Characteristics of lunar mare deposits in Smythii and Marginis basins: Implications for magma transport mechanisms

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Abstract. An analysis of 34 lunar lava flows and ponds in the eastern limb Smythii and Marginis basins was undertaken to examine and model the first stages of secondary crustal formation and assess processes involved in magma transport and eruption. In order to isolate the characteristics of single eruptive episodes, we focused on discrete mare ponds adjacent to the major maria. Mean values for areas and volumes of deposits estimated to be good candidates for single eruptive phases are large, approximately 950-1000 km² and ~200 km³ respectively. These eruptive volumes are commensurate with the largest known terrestrial eruptions (flood basalts). The lack of geomorphological structures indicative of shallow magma reservoirs indicates that deep, probably subcrustal source regions are prevalent. With respect to crustal thickness relationships, the magnitude and frequency of eruptive events are observed to be greatest in areas of thinnest crust. Specifically, regions of major maria (Mare Smythii, Mare Marginis) occur in areas of thinnest crust, while isolated ponds occur where the crust is relatively thicker. This is consistent with the correlation observed globally between crustal thickness and the magnitude and frequency of eruptive events. Ages of volcanic flows and pyroclastic events range from Early Imbrian to Imbrian-Eratosthenian but are concentrated most heavily between 3.80-3.60 Ga, suggesting a period of peak volcanism beginning around 3.85 Ga and lasting approximately 200 Ma. However, the existence of dark-halo crater clusters in non-mare units within the region suggests the presence of cryptomaria, which would indicate an earlier onset of volcanism and a volume of mare material potentially greater than that currently exposed on the surface. Typical nearest-neighbor distances suggest deposits derive from reservoirs <100 km in diameter. The observations made here are consistent with a magma transport model in which plumes rising diapirically stall at a density boundary under the lunar crust and propagate dikes to the surface through overpressurization.

1. Introduction

Lunar volcanic deposits (maria) represent only a small fraction (17%) of the surface area of the Moon [Head, 1976]. In terms of lunar evolution, however, these mare deposits are a crucial link between our understanding of the initial stages of primary crustal formation from the lunar magma ocean, and subsequent thermal evolution that produced a partial secondary crust [Taylor, 1989]. The characteristics of these mare deposits, their size and thickness, age and composition, morphology and setting, associated features and distribution across the lunar surface, provide important information for deciphering the processes responsible for their generation and formation. For example, the distribution of deposits across the Moon is highly heterogeneous. As seen in Figure 1, the nearside of the Moon has a significantly higher density of mare deposits than the farside. The reason for this dichotomy is unclear but must depend in part on the underlying mechanisms driving magma through the crust to the surface. It has been suggested, for instance, that crustal thickness plays an important role in the efficacy of magma transport [e.g., Solomon, 1975; Head and Wilson, 1992; Robinson et al., 1992], due to the density contrast between mare basalt magma and the highlands crust. Thus the mare distribution dichotomy may be due to the nearside-farside crustal thickness asymmetry observed by the Apollo and Clementine missions [e.g., Zuber et al., 1994; Neumann et al., 1996].

In order to resolve such questions, it is vital to reconstruct the conditions for ascent and eruption of magma so that subsurface processes may be constrained and modeled. Classification and analysis of common mare deposit characteristics provides a diagnostic tool in this regard. In previous efforts we have compiled and analyzed a database of characteristics for isolated mare deposits and individual eruptive events occurring in basins on the lunar western limb and farside [Yingst and Head, 1994, 1997a]. Here we expand this study to include isolated mare deposits in the eastern limb basins of Smythii and Marginis. We compare the results from these basins to those of previous average volume and distribution estimates in order to build a statistical picture of a typical lunar eruptive phase. We then interpret these results in terms of source region geometry and crustal thickness relationships, in order to understand them within the larger framework of magma transport and eruption mechanisms.

2. Method

The first step in our approach is to model the morphology of a typical lunar volcanic phase, defined here as a single dike
emplacement event that may have ranged from a small, short duration eruption up to a high-flux, longer duration eruption lasting several years. Accordingly, it is necessary to isolate those characteristics which are common to lunar eruptive events. This task is hampered by the complex nature of the large maria. Although flow fronts are locally observed in some regions [Gifford and El-Baz, 1981; Schaber, 1973], the morphology of eruptive episodes is difficult to identify against the background of other deposits displaying a variety of compositions, ages and stratigraphic positions.

In light of these constraints, we have adopted an approach similar to the one used to categorize and analyze pond characteristics in other farside basins [Yingst and Head, 1997a], where analysis is limited to significant populations of discrete, isolated mare patches, or ponds [Beals and Tanner, 1975; Whitford-Stark, 1982], regions more likely to represent individual volcanic phases. Such groupings are most common on the lunar limbs and farside, as seen in Figure 1.

