 high-Ti basalt source is consistent with the source models based upon Sr-Nd-Hf isotopic data (UNRUH et al., 1984; PACES et al., 1991) and the experimental work of GREEN et al. (1975), even though such analyses also were undertaken on basalts not considered to represent primary lunar mantle melts.

Synthesis

On the basis of the current data base, it is evident that fractional crystallization has played an important part in the development of the high-Ti basalt compositional range. With the division of Type B Apollo 17 basalts into B1 and B2, fractional crystallization can now be applied to all groups of high-Ti basalts to explain intra-group variation, but it cannot explain the inter-group compositional differences; source heterogeneity is required, although the Apollo 17 Type C and Apollo 11 B3 basalts have probably been derived from the same, or at least a similar, source (SNYDER et al., 1990a). Isotopic work is critical in demonstrating open-system evolution. The Sr and Nd isotope compositions suggest that in at least one high-Ti basalt group (e.g., Apollo 17 B2; PACES et al., 1991), fractional crystallization appears to have been an open- rather than a closed-system process. Can such open-system processes be used to establish a link between any of the ten currently proposed groups of high-Ti mare basalts? It is evident that the Apollo 11 high-Ti source is far more heterogeneous than that represented by the Apollo 17 equivalents. What is the nature of this heterogeneity? The need for this extreme heterogeneity, relative to Apollo 17, is primarily because of Apollo 11 group A, which may well represent a *single* flow. In addressing this question of source heterogeneity, we stress that future source modelling must use primary magma compositions which can be shown to be unaffected by fractional crystallization or open-system processes. This approach also will allow models of convective overturn within the cumulate source region to be more realistically evaluated.

The isotopic and trace-element compositions of the high-Ti mare basalts have initiated a controversy regarding source compositions. The isotopic modelling of PACES et al. (1991) indicates that no clinopyroxene was left as a residual phase in the high-Ti basalt source, but the trace-element modelling of SNYDER et al. (1990a,b, 1992) requires a significant proportion of clinopyroxene in the residue. Both are plausible models, but an integrated isotopic, trace- and major-element modelling effort is required to resolve this dichotomy.

LOW-Ti/LOW-Al/LOW-K MARE BASALTS

Introduction

Low-Ti mare basalts were returned from the Ocean of Storms (Oceanus Procellarum) and the Hadley Rille region near the Appenine Front by the Apollo 12 and 15 missions, respectively (Fig. 1). Four types of mare basalts have been described from the Apollo 12 site and two from Apollo 15.

Initial studies immediately recognized the importance of olivine fractional crystallization in the texturally and chemically diverse suite of Apollo 12 mare basalts (e.g., LSPET, 1970; KUSHIRO and HARAMURA, 1971). JAMES and WRIGHT (1972) subdivided these basalts into three major types on the basis of petrography and major-element chemistry: (1) olivine-pigeonite basalts; (2) ilmenite basalts; and (3) feldspathic basalts. Internal variations were ascribed to near-surface fractional crystallization of olivine (+ minor Cr-spinel; pigeonite replaces olivine in the pigeonite basalts) within thick flows up to 30 m (JAMES and WRIGHT, 1972; RHODES et al., 1977; WALKER et al., 1976; BALDRIDGE et al., 1979). However, the trace-element and isotopic variability was greater within the olivine-pigeonite grouping than expected by simple closed-system fractional crystallization (COMPSTON et al., 1971; PAPANASTASSIOU and WASSERBURG, 1971a). RHODES et al. (1977) assigned the olivine and pigeonite basalts to separate groups (Fig. 9a,b) by highlighting a distinct compositional hiatus between them (although still genetically linking these two groups), making a four-fold classification of Apollo 12 mare basalts: (1) olivine; (2) pigeonite; (3) ilmenite; and (4) feldspathic. RHODES et al. (1977) also classified previously analyzed basalts into this framework (see data sources given in Fig. 2).

Two distinct suites of Apollo 15 low-Ti mare basalts were initially identified on the basis of major-element chemistry (RHODES, 1972; MAXWELL et al., 1972; CHAPPELL et al., 1972; CHAPPELL and GREEN, 1973; HELMKE et al., 1973; CUTTITTA et al., 1973; NAVA, 1974; Fig. 9c,d). These two suites were defined as the "olivine-normative" basalts (ONB) and the "quartz-normative" basalts (QNB). The ONBs have a slightly greater compositional range than do the QNBs, but both have similar ages and isotopic characteristics (NYQUIST, 1977), although average ages suggest that the ONB suite is slightly younger than the QNB suite (3.2 Ga vs. 3.3 Ga). RHODES and HUBBARD (1973) and CHAPPELL and GREEN (1973) concluded that the two groups could not be related by simple fractional crystallization of a common parental magma or partial melting of a common source. BINDER (1976) and BINDER et al. (1980) separated the ONBs into three distinct flow units on the basis of small, but consistent, compositional variations, especially evident when norms were calculated. Variation within each flow unit was postulated to be controlled by olivine fractional crystallization.

Petrogenesis

Taken as a group, the Apollo 12 and 15 mare basalts delineate a consistent fractional crystallization trend of low-Ti basalts (Figs. 2 and 3). Low-Ti basalts from both landing sites contain similar major-element abundances on Figs. 2 and 3. As demonstrated above, previous site-specific studies have identified a number of basaltic groups which may or may not be related. For example, the original classification of olivine and pigeonite basalts as co-magmatic was questioned by RHODES et al. (1977) and can be seen to be incorrect from Fig. 10a,b, where the olivine basalts contain lower Sr and generally lower Rb abundances and, more importantly, higher Co/Sm ratios (11–18) than the pigeonite basalts (Co/Sm = 4–8). The change in Co/Sm ratio between, and the trends exhibited by, the olivine and pigeonite basalts cannot be achieved by simple closed-system fractional crystallization.

DUNGAN and BROWN (1977) concluded that the ilmenite basalt suite was generated by a parental melt which fractionated olivine, producing cumulates and fractionated derivatives. BEATY et al. (1979c) undertook a detailed study of the Apollo 12 feldspathic basalts (12031, 12038, and 12072) and concluded that 12031 could be derived from 12072 through fractional crystallization of Cr-spinel, olivine, and pigeonite (these minerals form the observed phenocryst assemblage). NYQUIST et al. (1979) demonstrated that 12031 is a plagioclase-rich pigeonite basalt and that 12038 is distinct (consistent with the observations of BEATY et al., 1979c), on the basis of initial $^{87}Sr/^{86}Sr$ ratios. NEAL et al. (1992) demonstrated that 12072 was in fact an olivine basalt. It may be that only three basalt groups are present at Apollo 12 (the need for a feldspathic group is disappearing), with 12038 being the "oddball."

Initial studies of Apollo 15 mare basalts (e.g., RHODES and HUBBARD, 1973; CHAPPELL and GREEN, 1973; HELMKE et al., 1973) demonstrated that within groups variation could be accounted for by fractional crystallization of <15% olivine (for the ONBs) and pigeonite (for the QNBs). RHODES and HUBBARD (1973) suggested that the QNBs could have been derived by olivine fractional crystallization from a more primitive magma. This is certainly feasible from the trends observed when La is plotted against Sm (Fig. 10c); but such a relationship is negated by the major elements (Fig. 9c,d), and both the light and heavy REE concentrations and slopes of the REE profiles (given by the La/Yb ratio; Fig. 10d). The differences between the two groups are primary features, related to varying degrees of partial melting within the lunar mantle or derivation from different depths in a heterogeneous mantle.

The olivine-normative mare basalt suite is more widespread than the quartz-normative suite at the Apollo 15 site (RYDER and STEELE, 1988) and has received more attention. MA et al. (1976, 1978a) concluded that neither near-surface fractional crystallization nor varied partial melting of a homogeneous source could explain the observed trace-element trends in the ONBs. These authors suggested that filter pressing of the residual liquid (i.e., short-range unmixing) in a single flow could have been the dominant process, but the data could also be interpreted as representing several flows. In addressing this ambiguity, RYDER and STEELE (1988) concluded that the textures, mineralogical characteristics, and chemical variation of the ONB suite are consistent with a sequence of thin fractionated flows, all from a common parent. This conclusion was generally supported by VETTER et al. (1988), who suggested that the more evolved QNBs can be derived from parental magmas similar to the primitive QNBs by crystal fractionation of olivine, pigeonite, and spinel. Also, VETTER et al. (1988) suggested that the high-SiO_2 ONBs may be parental to the QNB suite. VETTER and SHERVAIS (1989) proposed a dynamic melting model for the petrogenesis of Apollo 15 olivine-normative and quartz-normative basalts. Such a model can account for the differences in major-element chemistry and the similarities in trace-element chemistry between the ONB and QNB suites. In this model, melt extraction from the mantle source was incomplete. Between 5–10% of the melt from each melting event remained in the residuum as dikes and veins (cf. LANGMUIR et al., 1977). During subsequent remelting events, trace-element abundances were controlled by the remelting of the dike/vein material. Major-element compositions were controlled