For the purpose of comparing possible models of magma transport, regions of study were limited to those basins whose age, diameter, depth, morphology and associated crustal thicknesses differed from regions previously examined (e.g., Orientale and South Pole-Aitken basins [Gaddis, 1981; Yingst and Head, 1997a], Australe basin [Hiesinger et al., 1996]). On the basis of these criteria, lava ponds in the Smythii and Marginis basins on the eastern limb were mapped and their areas determined. Ages, modes of occurrence, topography, and associated features were identified using Lunar Orbiter, Apollo and Clementine data. Evidence was sought in each discrete pond for multiple flows in order to isolate those characteristics common to individual flows or eruptive episodes. To this end, ponds showing a homogeneous albedo, color, and crater-frequency distribution, as well as a lack of characteristics indicative of multiple flows (overlapping flows, several potential source vents, etc.), were estimated to be individual eruptive phases. Thus, although it is not possible unambiguously to determine the number of flows represented by a mare pond, the deposits described above were considered best candidates for estimates of individual eruptive phases. Clementine multispectral data has been used in the past to test the efficacy of estimating individual eruptive phases in this manner [Yingst and Head, 1997b]. However, Clementine images of the Smythii/Marginis region of the eastern limb tend to have low phase angles. Because the photometric properties of Clementine images at low phase angles change very rapidly [Nozette et al., 1994] in a manner that is currently not precisely modeled [e.g. McEwen, 1996; Pieters et al., 1997], multispectral data for the Smythii/Marginis region cannot at this time be used as an accurate indicator of subtle variations in soil mineralogy. For this reason, we have chosen not to include the Clementine multispectral dataset in this work.

Pond thicknesses were calculated using a variety of methods described in previous studies [Yingst and Head, 1997a; Gillis et al., 1997]. These thicknesses were used to estimate the total volume of each deposit or flow. Finally, associated

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Figure 1. Map showing the location of mare deposits on the lunar surface. Maria are shown in black. Distribution is concentrated on the nearside, while the farside and limbs have very few deposits. Regions mentioned in the text are noted as follows: SPA, South Pole-Aitken; Or, Orientale; SM, Smythii/Marginis; and MR, Mendel-Rydberg. Map after Schultz and Spudis [1983].
thicknesses based on Clementine altimetry models were noted [Zuber et al., 1994; Neumann et al., 1996]. Results from these analyses are presented in Tables 1 and 2.

3. Results and Interpretation

3.1. Setting

The Smythii and Marginis basins, shown in Figures 1 and 2, lie on the eastern limb of the Moon. While they are both dated as pre-Nectarian, Smythii has a more well-defined ring structure, one that appears to overlap Marginis basin. It is thus considered younger [Wilhelms, 1987]. Marginis basin is approximately 580 km diameter [Wilhelms, 1987], although exact measurements are difficult because only partial segments of the ring structure remain. We assume for the purposes of this study that Marginis basin is circular, extrapolating from the semicircular ring structure noted by Wilhelms and El-Baz [1977]. This suggests a basin area of ~2.6 x 10^6 km^2. Smythii, by comparison, is larger (840 km diameter and 5.54 x 10^5 km^2 in area) and better preserved. It displays an extensive central mare deposit (Mare Smythii) and a subcircular central furrowed and pitted plains-type deposit (noted INfp) upon which many floor-fractured craters lie [Schultz, 1976]. Most mare ponds within central Smythii basin lie within these craters.

3.2. Areas and Volumes

Thirty-nine volcanic deposits (shown in Figure 2) were mapped in the Smythii and Marginis basins; their characteristics are described in Tables 1 and 2. Two deposits (Mare Smythii and Mare Marginis) have areal extents and estimated volumes at least 10 times greater than any other pond in the region. These regions display a complex morphology and have evidence of multiple flows [Wilhelms and El-Baz, 1977], some of which are suggested to be younger than Apollo 12 basalts (3.20 Ga [Spudis and Hood, 1992]). These two deposits are more similar to the major maria than small, discrete lava ponds. For the purposes of creating a statistical picture of individual eruptive phases, we thus focus on analysis of the small individual deposits. We then compare the results to characteristics of the larger Maria Smythii and Marginis.