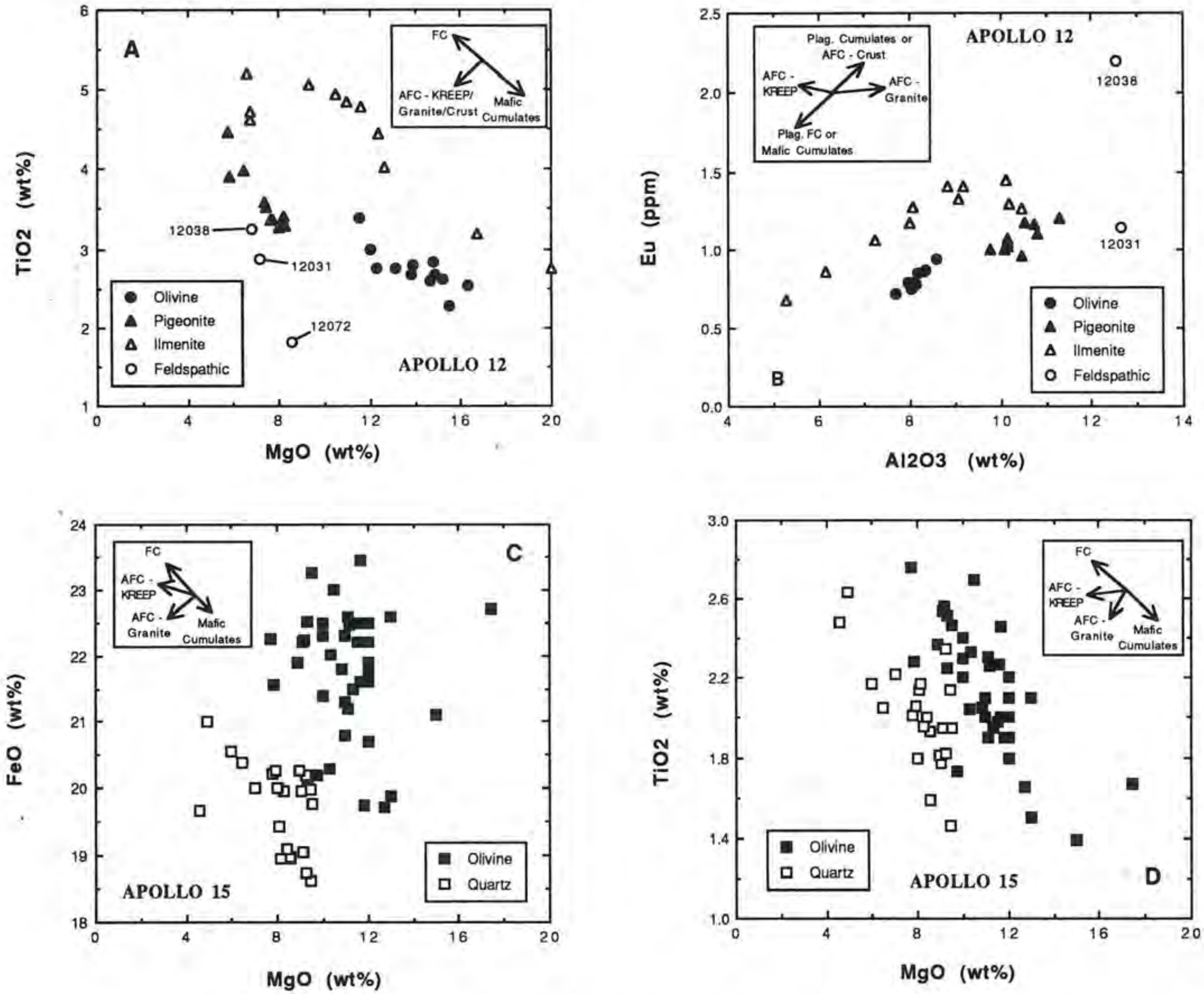


FIG. 9. Division of the low-Ti basalts into groups according to site. (a, b) = division of Apollo 12 basalts into olivine, pigeonite, ilmenite, and feldspathic groups on the basis of MgO (wt%) vs. TiO_2 (wt%) and Al_2O_3 (wt%) vs. Eu (ppm); (c, d) = division of Apollo 15 basalts into olivine-normative and quartz-normative groups on the basis of MgO (wt%) vs. FeO (wt%) and TiO_2 (wt%). Hypothetical evolution paths are also shown (descriptions are the same as in Fig. 7).

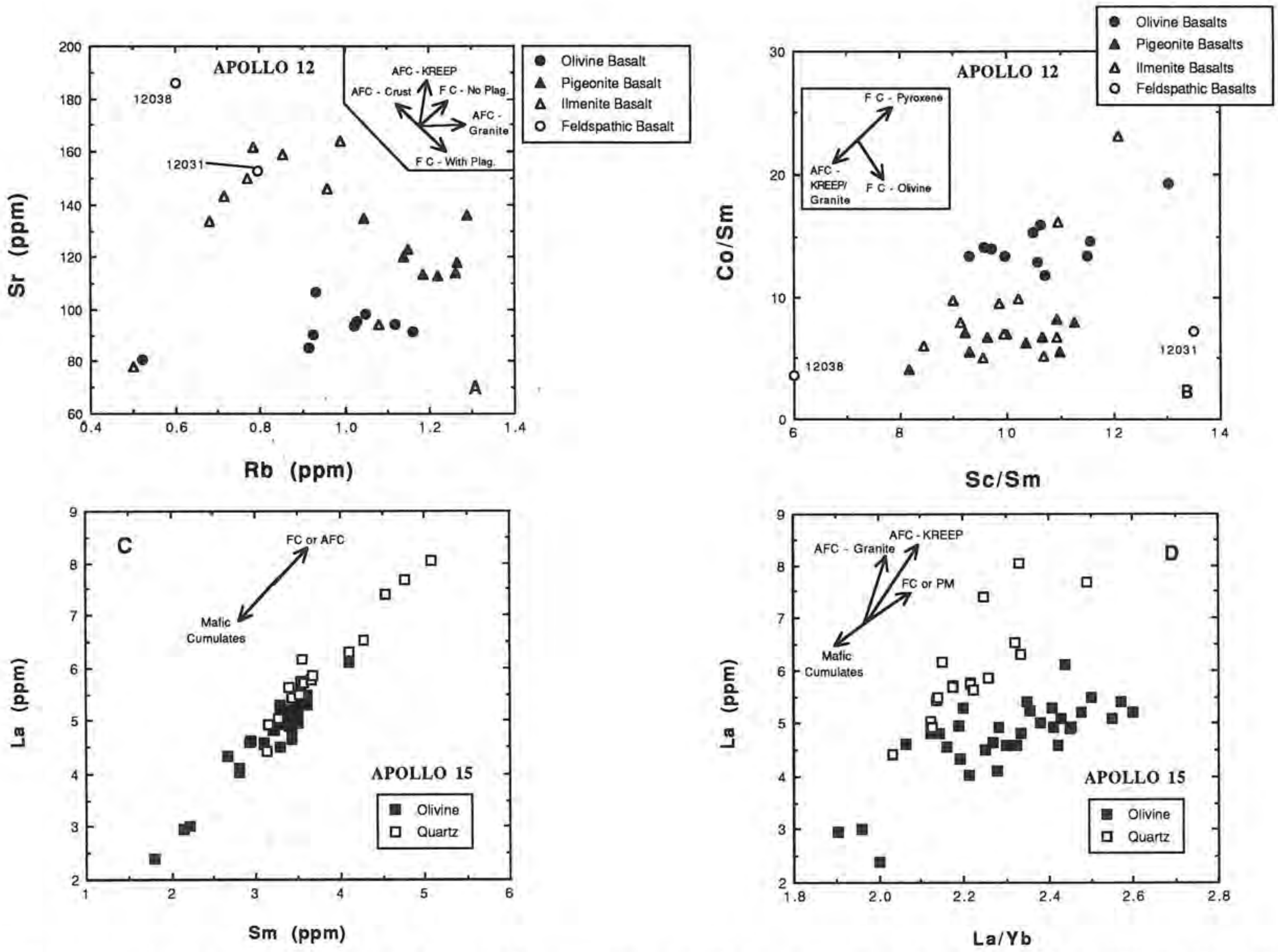


FIG. 10. Trace element evidence demonstrating that the Apollo 12 olivine and pigeonite basalts are not comagmatic and the Apollo 15 QNBs and ONBs could be related by fractional crystallization: (a, b) = Rb (ppm) vs. Sr (ppm) and Sc/Sm vs. Co/Sm (ppm), demonstrating that the Apollo 12 olivine and pigeonite are not related by fractional crystallization; (c, d) = Sm (ppm) vs. La (ppm) and La/Yb vs. La (ppm) demonstrating that although both groups of Apollo 15 low-Ti mare basalts form a single, continuous fractional crystallization trend on the basis of La and Sm, when the La/Yb ratio is plotted against La (ppm), it is evident that the QNBs and ONBs cannot be related by closed-system fractional crystallization. Note that if La/Yb > 1.5, the REE profile is LREE enriched. Hypothetical evolution paths are also shown (descriptions are the same as in Fig. 7).

by phase proportions and compositions in the refractory residuum. By definition, this refractory residuum was enriched in MgO and depleted in SiO_2, and, as such, subsequent melts derived from this would be more mafic and silica-undersaturated. However, subtle, yet distinct, differences present in the trace-element chemistry of the ONB and QNB suites, as noted above, may indicate that similar, yet separate, sources are responsible for the Apollo 15 low-Ti mare basalt suite.

Source Modelling

There have been several models developed in order to explain the source regions of Apollo 12 and Apollo 15 low-Ti mare basalts. Almost all require multiple source regions which are generally similar in character. For example, DUNCAN et al. (1976) concluded that low-Ti basalts were produced from a primitive Moon source, or at least one with whole Moon incompatible-element ratios. MA et al. (1976, 1978a) suggested on the basis of REE considerations that low-Ti mare basalts were derived from a more evolved source rich in clinopyroxene and containing no plagioclase. SHIH and SCHONFELD (1976) calculated a cumulate source for low-Ti basalts composed of olivine, clinopyroxene, and orthopyroxene, with minor plagioclase (or spinel at greater depths). However, if plagioclase flotation was required to form the lunar crust, the presence of plagioclase in the low-Ti source region would be problematical, especially as it is postulated to reside below the high-Ti basalt source (e.g., TAYLOR and JÂKES, 1974; TAYLOR, 1982). This problem was addressed by the Rb-Sr and Sm-Nd isotopic and REE modelling of NYQUIST et al. (1977, 1979), who concluded that in the case of the Apollo 12 basalts, the isotopic and REE compositions can be generated by small (2–10%) degrees of partial melting of cumulate sources composed of varying proportions of olivine, orthopyroxene, and clinopyroxene, but no plagioclase. An important feature of this model is that the parental LMO liquid had a LREE-enriched profile at the time of cumulate source formation.

The hybridization model of HUGHES et al. (1988) has the potential to overcome the problem of a fractionated REE pattern in the low-Ti mare basalt source by the mixing in late-stage cumulates and KREEPy inter-cumulus liquid through convective overturn of the cumulate pile. Using the same approach as in the high-Ti basalts, HUGHES et al. (1988) conducted source modelling of the Apollo 12 olivine basalts and the Apollo 15 ONBs. Constraints on Apollo 12 olivine basalt source regions, in light of Sm/Nd evolution (NYQUIST et al., 1979), show that significant LREE fractionation occurred at the time of magma generation. This requires residual pyroxene, and HUGHES et al. (1988) indicated a slight preference for orthopyroxene in the residuum with low degrees of partial melting ($F = 0.03$), consistent with Sm/Nd data. However, this modelling may be flawed in that average basaltic compositions again were used, rather than compositions which may represent primary lunar mantle melts.

The important observation of the low-Ti mare basalts is that they have been derived from a variety of source compositions. This is evident in the major- and trace-element differences between all types of Apollo 12 and Apollo 15 low-Ti basalts (Fig. 11a–d). Such observations again demonstrate the heterogeneity of the lunar mantle, as evidenced even within basalts of broadly similar composition.

Synthesis

As with the high-Ti mare basalts, there are a number of groups of low-Ti basalts defined by the current data base. Unlike the high-Ti basalts, there are no similarities between the groups from Apollo 12 and those from Apollo 15. Closed-system fractional crystallization is the main process responsible for intra-group variation, requiring a storage chamber or relatively thick flows (up to 30 m). Trying to obtain such flow thicknesses within the lunar maria is problematical, due to the low viscosity of mare basalt lavas, unless some sort of ponding occurred.

On the basis of our compiled data base, it is evident that there is no petrogenetic relationship between the Apollo 12 olivine and pigeonite basalts, and no connection via fractional crystallization between the Apollo 15 ONBs and QNBs. Both major- and trace-element data demonstrate source heterogeneity. The convective overturn hypothesis of HUGHES et al. (1988, 1989) and SPERA (1991) is the most convenient model for producing such source heterogeneities. However, as with high-Ti mare basalts, low-Ti source compositions must be determined using basaltic compositions unaffected by post-magma-generation processes; taking an average composition will yield misleading results. This point is carried over to the "dynamic melting" model of VETTER and SHERVAIS (1989) who use the same source to produce the Apollo 15 ONBs and QNBs, but have problems in reproducing major-element abundances. On the basis of current data, we feel it would be simpler to infer two similar, but distinct, source regions for the Apollo 15 low-Ti basalts. If the problems of major-element abundances in the dynamic melting model are overcome, it will then warrant more detailed consideration.

LOW-Ti/HIGH-Al/LOW-K MARE BASALTS

Introduction

Low-Ti/high-Al or high-alumina mare basalts were returned by the Apollo 14 and Luna 16 missions. The Apollo 14 mission returned samples from the Fra Mauro formation on the western side of the lunar equator (Fig. 1) in an attempt to sample Imbrium ejecta. The Luna 16 module landed in Mare Fecunditatis on the equator at the eastern extremity of the nearside. The study of these Apollo 14 and Luna 16 basalts has been hampered by sample size; the majority of basalts returned by Apollo 14 are present as clasts in breccias, and the automated Soviet Luna 16 probe only returned a small regolith sample (100 g) with basalt fragments of <100 mg.

High-alumina mare basalts from Apollo 14 contain 11–14 wt% Al_2O_3, <0.3 wt% K_2O, and K/La ratio ~100 (e.g., WARNER et al., 1980; DICKINSON et al., 1985; SHERVAIS et al., 1985a; NEAL et al., 1988a, 1989). The importance of the Apollo 14 site is highlighted by the fact that the oldest mare volcanism recognized to date (~4.3 Ga) is recorded here (TAYLOR et al., 1983; DASCH et al., 1987; Fig. 6).

SHERVAIS et al. (1985a) studied high-alumina basalt clasts from breccia 14321. These "14321-type" basalts are similar to other Apollo 14 high-alumina, low-Ti basalts (e.g.,

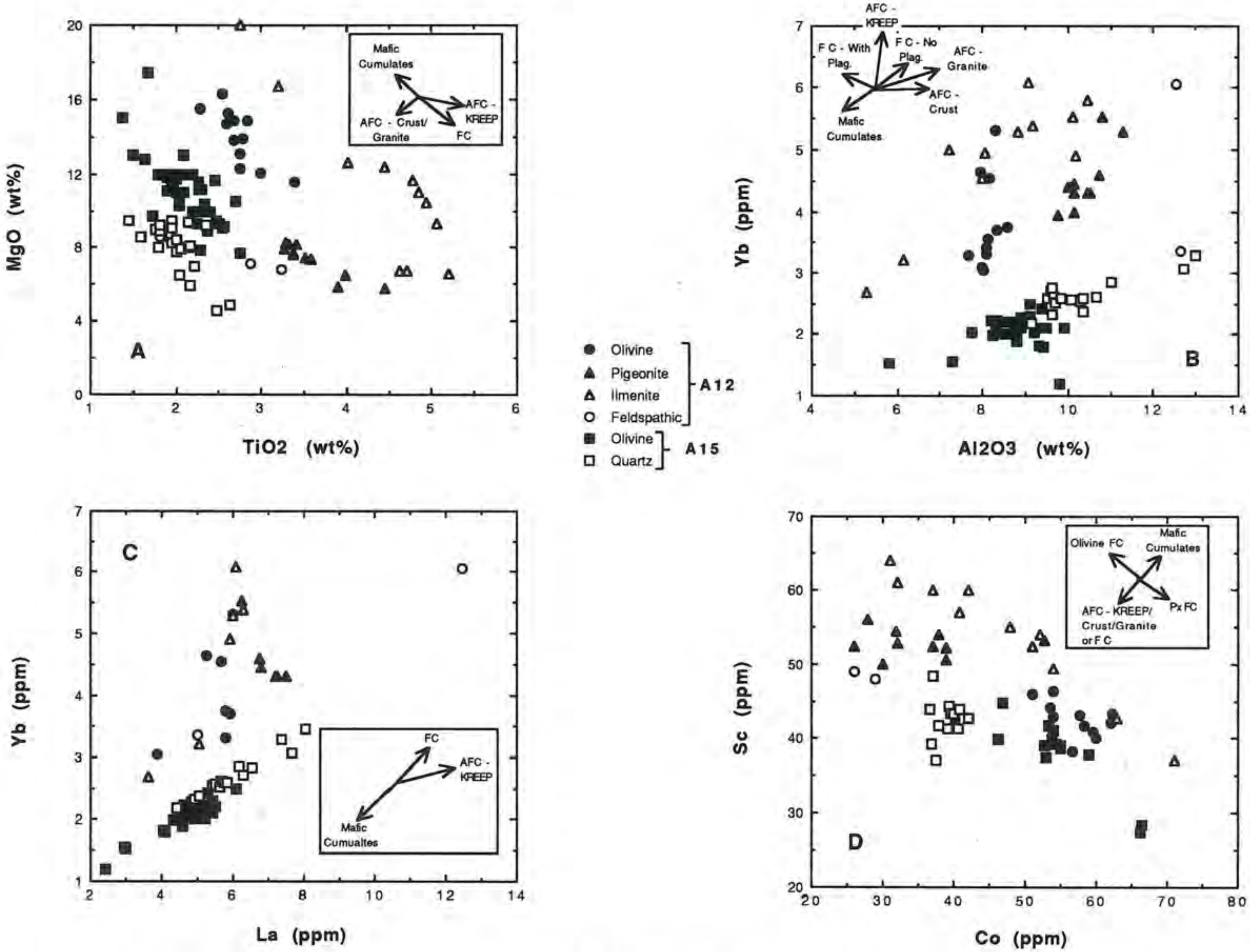


FIG. 11. The four plots presented here all demonstrate that although the Apollo 12 and Apollo 15 mare basalts are all classified as low-Ti, they are not of the same composition, whether major elements are considered, or the compatible and incompatible trace elements. Hypothetical evolution paths are also shown (descriptions are the same as in Fig. 7).

HELMKE et al., 1972; HUBBARD et al., 1972) in major-element chemistry, but are enriched in the incompatible trace elements relative to others (e.g., 14053 and 14072). Both SHERVAIS et al. (1985a) and DICKINSON et al. (1985) identified five groups of Apollo 14 high-alumina basalts based on whole-rock REE chemistry. These ranged from LREE-depleted to LREE-enriched varieties, and the two classification schemes are almost totally interchangeable (Fig. 12a,b). As more data were gathered, it became evident that the division of the high-alumina basalts into five groups was not justified (NEAL et al., 1988a, 1989). Rather, a continuum of high-alumina compositions exists at the Apollo 14 site (Fig. 13b).

The data base for Luna 16 high-alumina basalts is not as large as for the Apollo 14 varieties. KEIL et al. (1972) suggested that Luna 16 mare basalts could be divided into high- and low-K varieties, similar to high-Ti basalts. Unfortunately, we see no evidence to suggest that there are two groups of basalts divisible on the basis of K at Mare Fecunditatis, but there may be two groups present on the basis of MgO, TiO_2, and Hf abundances (Fig. 13a,b); one group (9–12 wt% MgO) is similar to the Apollo 14 high-alumina basalts, the other is distinct (3–7 wt% MgO).

Although basalts from Fra Mauro and Mare Fecunditatis are classified as high-alumina, they are not of the same composition. TiO_2 and FeO abundances are similar (Figs. 13a and 14d), but Luna 16 basalts are generally lower in MgO and Cr than the Apollo 14 varieties (e.g., Fig. 14a,d), and higher in Al_2O_3 and CaO (Fig. 14b,c).

Petrogenesis

Both the Luna 16 and Apollo 14 high-alumina mare basalts broadly define potential fractional crystallization trends (Fig. 14a–c). When it was considered that various groups of high-alumina basalts were present at Apollo 14, both SHERVAIS et al. (1985a) and DICKINSON et al. (1985) noted that their

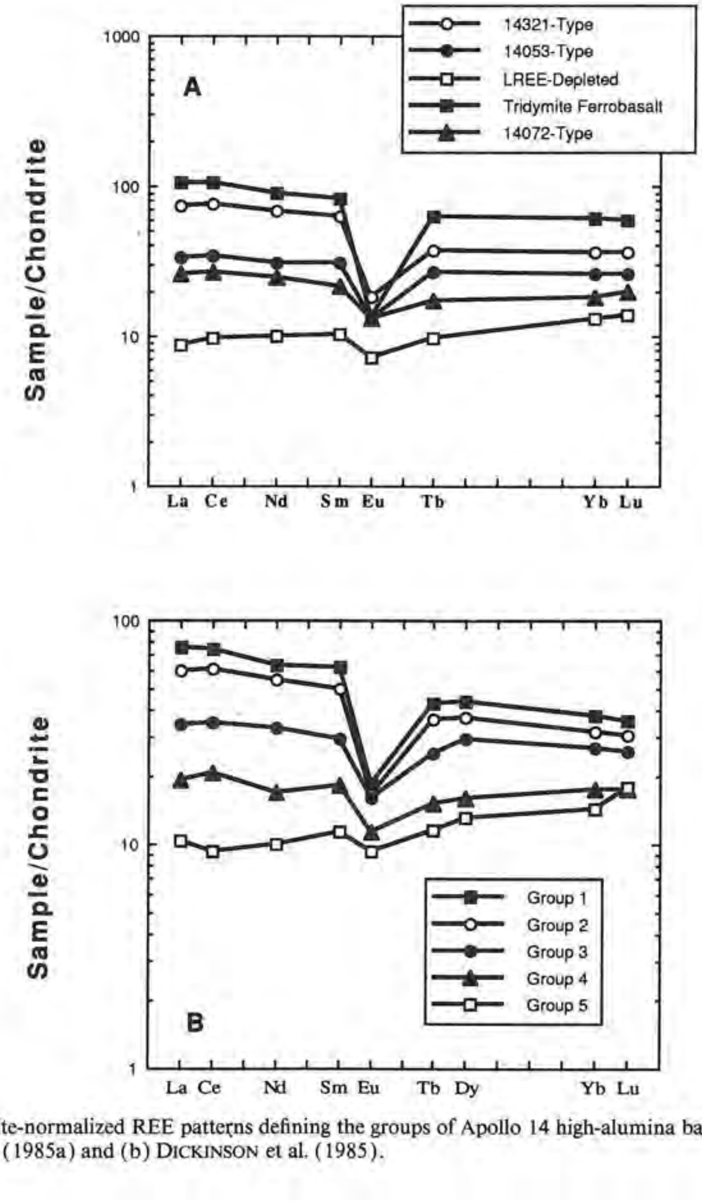


FIG. 12. Chondrite-normalized REE patterns defining the groups of Apollo 14 high-alumina basalts as reported by (a) SHERVAIS et al. (1985a) and (b) DICKINSON et al. (1985).