Of the remaining 37 deposits in both basins, five show evidence that they represent multiple eruptive episodes (differing crater-frequency data [Wilhelms and El-Baz, 1977]). One multiphase deposit (Joliot; pond 1 in Marginis basin) shows morphological evidence of being comprised of more than one eruptive phase in that flow boundaries are evident. However, the precise boundaries of these flows have not yet been determined and the flows have not been dated. The remaining four of the five multiphase ponds (Camibens, Haldane, Kiess, and Tasso S in Smythii basin) have each been divided into two distinct deposits of differing ages and eruption histories [Wilhelms and El-Baz, 1977], yielding 41 separable volcanic deposits. The deposits noted as Late Imbrian/Eratosthenian dark material (Eld) in these four craters, as well as Smythii W, Kastner NE and Widmannstätten, are interpreted to be dark mantle deposits rather than effusive eruptions whose thicknesses can be estimated by the methods stated above and are therefore not considered as mare ponds. Although it is possible that these Eld deposits contain effusive as well as pyroclastically emplaced elements, because of the
### Table 2. Characteristics of Lava Ponds in Smythii Basin

<table>
<thead>
<tr>
<th>Pond</th>
<th>Name</th>
<th>Location</th>
<th>Mode of Occurrence</th>
<th>Crater Age</th>
<th>Rim Height Diameter, km</th>
<th>Area, km²</th>
<th>Thickness, km</th>
<th>Volume, km³</th>
<th>Method of Thickness Estimation</th>
<th>Associated Features</th>
<th>Age</th>
<th>Inferred Crustal Thickness, km</th>
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<td>9°N, 85°E</td>
<td>CF</td>
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<td>150</td>
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<td>Im₃</td>
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<td>pNc</td>
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<td>Im₃</td>
<td>Im₃</td>
<td>65</td>
</tr>
<tr>
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<td>pNc</td>
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<td>65</td>
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<td>1675</td>
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<td>FFC</td>
<td>Im₃</td>
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<td>FFC</td>
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<tr>
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<td>0.500</td>
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<td>Im₃</td>
<td>M</td>
<td>Im₃</td>
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</table>

*Mode designations are CF, crater floor; IC, inter-crater region.

Age designations are Ic₃, Im₃, younger than Orientale basin; Nc, Nectarian; pNc, pre-Nectarian.

Method designations are CD, crater excavation relationships; PBC, partially buried or filled craters; CD, crater diameter/depth, rim height, floor diameter, etc., estimates; PBT, partially buried topography other than craters (e.g., kipukas); S, stratigraphic onlap; O, other (see text for description of methods).

Associate features are designated by the following nomenclature: M, mare ridge; RF, rift or scarp; R, ridge; LR, linear rille; SR, sinuous rille; FFC, fractured floor crater designated type III by Schultz [1976].

Age designations are Eld, Late Imbrian/Eratosthenian dark material; Im₃, Late Imbrian; Im₂, Early Imbrian, possibly as old as Orientale basin [from Wilhelms and El-Baz, 1977].

Dark mantle deposit.

Multiple phase deposit.
Figure 2a. Sketch map of Smythii/Marginis basin region showing the location of mare deposits. Ponds studied are indicated and correlate to numbers in Tables 1 and 2. Older Late Imbrian mare material (Im₁) is indicated in black. Younger Late Imbrian mare material (Im₂) is shown in grey, while striped areas denote Eratosthenian/Imbrian dark mantle material. Floor-fractured craters are indicated by a dot pattern. Ages noted for mare deposits are based upon Hiesinger et al. [1997] and Wilhelms and El-Baz [1977]. Regions within Mare Marginis dated through crater size-frequency data by Hiesinger et al. [1997] that are discussed in the text are noted by Roman numerals.
Figure 2b. Airbrush map of the Smythii/Marginis basins superposed onto the Clementine altimetry data of the region (based on data from Zuber et al. [1994]), showing the topography and setting of the basin.
mantling nature of pyroclastic deposits, it is difficult to distinguish between material emplaced by effusive flow and material emplaced by pyroclastic means. Consequently, if sufficient pyroclastics exist in a region to create any ambiguity we do not treat the deposit further as an effusive eruptive episode. The nature of these dark mantle deposits will be discussed in more detail below. Thus, on the basis of their characteristics (Tables 1 and 2) 34 out of 41 separable volcanic deposits and flows in Smythii and Marginis basins are classified here as mare ponds and of these, 33 ponds are interpreted to represent individual effusive eruptive phases.

Areas for the 34 mare ponds in the Smythii/Marginis region are shown as an area-frequency distribution in Figure 3. Estimates of individual eruptive phases are shown in Figure 3b. Joliot, which is assumed to be a multiphase pond whose individual flows are not currently measured individually, is excluded from this figure.