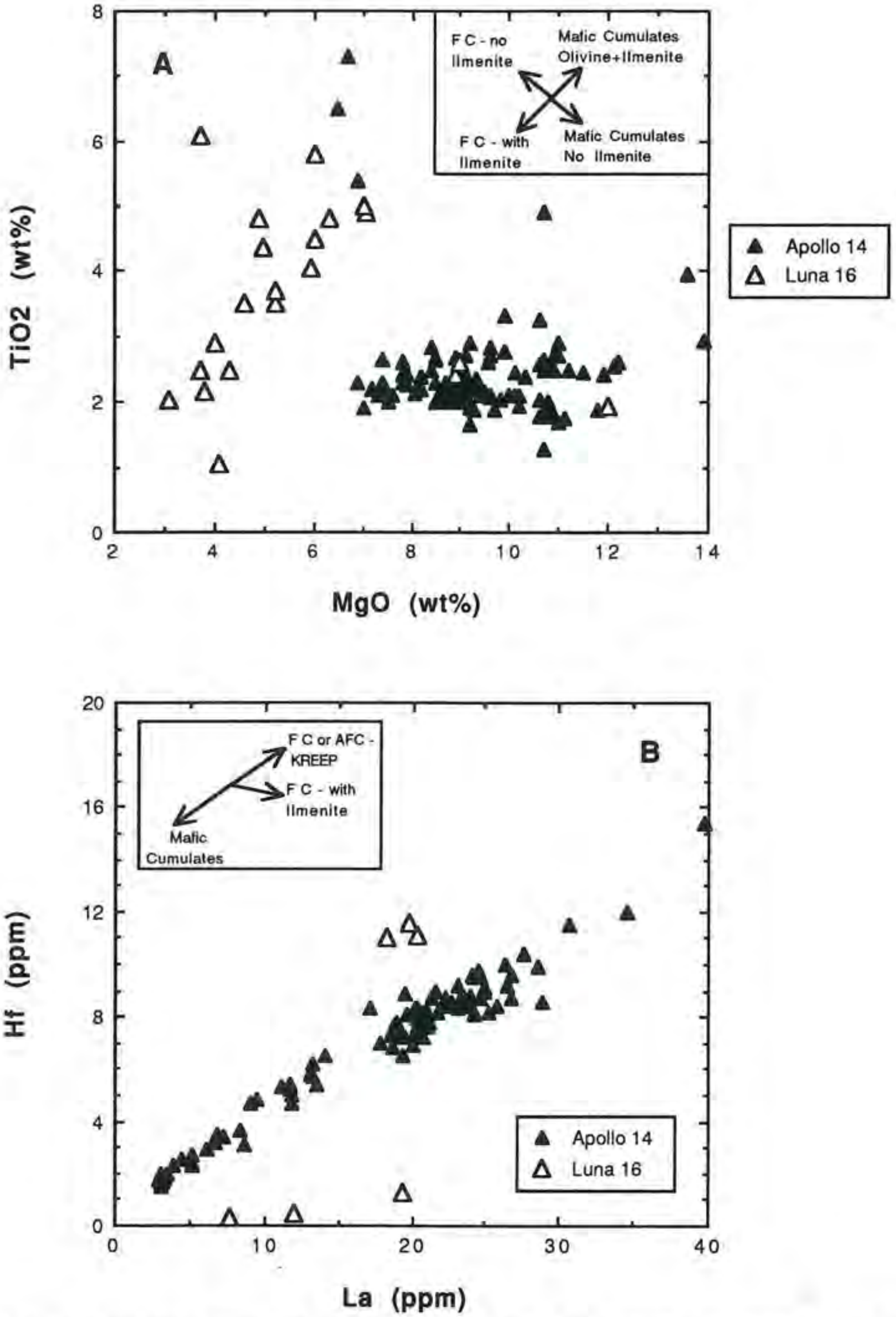


FIG. 13. Major- and trace-element plots of the high-alumina basalts, demonstrating the possibility of two groups at Luna 16 but a continuum at Apollo 14. Hypothetical evolution paths are also shown (descriptions are the same as in Fig. 7).

individual groups could not be related by fractional crystallization. Processes such as variable degrees of partial melting and KREEP assimilation were invoked to generate the observed compositions. NEAL et al. (1988a, 1989) expanded the concept of KREEP assimilation as a process to generate the continuum of Apollo 14 high-alumina basalt compositions (Fig. 13b). These authors used the LREE-depleted ("Group 5") olivine basalts as a parental magma which assimilated KREEP while fractionating the liquidus assemblage (AFC: $r = 0.22$ = mass assimilated/mass crystallized), with all observed compositions being generated by up to 15.5% KREEP assimilation in consort with 70% fractional crystallization. However, DASCH et al. (1987) demonstrated that the mare basaltic volcanism represented by samples returned from Apollo 14 lasted approximately 300 to 400 Ma. Clearly, a single AFC process could not have operated for this period of time. DASCH et al. (1987) concluded that both KREEP assimilation and partial melting of a bulk Moon source appear

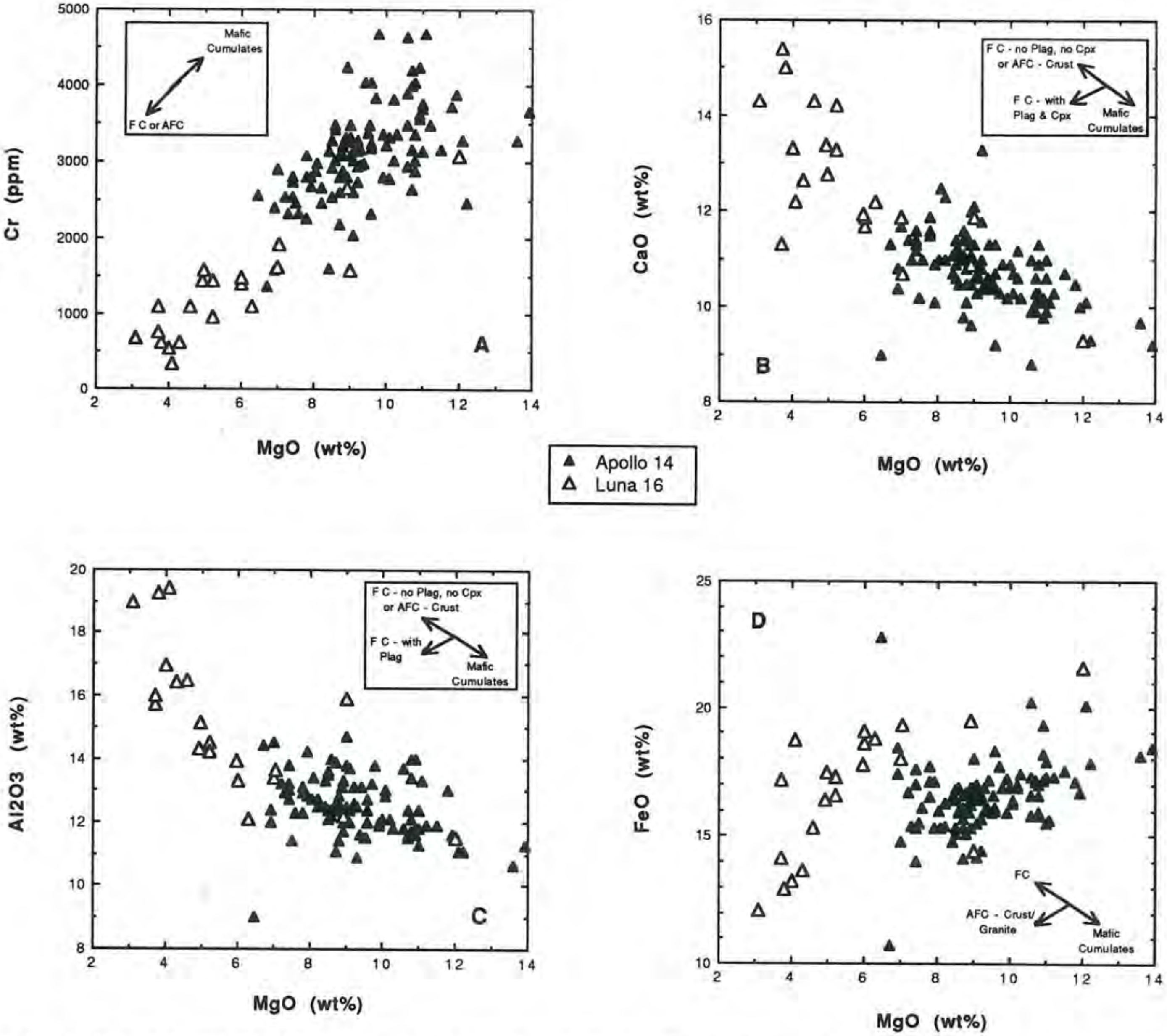


FIG. 14. Element-element plots of high-alumina mare basalts demonstrating the differences in composition between Apollo 14 and Luna 16 high-alumina basalts. Hypothetical evolution paths are also shown (descriptions are the same as in Fig. 7). See text for discussion.

to be reasonable mechanisms for forming the groups of basalts defined by DICKINSON et al. (1985).

NEAL and TAYLOR (1990) modified the AFC model using the Sr isotope data and ages reported by DASCH et al. (1987) and PAPANASTASSIOU and WASSERBURG (1971b). A cyclical AFC model, using the parameters given above, was developed for Apollo 14 high-alumina basalt petrogenesis. AFC events were proposed at 4.3, 4.1, and 3.9 Ga, and basalts of all compositions would occur at each proposed age. Although the trace elements and Sr isotopes of the Apollo 14 high-alumina basalts can be modelled by KREEP AFC, the major elements are more problematical (Figs. 13a and 14), but if the isotopes do represent cyclical AFC, this could explain the lack of well-defined major-element trends in the Apollo 14 high-alumina basalt data (e.g., Fig. 14). Several overlain AFC cycles may be present, in which case the major elements would not be as systematic in their behavior as the trace elements. SHIH and NYQUIST (1989) suggested that the data distribution for Apollo 14 basalts could be explained by a decoupling of the assimilation and fractional crystallization processes. Their model requires that KREEP assimilation occurred first to elevate La abundances, followed by olivine fractional crystallization (<15%) to give the range in MG# within each broad group (Fig. 15d). SHIH and NYQUIST (1989) also suggested that the variation in La (Fig. 15d) may be due to differences in degrees of partial melting, or that the samples were derived from sources containing different La abundances, followed by olivine fractional crystallization. However, most of these basalts had been analyzed by INA where the Mg precision is poor; NEAL and TAYLOR (1990) suggested that the errors associated with the calculation of the MG# would make definitive modelling of this parameter impossible.