Pond areas range from 170 km$^2$ to 6575 km$^2$ (mean value 1120 km$^2$) but are concentrated in the lower portion of this range. For those ponds judged to be single eruptive episodes, areas also range from 170 km$^2$ to 6575 km$^2$, with a mean value of 965 km$^2$. The total area of all mare ponds in the Smythii/Marginis region is approximately 38,015 km$^2$, representing less than 5% of the total area of these basins. Coverage by lava ponds is higher within Marginis basin. About 21,950 km$^2$, or 8% of the Marginis basin area, is covered with lava ponds, compared to 16,065 km$^2$, or 3% of the area of Smythii basin. Mare coverage for this region (ponds and major maria) is about 102,845 km$^2$, or 13% of the total area in the two basins. For Marginis this equates to approximately 56,895 km$^2$, or 22% of the basin, while for Smythii total mare coverage is 45,950 km$^2$, which is 8% of the basin area. Thus, effusive mare deposits in these two basins make up approximately 13% of the total area of the basins, with ponds representing about 5% and the maria about 8% (Figure 2a).

Ponds in the Marginis region tend to be concentrated in the southern half of the basin, the only exceptions being the large crater floor ponds Joliot and Hubble (ponds 1 and 2). Pond density for Marginis basin is 1 pond per 20,000 km$^2$, while for the south portion of the basin it is 1 per 12,000 km$^2$. For Smythii, deposits that occur within the central region are relatively evenly spaced in the confines of the central ring, while those ponds occurring outside this region are concentrated in the northwest portion of the basin. Thus, while the pond density for the basin as a whole is only one deposit per 171,000 km$^2$, the deposit density within the central basin ring (area ~120,000 km$^2$) is one per 6,600 km$^2$. In general terms, the mean pond density for the entire Smythii/Marginis region is about one pond per 121,000 km$^2$, but the concentration of mare deposits appears to be related to the state of preservation of the basin. In the better preserved Smythii basin, deposits are highly concentrated in the center of the basin, while in Marginis basin the distribution of deposits is much more diffuse. Deposits in Marginis are not contained by the highly degraded topography of the central depression or basin rings.

Volumes of ponds and individual flows were estimated using the various methods described in Gaddis [1981] and Yingst and Head [1997a]. Specifically, in those cases where such craters are available, pond thicknesses were estimated by calculating the depth of relatively young, optically mature post-mare craters based on crater diameters [Pike, 1977, 1980]. The depth of excavation was then determined from this value [Stöffler et al., 1975] and the pond thickness was estimated, where dark ejecta indicates a minimum depth and bright, presumably highland ejecta indicates a maximum depth for the mare deposit. Crater depth/diameter relationships derived by Pike [1977] were also used to estimate the geometry and depth of fill for flooded or partially embayed craters, where such existed, so that the thickness of the embaying unit could be determined. For ponds lying on relatively fresh crater floors, the measured diameter of the floor was compared to the calculated floor diameter of the original crater based on depth/diameter measurements [Pike, 1977]. The difference between these two values was then used to derive a thickness estimate [Whitford-Stark, 1979]. Other indicators of deposit thickness used were shadow measurements on flow fronts where available, and partially buried topography other than craters, where the elevation of the pre-existing topography was known from topographic maps or other sources. For a more comprehensive treatment of these pond thickness estimate methods the reader is directed to Gaddis [1981] and

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Yingst and Head [1997a]. The techniques used for each lava pond are indicated in Tables 1 and 2, and the compiled data are shown in a frequency distribution plot in Figure 4.

The total calculated volume for mare ponds in the Smythii/Marginis region is 6755 km$^3$, where 3445 km$^3$ is in Smythii basin, and 3310 km$^3$ is in Marginis basin. As shown in Figure 4a, the range for pond volumes in both basins is 15 km$^3$ to 1045 km$^3$, with an average value of 200 km$^3$. The largest volumes occur within the large craters (e.g., Neper, Kiess, and Helmert-Kao). In the case of those ponds or flows estimated to be individual eruptive events, volumes also range from 15 km$^3$ to 1045 km$^3$, as shown in Figure 4b. The total volume of ponds in Smythii basin is estimated to be -3450 km$^3$, or about 19% of the total volume of mare material in the basin. Similarly, the total volume of ponds in Marginis basin is estimated as -3305 km$^3$, or 12% of the total volume of basin mare material. The mean value for volumes of individual eruptive phases for the region as a whole is 195 km$^3$ (190 km$^3$ average for Smythii basin and 270 km$^3$ for Marginis basin). Such average values bracket the mean volume for individual eruptive phases in the Orientale/Mendel-Rydberg basins (about 240 km$^3$ [Yingst and Head, 1997a]).

3.3. Modes of Occurrence

We noted two modes of occurrence for mare ponds: (1) within crater floors, and (2) in inter-crater highland regions. These are shown in terms of area- and volume-frequency distribution plots in Figures 5 and 6. For the Smythii/Marginis region as a whole, the majority of ponds and flows occur on crater floors (24, representing 71% of all occurrences). In the case of single eruptive episodes, most of the discernible flows (23 out of 34) lie within craters. In addition, crater-floor occurrences tend to represent the highest volume mare ponds. Thus a regional preference for crater floors is observed both for mare pond occurrence and total volume extruded. The majority of these crater floor ponds lie in the relatively young (Late Imbrian-aged) craters that dominate the floor of Mare Smythii, suggesting that these ponds may represent a later stage of volcanism, an observation that we will return to later. Of the crater floor ponds that occur in older craters, eight (33% of crater floor occurrences) lie in pre-Nectarian craters outside the central basin, and only 4 ponds (17% of crater floor occurrences) occur in Nectarian-aged craters. In contrast, 12 ponds, or almost 50% of crater occurrences, lie within Imbrian-aged craters. Two such ponds (ponds 10 and 11 in Smythii basin; Figure 2a, Table 2) are shown in Figure 7. The craters in which these ponds occur are characterized by updomed floors, moats, and a system of subconcentric fractures, previously classified by Schultz [1976] as type III floor-fractured craters. It has been suggested that this type of uplifted and fractured crater morphology is due to endogenic modification due to viscous relaxation of crater topography over time [Danes, 1965; Baldwin, 1968; Hall et al., 1981] or through surface failure in response to igneous intrusion [Schultz, 1976; Brennan, 1975; Wichman and Schultz, 1995]. Eruption of magma has resulted in partial burial of several of the floor fractures in Cam6ens. Similarly, the pond in Doyle appears to have partially buried sections of the fracture system in the southwestern crater floor. Mare emplacement subsequent to fracture formation is thus suggested for both ponds.

Two separate volcanic episodes are discernible in Cam6ens, distinguished by differences in albedo, texture, and crater density. The northern deposit has fewer craters than the flow in
the central portion of the crater, a fact that led Wilhelms and El-Baz [1977] to interpret this deposit as younger. In addition, the northern deposit displays a very low albedo and a more diffuse boundary than the central deposit, such that while the central deposit represents an effusive volcanic eruption, the northern deposit is denoted as a dark mantle deposit [Wilhelms and El-Baz, 1977]. The northern deposit extends beyond the crater rim into the surrounding highlands, forming fine-textured lobes of a more rugged relief than the central deposit.

The diffuse dark material associated with the northern deposit appears to disregard boundaries formed by the preexisting topography. Thus, the above morphology suggests a pyroclastic emplacement mechanism for this deposit rather than an effusive one [Lucchitta, 1972; McGetchin and Head, 1973; Pieters et al., 1973, 1974; Wilhelms, 1987]. A 15 km long linear rille extends from the northwestern edge of the pyroclastic material in Camões crater to the southern rim of
Doyle, cutting through the younger Camões dark deposit. In
contrast to this pair of deposits, the pond in Doyle crater is
smaller in both area and volume, has a higher albedo and
shows a rougher texture than the central flow in Camões. The
mare material approximately follows the concentric contours
of the crater rim, but is confined to the crater's western
portion. No other features aside from crater floor fractures
appear to be associated with this pond.

On the basis of the preceding observations, the history of
this region may be reconstructed. Some time during the
Imbrian period impacts created the initial craters Doyle and
Camões. Fracturing of the crater floors followed. Mare
material was then emplaced, embaying some fractures in both
craters and filling the floor of Camões. A later episode of
pyroclastic activity occurred in the northern moat region of
Camões, mantling a portion of the crater rim and the
surrounding highlands. Either concurrent with, or after this
last volcanic episode, a linear rille formed from the
northwestern rim region of Camões to the southern rim of
Doyle crater.

The remainder of ponds in Smythii/Marginis (10 deposits,
or 29% of all occurrences) lie in inter-crater regions,
apparently unrelated to any other formation or feature. An
example of such a pond is located north of Al-Biruni crater
(pond 3 in Marginis Basin; Figure 2a), shown in Figure 8. In
the case of this pond, younger Late Imbrian mare material has
flooded a region of lower-elevation furrowed highlands north
of Al-Biruni crater [Wilhelms and El-Baz, 1977], embaying
several small craters and burying the hummocky pre-existing
topography. There are no vents or other structures visible that
would suggest the source of the mare material. Instead, effusive
eruption of mare material appears to have buried any
responsible source vents. Smaller craters on the surface of the
deposit are evidence of younger (post-Late Imbrian) impacts.
No other outstanding features are present in this mare deposit.
Thus the morphology and placement of this pond in a region
of low elevation make this deposit typical of ponds occurring
in inter-crater areas in Smythii and Marginis basins.

On the basis of these observations, the geologic history of
this pond may be determined. Formation of the many large
surrounding craters occurred first, including the formation of
the Nectarian-aged Al-Biruni crater to the south. Formation of
the furrowed inter-crater plains occurred during or immediately
after this period, both in this region and in central Smythii
basin. This was followed by emplacement of the mare pond by
effusive volcanism. Finally, subsequent to emplacement of
mare material, continued impact activity formed the small
younger craters in the mare pond and the surrounding
highlands.