Due to the smaller data base and minute sample size, the petrogenetic interpretation of Luna 16 basalts has not been as detailed. BENCE et al. (1972) and KURAT et al. (1976) suggested that the range in basaltic compositions could be explained by near-surface fractional crystallization involving plagioclase, olivine, and ilmenite, but the Luna 16 basalts could not be produced by fractional crystallization of low-Ti or high-Ti parental compositions. Note that these basalts define a trend of increasing Al_2O_3 (Fig. 14c) and decreasing MgO and FeO (Fig. 14d), both consistent with crustal assimilation, but these samples are so small that the analyses are statistically troubling. Due to these sampling problems, the Luna 16 data base should only be used sparingly at best for petrogenetic interpretation, and only then to test petrogenetic models, not formulate them.

Source Modelling

In spite of the aluminous nature of these basalts, significant plagioclase need not have been present in the source. As pointed out by RIDLEY (1975), >10–15% plagioclase requires that the source have a marked positive Eu anomaly. WARNER et al. (1980) suggested that pyroxene was a residual phase in the source region of low-Ti, high-alumina because of the relative abundances of the REEs and Sc.

SHERVAIS et al. (1985a) concluded that 3–30% partial melting of a cumulate source consisting of predominantly olivine, orthopyroxene, clinopyroxene, and minor plagioclase plus 2% trapped liquid (all derived from a LREE-enriched LMO) could produce all Apollo 14 high-alumina basalts, except for their LREE-enriched 14321-type (see Fig. 12a). In a similar approach, DICKINSON et al. (1985) used the 12038 source (87% olivine, 8% plagioclase, 5% orthopyroxene; determined by UNRUH et al., 1984) as a feasible source composition for all their Apollo 14 high-alumina basalt groups. All of these authors have concluded that KREEP assimilation could have played a role in the evolution of these basalts. As such, many of the basaltic compositions do not represent primary melts and cannot be used to calculate source compositions. This was emphasized by the AFC model of NEAL et al. (1988a, 1989) and NEAL and TAYLOR (1990), and one of the results of this model was to identify a parental composition which contained no identifiable KREEP contamination. This composition is equivalent to the LREE-depleted variants described by DICKINSON et al. (1985; "Group 5"), SHERVAIS et al. (1985a; LREE-depleted olivine basalts), and NEAL et al. (1988a, 1989).

HUGHES et al. (1990) applied the convective-overturn model to the parental Apollo 14 high-alumina composition, identified by the AFC model, in order to estimate the source mineralogy. They described the source of this parental high-alumina magma as being dominated by olivine, with minor late-stage LMO phases of clinopyroxene, pigeonite, plagioclase, and 0.2–0.8% trapped liquid. These authors further concluded that even though the parental magma had been LREE depleted, the model required a hybridized source containing a KREEPy component. This supported the argument of DICKINSON et al. (1989), who suggested a metasomatized source for the Apollo 14 high-alumina basalts on the basis of Ge abundances. Further evidence supporting the idea of hybridized sources was presented by SHERVAIS and VETTER (1990), who concluded that these mantle sources already contained the KREEPy component such that post-magma-generation KREEP assimilation was not required.

The source material for Luna 16 high-alumina basalts is distinct, but not totally dissimilar from that for the Apollo 14 variants. The source for Luna 16 basalts has been placed below the crust but above those source regions for other mare basalts, between 60–100 km deep (REID and JÂKES, 1974; TAYLOR and JÂKES, 1974; BINDER, 1976). KURAT et al. (1976) concluded that the source for the Luna 16 basalts was between 40–100 km and was not unlike ferroan anorthosite in some properties (e.g., plagioclase and Sr-rich). MA et al. (1979) stated that the REE content of the Luna 16 high-alumina basalts requires that the source material crystallized from a LREE-enriched magma. The basalts were produced by ~15–30% partial melting of a source rich in clinopyroxene and plagioclase (MA et al., 1979).

Synthesis

Luna 16 high-alumina basalt petrogenesis is somewhat sketchy due to the relatively smaller data base and the relatively small size of analyzed samples. We agree with BENCE et al. (1972) that the array of compositions is due to near-surface fractional crystallization, but it is not clear whether this involved one or several magmas, or if fractional

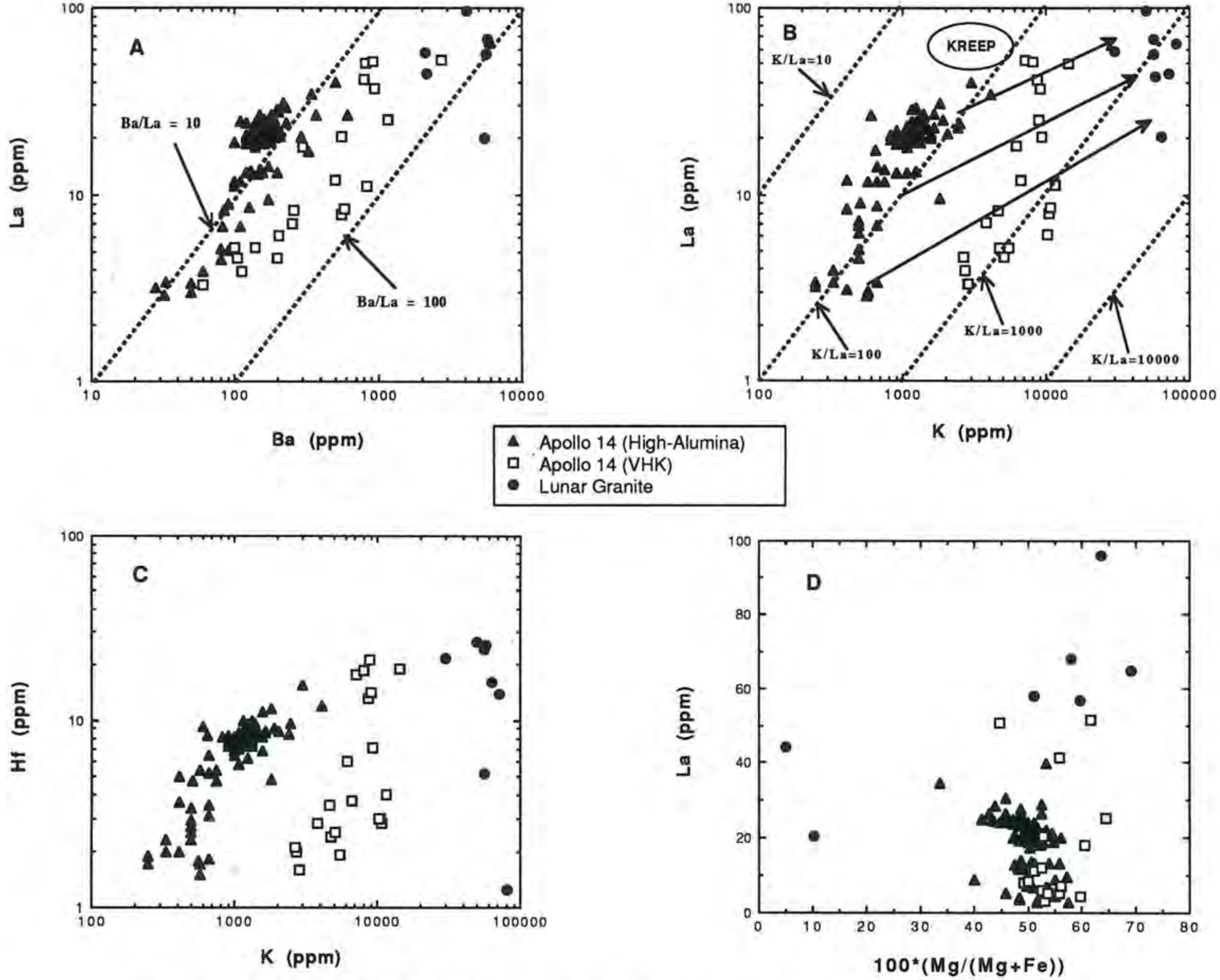


FIG. 15. Element-element plots for the very-high-potassium (VHK) basalts compared with the Apollo 14 high-alumina basalts and lunar granite. The filled arrows in 15b represent AFC paths. See text for discussion.

crystallization occurred in subcrustal magma chambers or in flows at the surface.

The petrogenesis of Apollo 14 high-alumina basalts is controversial. The range in Sr isotopes and continuum of trace-element compositions is consistent with open-system evolution (i.e., AFC with KREEP), but the range in ages of these basalts is not compatible with a single AFC cycle. Such age ranges tend to support a hybridized source model for the Apollo 14 high-alumina basalts, in that the KREEP component was already present within the source regions (HUGHES et al., 1990). The hybridized source model finds support in the convective-overturn hypothesis of HUGHES et al. (1988, 1989), which has the ability to mix KREEPy components into the Apollo 14 high-alumina basalt source region. On the basis of the present data base, we cannot totally refute either the AFC or hybridized source models for high-alumina basalt petrogenesis at Apollo 14, although it is significant that the Sr isotopes are compatible with several (at least three) AFC cycles.

LOW-Ti/HIGH-Al/HIGH-K MARE BASALTS

Introduction

A subset and less common member of the low-Ti, high-alumina basalt suite at Apollo 14 is that of the very-high-K or VHK basalts (e.g., WARNER et al., 1980; SHERVAIS et al., 1984b, 1985b; GOODRICH et al., 1986; SHIH et al., 1986; NEAL et al., 1988b). For major elements, the VHK basalts are essentially the same composition as the high-alumina basalts. But the VHK basalts contain higher (0.3 wt%) K_2O, K_2O/Na_2O (>1), and K/La (>150). These basalts also are characterized by unusually high concentrations of Rb and Ba and consequent high ratios of these elements to other incompatible elements exceeding "normal" lunar values (SHERVAIS et al., 1985b). K-feldspar and K-, Si-rich glass are ubiquitous components of these basalts. Also, both Fo- and Fa-rich olivines are present, and plagioclase compositions exhibit a larger range (e.g., An_{75-95}) than in high-alumina basalts.

Petrogenesis

The VHK basalts have essentially the same major-element chemistry as other members of the high-alumina basalt suite, except for elevated K. SHERVAIS et al. (1985b) concluded that long-term alkali-element enrichment of the source region is inconsistent with high Ba/La ratios (Fig. 15a) and low initial $^{87}Sr/^{86}Sr$ ratios (SHIH et al., 1986) of the VHK basalts. A model for the VHK basalts involving partial assimilation of lunar granite by a high-alumina parental magma was proposed by SHERVAIS et al. (1984b, 1985b; Fig. 15b). Results from isotopic studies of VHK basalts demonstrated an approximately ten-fold fractionation of Rb/Sr, but no significant Sm/Nd fractionation relative to the source materials at the time of crystallization (SHIH et al., 1986, 1987). SHIH et al. (1986, 1987) supported the conclusions of SHERVAIS et al. (1984b, 1985b) that the isotopic data could be adequately modelled by a high-alumina parental magma assimilating up to 5% lunar granite. This model was supported by WARREN et al. (1986) and by the study of GOODRICH et al. (1986), who further suggested that the assimilated material could also be a fractionated derivative of urKREEP.

NEAL et al. (1988b, 1989) proposed an integrated assimilation and fractional crystallization (AFC) model for VHK basalt petrogenesis. It was suggested by these authors that KREEP and granite were intimately related at the Apollo 14 site (after the K-fraction / REEP-fraction model of NEAL and TAYLOR, 1989). This required that a high-alumina basalt magma initially evolved by KREEP assimilation, but at various points along this AFC path, granite replaced KREEP as the assimilant, producing VHK "offshoots" (Fig. 15b). As the major-element chemistry is generally the same for both VHK and high-alumina basalts, the proportions and compositions of fractionating phases in this AFC model are the same as in the high-alumina basalt AFC model. The difference is in the ratio of mass assimilated / mass crystallized, which is considerably higher for the VHK basalts ($r = 0.5$ vs. 0.22) in keeping with the lower melting point of granite. As can be seen from Fig. 15a–c, no single high-alumina basalt can generate all observed VHK basalt compositions. NEAL et al. (1988a, 1989) concluded that at least three VHK basalt flows exist at Apollo 14, consistent with the isotopic work of SHIH et al. (1986, 1987).

Synthesis

As demonstrated by SHERVAIS et al. (1985b), the trace-element and high-K abundances of VHK basalts are not a direct product of partial melting, so source modelling on these basalts is not valid. The petrogenesis of VHK basalts appears to be relatively straightforward, although refinements may have to be made. All evidence points toward a parental high-alumina magma assimilating lunar granite, but the range in VHK basalt compositions requires at least three AFC episodes. The required high-alumina magmas must be primitive (low incompatible-element abundances), intermediate, and evolved (high incompatible-element abundances) in order to produce all reported VHK compositions.

VERY LOW-Ti/LOW-Al/LOW-K BASALTS

Introduction

Very-low-Ti or VLT basalts were returned from the Taurus-Littrow region by the Apollo 17 mission and from Mare Crisium by the Luna 24 mission (Fig. 1). The VLT basalts were first discovered in thin sections from the Apollo 17 deep drill core (VANIMAN and PAPIKE, 1977; TAYLOR et al., 1977), but have since been found in impact melt rocks at the Apollo 17 site (WARNER et al., 1978; TAYLOR et al., 1978) and as basaltic clasts in breccias 73255 and 72235 (JAMES and MCGEE, 1980). Indeed, the VLT mare basalts are the most abundant basaltic type returned by the Luna 24 mission (e.g., MA et al., 1978b; RYDER and MARVIN, 1978).

Petrogenesis

Unravelling the petrogenetic history of any mare basalt is problematical, but it is acutely so with the VLT varieties because of their small sample size—only a few milligrams of sample have been analyzed for whole-rock chemistry.

Therefore, caution is required in interpreting the trace-element chemistry because of short-range unmixing considerations. For example, the VLT basalts from Apollo 17 and Luna 24 (including the ferrobasalts) exhibit a wide range of compositions, although all are slightly LREE-depleted (Fig. 16).

What is intriguing about the Luna 24 VLT basalts is that many contain positive Eu anomalies, with a few examples with slight negative anomalies (Fig. 16), whereas the Apollo 17 VLT basalts all exhibit only negative Eu anomalies (WENTWORTH et al., 1979). It is difficult to categorically state whether the variability of the Eu anomalies is a result of source compositions or sampling errors.

To some extent, sampling errors have been reduced for the Mare Crisium varieties by examining the fine-grained basalts, vitrophyres, and glasses (e.g., RYDER and MARVIN, 1978; GROVE and VANIMAN, 1978; MA et al., 1978b; NORMAN et al., 1978; COISH and TAYLOR, 1978). This was not the case for VLT basalts at the Apollo 17 site (e.g., WENTWORTH et al., 1979), so care is required in interpreting these samples.

The range in VLT compositions, especially from Luna 24, suggests a relationship through simple fractional crystallization. As noted above, these basalts define a trend of increasing Al_2O_3 contents with decreasing MG# (Fig. 3a). Although the most primitive samples from each suite have similar major-element compositions, Apollo 17 VLT basalts do not exhibit fractional crystallization trends in the direction of the Luna 24 VLT series (Fig. 17a–c) (GROVE and VANIMAN, 1977; RYDER and MARVIN, 1978). Both major and trace elements suggest similar, yet distinct, source regions for VLT compositions at these sites (Fig. 17a–d). Variation within the Luna 24 crystalline VLT suite is greater than expected for short-range unmixing and has been attributed to olivine fractional crystallization (e.g., TAYLOR et al., 1978), a conclusion apparent in Figs. 2 and 3. NORMAN et al. (1978) and COISH and TAYLOR (1978) suggested that the fine-grained crystalline Luna 24 VLT basalts were products of a more primitive parent, represented by the olivine vitrophyres, whereas RYDER and MARVIN (1978) and GROVE and VANIMAN (1978) suggested that the Luna 24 green glass is parental to the VLT, ferrobasaltic suite through ~50% fractional crystallization of olivine. TAYLOR et al. (1978) concluded that an Apollo 17 VLT glass composition is parental to the Luna 24 VLT/ferrobasalt suite, which would document the first link between the enigmatic glasses and crystalline mare basalts. It also has been suggested that an Apollo 15 green glass composition could be parental to the VLT basalts, again through olivine fractional crystallization (TAYLOR et al., 1977; VANIMAN and PAPIKE, 1977). However, GROVE and VANIMAN (1978) and MA et al. (1978b) have demonstrated that this is not feasible based upon experimental petrology and REE-profile comparisons.

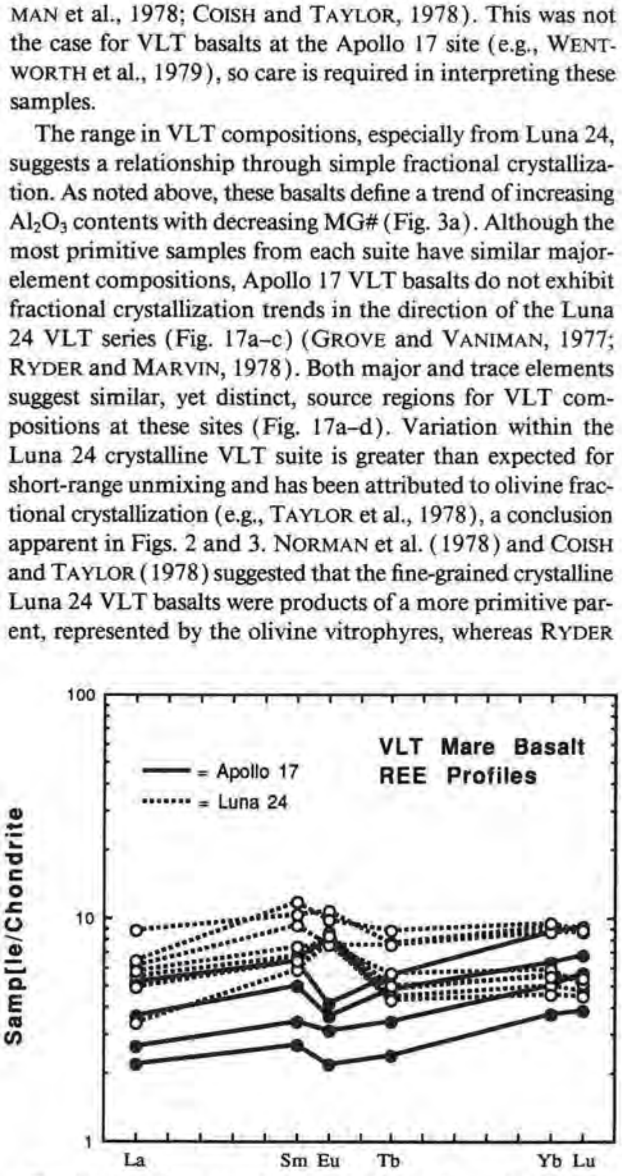


FIG. 16. Chondrite-normalized REE plots for the very-low-Ti mare basalts. Note the positive Eu anomalies present in the Luna 24 samples.

Source Modelling

As the Apollo 17 and Luna 24 VLT basalts cannot be related, it is apparent that two separate sources are required (e.g., RYDER and MARVIN, 1978). Source modelling of the Apollo 17 VLT basalts was undertaken by WENTWORTH et al. (1979), who demonstrated that the observed VLT liquid composition can be produced by 1–2% partial melting of a 90% olivine, 10% orthopyroxene cumulate, assuming this cumulate crystallized from a LMO melt with a flat REE pattern. WENTWORTH et al. (1979) further suggested that if this cumulate source formed from a magma with a fractionated REE pattern [e.g., $(La/Lu)_N \sim 2$], the VLT basalt REE pattern could be generated by ~4% partial melting of 70% olivine, 30% orthopyroxene source mineralogy.

Many Luna 24 VLT rocks exhibit positive Eu anomalies (Fig. 16). If this is an inherent feature rather than a sampling error, it alone demonstrates the difference between Apollo 17 and Luna 24 VLT basalts. This, coupled with the relatively high-alumina nature of the Luna 24 VLT basalts, led RYDER and MARVIN (1978) to suggest that the source may have contained plagioclase, even though it lay below the lunar crust (i.e., within the early olivine- and orthopyroxene-dominated LMO cumulates). A plagioclase-bearing source is compatible with Rb/Sr considerations (NYQUIST et al., 1977). MA et al. (1978b) concluded that the varying REE contents coupled with the nearly constant Sc and Eu abundances suggested that small amounts of clinopyroxene and plagioclase remained in the residue after partial melting. Such compositions would not be consistent with assimilation of plagioclase-rich crustal materials, as Eu abundances and Rb/Sr ratios would show considerable variation, not the constant values consistent with buffering by plagioclase left in the source.