For the Marginis basin, the majority of ponds (seven
ponds, or 58% of occurrences) lie in inter-crater regions, with
areas ranging from 175 km² to 3595 km², and volumes ranging
from 20 km³ to 715 km³. The remainder occur in crater floors
(five ponds, or 42% of occurrences), displaying areas ranging
from 290 km² to 6155 km², and volumes in the range of 60
km³ to 710 km³. In contrast, ponds within Smythii basin tend
to occur most frequently in crater floors (19 ponds,
representing 86% of occurrences), with areas ranging from 170
km² to 6575 km². Volumes for crater floor occurrences range
from 15 km³ to 1045 km³. Three ponds (14% of occurrences)
lie in inter-crater regions, showing areas ranging from 255
km² to 1675 km², and volumes in the range of 25 km³ to 250
km$^3$. All multiphase ponds occur within crater floors. The pond associated with crater Haldane (pond 12 in Table 2) breaches the crater rim and displays material both on the crater floor and in the inter-crater region to the north. However, because topographic evidence suggests that the source vent for this pond lies within the interior of crater Haldane, this pond is classified as a crater floor occurrence in Figures 5 and 6, and in Table 2.

3.4. Associated Features

Linear rilles, mare ridges, dark-halo craters, and floor-fractured craters are among the variety of features found associated with Smythii/Marginis region lava ponds, as noted in Tables 1 and 2. Examples of these features are labeled in Figure 9. For example, linear rilles are seen in Neper crater pond. Mare ridges are seen in Mare Smythii and Mare Marginis, as well as in the crater Joliot. They tend to be associated with large, contiguous maria. Clusters of dark-halo impact craters lie within the non-mare units north and east of Mare Marginis, and circumferential to Mare Smythii [Schultz and Spudis, 1979, 1983]. As was noted before, several ponds are also associated with floor-fractured craters (e.g., Figure 7 [Schultz, 1976]).

Linear rilles (simple graben) occurred in four ponds (12%); all of these occurrences were in craters (e.g., Camões, pond 11 in Figures 2a and 7). Volumes for these ponds range from 40 km$^3$ to 780 km$^3$, with an average volume of 300 km$^3$. The nonarcuate nature of these linear rilles, along with their association with mare deposits, favors a volcanic mode of origin, such as the near-surface emplacement of a dike [Pollard et al., 1983; Head and Wilson, 1993].

Mare ridges were observed in the crater Joliot, as well as in Mare Smythii and Mare Marginis, the largest volume deposits. Dynamic models for the formation of these ridges [e.g., Melosh, 1978; Solomon and Head, 1980; Pullan and Lambeck, 1981; Golombek, 1985] tend to attribute ridge origin to the stresses placed on the lithosphere from large volumes of mare material. Such a conclusion is consistent with the observed association of mare ridges in Smythii and Marginis basins with the largest and thickest mare deposits.

Dark-halo impact craters have been mapped by Schultz and Spudis [1979, 1983] in highland regions surrounding Mare Smythii and Mare Marginis. Specifically, these craters appear to be concentrated within central Smythii basin and in the Al-Khwarizmi/King and Lomonosov-Fleming basins east and north of Smythii respectively. Such craters have been used as indicators of hidden volcanic mare material, termed cryptomare [Head and Wilson, 1992], that has been buried by basin or crater ejecta deposits [Schultz and Spudis, 1979, 1983; Hawke and Bell, 1981; Bell and Hawke, 1984; Antonenko et al., 1995]. The presence of clusters of dark-halo craters suggests that there was early volcanic activity in this region, the nature of which is not fully represented by the surface mare deposits. In addition, elevated concentrations of Mg, Fe and Ti, as well as low Al concentration in some portions of Al-Khwarizmi/King [Clark and Hawke, 1991] suggest the presence of a partially obscured basaltic component [Schonfeld and Bielefeld, 1978; Hawke et al., 1985]. Thus volumes indicated by mare ponds should be considered a minimum value for the total mare volume in this region.

There are several floor-fractured craters in the Smythii/Marginis region [Schultz, 1976; Wolfe and El-Baz, 1976; Wichman and Schultz, 1995]. Many of these craters have lava ponds associated with them. Eight ponds (21% of occurrences) occur in floor-fractured craters (e.g., Haldane [Wolfe and El-Baz, 1976], Doyle and Camões, Figure 7). The fractures associated with these craters are concentric to the crater rim, implying tectonic modification of the original crater such as uplift of the local topography by some mechanism [e.g., Schultz, 1976], rather than a strictly endogenic origin such as dike emplacement [e.g., Pollard et al., 1983; Head and Wilson, 1993]. The nature of floor-fractured crater formation will be discussed in more detail later.

3.5. Stratigraphy

Initially, the volcanic deposits of the Smythii/Marginis region were divided by Wilhelms and El-Baz [1977] into older
E1-Baz [1977] have identified these latter deposits as dark volcanic deposits. The very low albedo, fine texture and rugged relief of these deposits, along with the diffuse boundaries displayed by many of them (e.g., Camøens in Smythii basin, shown in Figure 7), are all consistent with a pyroclastic origin [Lucchiata, 1972; McGetchin and Head, 1973; Pieters et al., 1973, 1974; Wilhelms, 1987]. We thus interpret the seven deposits and flows noted as EId by Wilhelms and E1-Baz [1977] to be pyroclastic deposits of upper Late Imbrian or possibly Eratosthenian age. On the basis of albedo characteristics and the density of small superposed craters, the deposits have been further divided by Wilhelms and E1-Baz [1977] into 41 separate and datable lava ponds, dark mantle deposits or flows. These 41 volcanic deposits are shown in an age-frequency distribution plot in Figure 10. Hiesinger et al. [1997] have dated several ponds and mare regions in Marginis basin using crater size-frequency distribution measurements. These ages are displayed in Figure 10 as well. For those ponds whose ages do not agree between the two studies (three ponds in Smythii basin, dated as EId by Wilhelms and E1-Baz [1977]), we thus interpret the seven ponds by these authors as EId due to their relative pyroclastic origin rather than youth. Thus for these deposits albedo would be suspect as an indicator of age.

Crater age-frequency data accumulated recently [Hiesinger et al., 1997] suggest that a great deal of volcanism in Marginis basin occurred in a relatively limited timespan, ranging from 3.80-3.60 Ga. Based on the available evidence, it is reasonable to assume that the stratigraphic ages indicated by Hiesinger et al. [1997] may be typical of the entire period of volcanic activity in Marginis. In addition, because those ponds in Smythii basin dated as EId on the basis of albedo differences [Wilhelms and E1-Baz, 1977] are likely dark due to composition rather than age, it is possible that these deposits may have been formed exclusively in the upper Late Imbian. Thus the weight of evidence suggests that, for Marginis basin specifically, and by extrapolation the region in general, the majority of volcanic activity was in the latter part of the Late Imbian period.

Out of 41 individually mapped volcanic deposits and mare flows within the Smythii and Marginis basins, most (31 deposits, or 76%) are dated as Im1 (younger Late Imbian) [Wilhelms and E1-Baz, 1977; Hiesinger et al., 1997]. Of the remaining 10 deposits, two (5% of occurrences) are Im1 (older Late Imbian), seven (17%) are Late Imbian/Eratosthenian dark mantle deposits and one (in Hubble crater, pond 2 in Marginis basin, Figure 2a) is dated as 3.80 Ga, contemporaneous with the Early/Late Imbian division. The oldest ponds (Im1) are confined to the eastern portion of Marginis basin, while the dark mantle deposits (EId) rim the central interior of Smythii basin, occurring within, or mantling the rims of, several of the prevalent Imbian-aged craters there. For the Marginis basin area, two of the 12 deposits are Im1, (17%) and nine (75%) are Im2, while the pond occurring in Hubble crater (pond 2 in Marginis basin, Figure 2a) straddles the Early/Late Imbian periods. Likewise, all but one of the deposits in Smythii basin are dated either Im1 (21, accounting for 76% of deposits) or Late Imbian/Eratosthenian (seven deposits, or 24% of occurrences), only one (Erro NW, pond 2 in Figure 2a, Table 2) is dated Im1. This distribution of observed surface deposits implies that volcanism was active over a similar time period for both basins, with volumetrically small activity in the early part of the Late Imbian and more volumetrically significant activity during the latter part of the Late Imbian. Such activity is consistent with the volcanic flux measured for the Moon as a whole. Volcanic flux estimates based on lunar eruption rates [Hartmann et al., 1981] suggest that >90% of the volume of known volcanic deposits were emplaced in the Late Imbian period (3.80-3.20 Ga), peaking at ~200 Ma into the Late Imbian period (around 3.60 Ga [Head and Wilson, 1992]). While <5% was emplaced in the Eratosthenian period [Head and Wilson, 1997]. The distribution of volcanism during peak activity was widespread, involving most of the large nearside basins, including nearby Crisium [Wilhelms, 1987]. Thus, based upon the ages of the surface mare deposits, we conclude that the most volcanically active period in Smythii/Marginis basins (the Late Imbian) coincides with the most widespread and volcanically active time in lunar history.