NYQUIST et al. (1978) applied a cumulate remelting model to the Luna 24 VLT compositions, assuming the cumulate source crystallized from a slightly LREE-enriched magma. Their model requires low to moderate degrees of partial melting (3–10%) of a source dominated by orthopyroxene and olivine, but also containing minor clinopyroxene and plagioclase. NYQUIST et al. (1978) justify such a source composition by noting that (1) the LREE depletion for Luna 24 VLT basalts is consistent with the presence of clinopyroxene and orthopyroxene; (2) the higher HREE abundances of the Luna 24 relative to other mare basalts is consistent with the presence of orthopyroxene; and (3) the low Rb/Sr ratios, low initial $^{87}Sr/^{86}Sr$ ratios, and positive Eu anomalies are consistent with the presence of plagioclase. Although the rationale for pyroxene and plagioclase in the source appears to be valid, the high proportion of orthopyroxene is not necessary because the Luna 24 basalts do not contain higher HREE abundances relative to other mare basalts (Fig. 18a). Rather, the Luna 24 VLT basalts contain relatively low HREE and LREE abundances relative to the LREEs (Fig. 18a,b). Therefore, this model requires some modification, basically increasing the proportion of olivine relative to orthopyroxene (i.e., after the model of WENTWORTH et al., 1979).

Synthesis

Two groups of VLT basalts may be present at Mare Crisium: one is similar to the Apollo 17 VLT basalts and one is distinct. Olivine fractionation has been important in the evolution of the VLT mare basalts at least at the Luna 24 site. Although the details of VLT source composition are clearly arguable, it is obvious that the majority of Luna 24 VLT basalts require a source distinct from similar basalts at Apollo 17. The relatively high-alumina nature of the Luna 24 varieties, coupled with small positive Eu anomalies, demonstrates the need for plagioclase in the source region. Indeed, many of these Luna 24 basalts are saturated with both olivine and plagioclase, and some even plot in the plagioclase field of an olivine-quartz-plagioclase pseudoternary. Furthermore, the VLT nature requires that practically no oxide phase be present in the source. PAPIKE and VANIMAN (1978) and GROVE and VANIMAN (1978) suggested that VLT basalts were derived from early LMO cumulates dominated by olivine and orthopyroxene, but containing no oxide phase and also no plagioclase. Plagioclase can be conveniently incorporated into the VLT source by the convective-overturn model of HUGHES et al. (1988, 1989), but this would also bring in the late-stage, dense, ilmenite-rich cumulates. Clearly, the details of this model and its applicability to VLT basalts require further work. However, the VLT basalts have been invaluable in our understanding of the lunar interior and igneous processes.

SYNOPSIS

The mare basalt data base, compiled since the return of the first lunar samples, allows a classification to be formulated based primarily on TiO_2 content, with a secondary subdivision using Al_2O_3 content, and a tertiary subdivision using K abundance. We have suggested that the original three-fold classification of very-low-, low-, and high-Ti basalts be modified; the VLT class remains the same (<1 wt% TiO_2), but the low-Ti basalts encompass 1–6 wt% TiO_2 and the high-Ti basalts encompass 6–15 wt% TiO_2. The most important aspect of our classification is that each of the three groups defined on the basis of Ti-contents contains subgroups of low- (<11 wt%) and high-Al_2O_3 (>11 wt%) basalts, and each of these is further subdivided into low- (<2000 ppm) and high-K (>2000 ppm) groups. Therefore, on the basis of this new classification scheme there are potentially twelve groups of mare basalts (see Fig. 5), of which only six have been sampled to date: (1) high-Ti/low-Al/low-K (referred to as "high-Ti/low-K"); (2) high-Ti/low-Al/high-K (referred to as "high-Ti/high-K"); (3) low-Ti/low-Al/low-K (referred to as "low-Ti"); (4) low-Ti/high-Al/low-K (referred to as "high-alumina"); (5) low-Ti/high-Al/high-K (referred to as "VHK"); and (6) VLT/low-Al/low-K basalts (referred to as "VLT"). Such a classification is not perfect, as witnessed by the Luna 24 VLTs "encroaching" the high-alumina field, but it does allow a convenient evaluation of source regions and post-magma-generation processes for basalts of broadly similar compositions. This classification is also flexible inasmuch as future sample returns of mare basalt may contain previously unsampled varieties which can be conveniently "classified." Our newly proposed classification will not be rigorously tested until future missions to the Moon are undertaken.

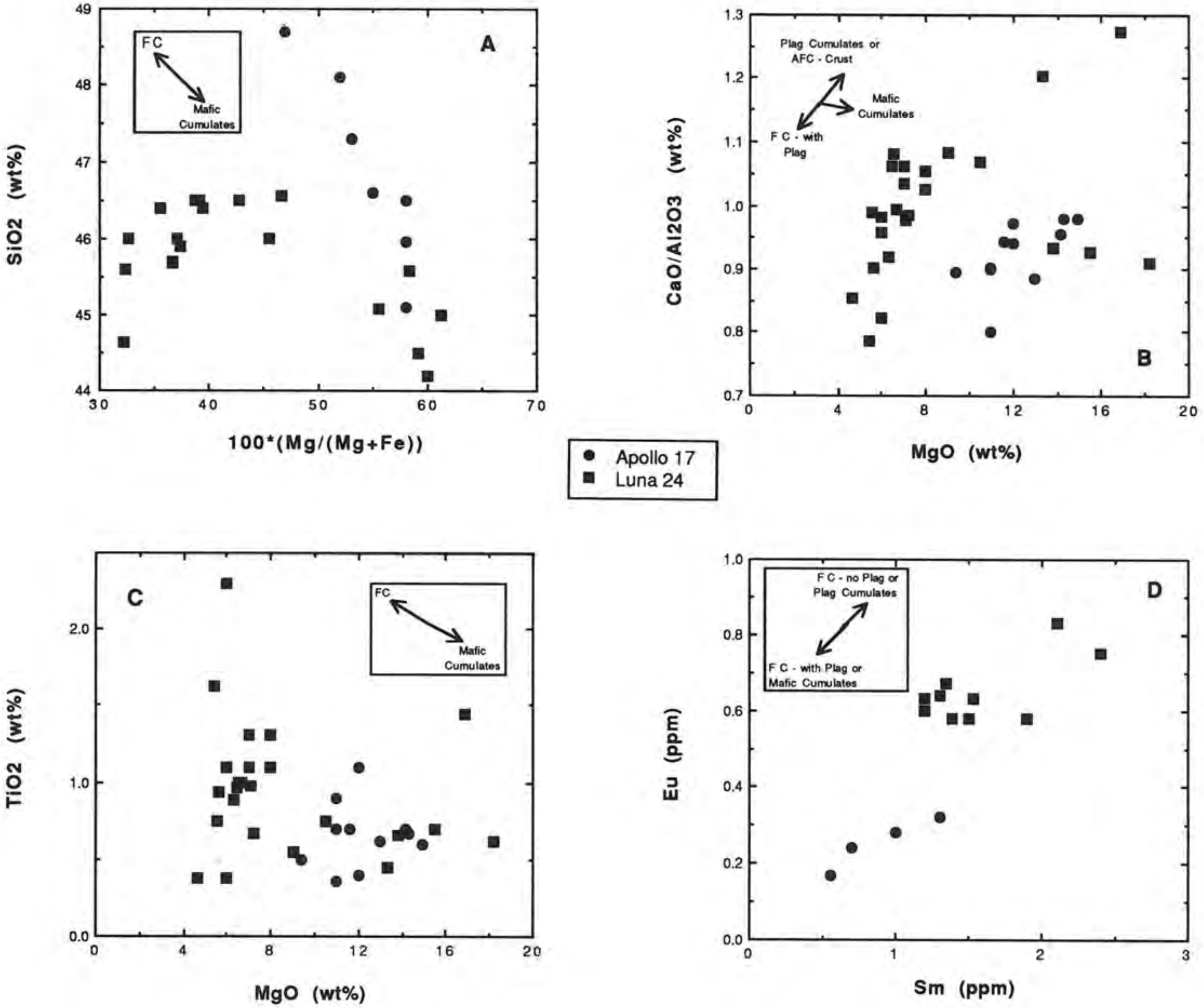
The petrogenetic modelling of basalts highlights the significance of source heterogeneity and fractional crystallization at a single site (e.g., Apollo 11, 12, 15, 17) for low- and high-Ti mare basalts. As part of this synopsis, we will briefly summarize the pertinent features discussed above using the primary subdivisions based upon Ti contents.

High-Ti Mare Basalts

The range in composition of the high-Ti basalts requires source heterogeneity to account for at least the range in K contents, but the location of these sources is debatable. If the LMO "layer-cake" hypothesis is correct, the high-Ti source should be composed of late-stage clinopyroxene and ilmenite cumulates, placing the source close to the crust-mantle boundary. If this is so, the MG#'s of these basalts cannot be generated. Such a problem may be overcome by convective overturn of the cumulate pile in order to mix in some early, high-Mg cumulates, but at what depth this source would be located is unknown. It must be above the pressures at which eclogite forms, as such a source would not yield the observed trace-element patterns. Therefore, we suggest that the general high-Ti source was generated by limited or localized overturn of the LMO cumulate pile. Also, there is some debate as to whether the Fe-Ti oxide minerals were exhausted or left as a residual phase after melting of the high-Ti source. The observed scatter within each group is attributed to either closed- or open-system fractional crystallization.

Low-Ti Mare Basalts

Low-Ti/low-Al/low-K basalts (from Apollo 12 and Apollo 15) also require heterogeneous source regions and fractional crystallization in order to account for the range of compositions within and between the different groups. There is some suggestion that the Apollo 12 olivine basalts may have experienced open-system fractional crystallization because of the range in initial Sr isotopic ratios. The source region must have been dominated by early LMO cumulates (olivine and pyroxene) with a lower abundance of Fe-Ti oxide phases relative to the high-Ti sources. The high-alumina basalts from Luna 16, and particularly from Apollo 14, have a wide range of composition. The Luna 16 data suggest that closed-system fractional crystallization generated the range of compositions, but to what extent sampling errors have affected these results is not known. Apollo 14 high-alumina basalt compositions are consistent with KREEP involvement in their petrogenesis, but whether this KREEP component was in the source


A
FC
Mafic Cumulates
SiO2 (wt%)
100*(Mg/(Mg+Fe))
49
48
47
46
45
44
30
40
50
60
70
B
Plag Cumulates or AFC - Crust
Mafic Cumulates
F C - with Plag
CaO/Al2O3 (wt%)
MgO (wt%)
1.3
1.2
1.1
1.0
0.9
0.8
0.7
0
4
8
12
16
20
● Apollo 17
■ Luna 24
C
FC
Mafic Cumulates
TiO2 (wt%)
MgO (wt%)
2.0
1.0
0.0
0
4
8
12
16
20
D
F C - no Plag or Plag Cumulates
F C - with Plag or Mafic Cumulates
Eu (ppm)
Sm (ppm)
1.0
0.8
0.6
0.4
0.2
0.0
0
1
2
3

(hybrid source model) or added after magma generation (AFC model) cannot be unequivocally stated. Source modelling of these basalts requires that plagioclase be involved, but the amount is limited to <10–15% in order to maintain the inherent negative Eu anomaly observed in high-alumina basalts. Inasmuch as plagioclase flotation during LMO crystallization was responsible for lunar crust formation and the negative Eu anomaly in the mantle, either the plagioclase separation was incomplete (e.g., SNYDER et al., 1992) or the source was situated at a shallow depth, being close to the crust. The VHK mare basalts formed due to a high-alumina magma assimilating lunar granite. These low-Ti variants were produced by granite periodically substituting for KREEP along the high-alumina AFC trend, thereby producing VHK compositions.

VLT Mare Basalts

The VLT mare basalts must be interpreted with extreme caution because of their small sizes, which preclude definitive petrogenetic interpretation for these basalts, and maybe all Soviet Luna samples, due to the problems of representative sampling. Apparent fractional crystallization trends are delineated by these basalts, but whether this is closed- or open-system is not known. Isotopic studies have been too few to accurately characterize such behavior. Source regions must have been almost devoid of Fe-Ti oxide minerals and dominated by early LMO cumulates (e.g., olivine and orthopyroxene, as per WENTWORTH et al., 1979), suggesting derivation from deep within the lunar mantle. However, the presence of positive Eu anomalies in many Luna 24 VLT basalts requires the presence of plagioclase in the source (many lie on the cotectic of the Ol-An-Qz pseudoternary), which requires either proximity to the crust or crustal assimilation (the latter is not immediately evident on the basis of the current data base). Getting components from depth and near the crust in the same source requires either convective overturn of the cumulate pile or incomplete plagioclase separation during crust formation. Plagioclase does not begin to fractionate until at least 60–70% of the LMO has crystallized, which would require the VLT source to be comprised of relatively late, rather than early, cumulates.

So where do we go from here? While the LMO hypothesis can explain many of the observed features of mare basalts, it is evident that the idealized layering of early cumulates at the base of the LMO and late cumulates just below the crust is too simple. Rather, we concur with SNYDER et al. (1992) that separation of the cumulus phases was incomplete (at least for the high-alumina and maybe Luna 24 VLT basalts) and some limited overturn of the LMO cumulate pile could have occurred. There is the possibility of incomplete plagioclase separation from the mafic mantle cumulates, which would explain the high-alumina and Luna 24 VLT basalt compositions; such a phenomenon would also alleviate the necessity of any overturn of the LMO on a grandiose scale. Although the global overturning hypothesis (e.g., HUGHES et al., 1988, 1989; RYDER, 1991b) is attractive, would this place the late-stage, Fe-Ti oxide-rich cumulates (i.e., the high-Ti basalt source) within the eclogite field? An eclogitic source would not produce the observed mare basalt compositions.

The modelling of source compositions has been critical in determining the nature of the lunar mantle and, therefore, lunar evolution. As such, the post-magma-generation evolution of each basaltic type must be fully understood in order to define a parental magma which is representative of the source from which it was derived (i.e., has not experienced any fractional crystallization or assimilation). In the past, source modelling has often been undertaken on compositions which have experienced post-magma-generation evolution, and results do not reflect the composition of the true source. As such, a re-evaluation of source compositions for the different basaltic types is essential in understanding the nature and evolution of the lunar interior.

We suggest that future research should focus upon the nature of the lunar mantle. Questions to be addressed should include (1) accurate identification of parental magma compositions of each group of mare basalts in order to determine precise source compositions; (2) the nature of source heterogeneity (e.g., metasomatism of a single source as proposed for Apollo 17, or totally different sources); (3) the problem of overturning the cumulate pile in order to yield a source containing early and late LMO fractionates (i.e., limited or Moon-wide overturn); and (4) incomplete separation of plagioclase from the mafic mantle cumulates during the formation of the lunar crust. Further pertinent questions regarding mare basalt petrogenesis involve the nature of post-magma-generation processes. For example, it is evident that fractional crystallization (whether open- or closed-system) played an important role in generating the compositional ranges within the various basaltic groups. Did this fractional crystallization occur at the surface within thick flows, or in subcrustal magma chambers? Models have suggested that Apollo 14 high-alumina and VHK basalts formed through assimilation and fractional crystallization (open-system), the high-alumina basalts evolving by assimilation of KREEP, which was occasionally replaced by granite to produce the VHK offshoots. Therefore, we suspect a relationship between granite and KREEP (e.g., NEAL and TAYLOR, 1989), but what is the exact nature of this? Furthermore, if sub-crustal magma chambers existed, then it is possible that replenishment-fractional-crystallization (RFC) processes may have occurred. Such RFC possibilities have not, as yet, been tested.

As is evident from the above discussion, the generation of the mare basalts has been attributed to a variety of igneous

FIG. 17. Major- and trace-element compositions of the VLT basalts. (a) = MG# (molar) vs. SiO_2 (wt%); (b) = MgO (wt%) vs. CaO/Al_2O_3 ratio; (c) = MgO (wt%) vs. TiO_2 (wt%); (d) = Sm (ppm) vs. Eu (ppm). In 17 (a)–(c) it is suggested that there may be two groups of VLT basalts at the Luna 24 site, one of which is similar to those from Apollo 17, but the two groups cannot be related, whereas (d) defines two groups of VLT basalts on the basis of site, and no Luna 24 basalts plot with those from Apollo 17. Whether this is real or due to the sparse trace-element data for these basalts is unclear. Hypothetical evolution paths are also shown (descriptions are the same as in Fig. 6). Basalt data are from HASKIN (1978); LAUL et al. (1978); MA et al. (1978b); RYDER and MARVIN (1978); TAYLOR et al. (1978); VANIMAN and PAPIKE (1977); WENTWORTH et al. (1979).

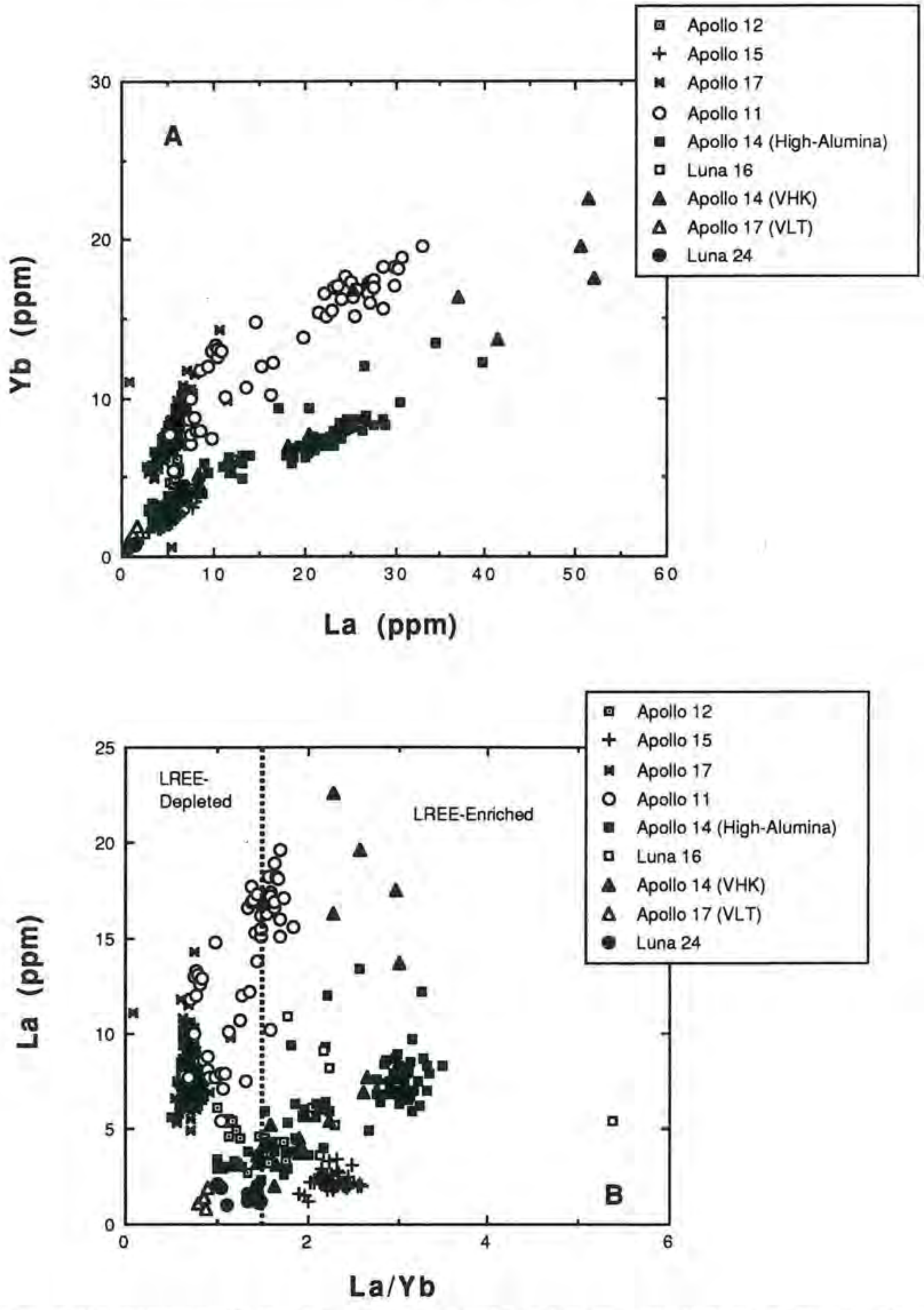


FIG. 18. (a) La (ppm) vs. Yb (ppm) demonstrating that VLT basalts from Luna 24 do not contain the highest HREE abundances relative to other mare varieties; (b) La/Yb ratio vs. Yb (ppm) for all mare basalts to highlight the LREE-enriched and depleted signatures of all mare basalts. Data sources are as in Fig. 2.

processes. With continued research, petrogenetic models can be expected to evolve or to be superseded. A few will stand the test of time, but each will require testing. Furthermore, some results have raised more questions than they have answered (e.g., Apollo 14: coupled, decoupled, or cyclical AFC, or hybrid sources?). Many small samples remain to be examined in the present collections; however, the continuing problems of representative sampling will have to be overcome.

A return to the Moon for more samples is essential, and this is even more important considering that none of the samples gathered and returned to date were taken from outcrop; they were all portions of the regolith, created by meteorite impact processes. Stratigraphic sampling of an *in situ* lava pile (such as at Hadley Rille) is essential for substantiating models presented above. An increase in the detailed trace-element and isotopic data base for volcanic glasses is required in order to fully evaluate any relationship with crystalline basalts. In the absence of a new sampling mission, existing data bases for the mare basalts must be expanded, scrutinized, and evaluated in our quest for knowledge of the origin and evolution of the Moon.

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