![Figure 10. Distribution of mare pond ages within the Smythii/Marginis basin area. The inset stratigraphic column represents stratigraphic ages assigned by Hiesinger et al. [1997] on the basis of crater size-frequency distributions for some Marginis ponds and portions of Mare Marginis indicated by Roman numerals in Figure 2a. Note that the cluster of ages around ~3.80-3.60 Ga represents regions widely separated in space and that in the case of the pond in Ibn Yunus and areas of Mare Marginis directly south of this crater, neighboring regions display very disparate ages.](image-url)
3.6. Distribution

Nearest-neighbor distances were measured for the 34 mare ponds in Smythii/Marginis by taking the average value for the measured center-to-center distance of the five closest deposits to each pond, after Yingst and Head [1997a]. A frequency distribution plot for the average nearest-neighbor separation distances in the Smythii/Marginis region is shown in Figure 11. Center-to-center separation distances, shown in Figure 11a, range from 50 to 250 km and are typically in the range of 60-150 km, while edge-to-edge separation distances, shown in Figure 11b, range from 15 to 160 km, with a mean value of 65 km. These ranges are similar to those observed for ponds in the South Pole-Aitken and Orientale basins on the western limb and farside [Yingst and Head, 1997a]. Ponds within Smythii basin may be divided into two groups; those within the confines of the central ring (14 deposits) and those outside the central basin ring (eight ponds), so that comparison of distribution with respect to varying topography and crustal thickness may be analyzed. Mean values for center-to-center distances, shown in Figure 11c, are estimated to be 85 km within the inner basin ring and 160 km between the inner and outer rings. Mean values for edge-to-edge distances, shown in Figure 11d, are calculated to be 40 km in the basin center and 110 km outside the basin center. The fact that ponds are not only more frequent but are also more closely packed within the basin center suggests that pond distribution and spacing is dependent to some extent upon topography and the associated crustal thickness.

4. Comparison and Global Context

The significant number of lava ponds analyzed in this and other studies [e.g., Gaddis, 1981; Yingst and Head, 1997a; Hiesinger et al., 1996] permit comparisons to be made between the characteristics of ponds in disparate regions, allowing improved modeling of lunar eruptive commonalities in a global sense. Values for these characteristics are

![Figure 11](image-url). Frequency distribution plot of average nearest-neighbor distances measured for ponds in the Smythii/Marginis basin region. Center-to-center distances (Figure 11a) are larger than edge-to-edge distances (Figure 11b) because of the irregular shape of many of the ponds examined. Also shown are frequency distribution plots of center-to-center (Figure 11c) and edge-to-edge (Figure 11d) distances for ponds within Smythii basin alone, demonstrating the difference between deposits in the interior of the basin and those in the exterior rings.
summarized in Table 3. As has been previously stated, we interpret the 33 flows and ponds noted as individual phases in Tables 1 and 2 to be our current best estimate of individual eruptive events. It is this subset of deposits with which we will now be concerned.

4.1. Areas and Volumes

As stated previously, the mean areal extent of those lava ponds in Smythii/Marginis basins which are estimated to be the best candidates for individual eruptive episodes is 965 km², while the average volume is 195 km³. This compares to approximate mean areas of 1115 and 2080 km² and mean volumes of approximately 240 and 860 km³ for Orientale and South Pole-Aitken basins, respectively. A volume-frequency distribution plot for these deposits is shown in Figure 12. From these data, it is clear that South Pole-Aitken is characterized by a wider range of values, with typically higher volumes for individual ponds, while both Orientale and Smythii/Marginis have a narrower range of values, with occurrences peaking at lower volumes (less than 150 km³).

In all basins, however, mare ponds display volumes that are high by terrestrial standards. For example, typical flows for a single eruption from a shallow source region such as Hawaii average less than 1 km³ per eruption [Peterson and Moore, 1987]. On the other hand, the Laki eruption in Iceland, one of the largest historical terrestrial eruptions, was measured at ~12 km³ [Jónsson, 1983], which is comparable to the lowest volumes observed in this study. A more striking terrestrial comparison can be made using flood basalt provinces, volcanic regions believed to be associated with deep-seated source regions [Campbell and Griffiths, 1990]. For example, the Roza Member of the Columbia River Basalt province has an estimated volume of 1200 km³ [Tolan et al., 1989]. Such large volumes suggest by comparison that deep, rather than shallow source regions are indicated in the basins studied.

4.2. Morphology and Associated Features

The morphology and structures associated with mare deposits are indicators of the conditions that existed during magma extrusion. The typical morphology for lava ponds in this region is similar to that observed in South Pole-Aitken and Orientale basins. Ponds tend to be relatively smooth, with no domes, calderas, or other similar structures evident. Features are limited to linear rilles, dark-halo craters, and tectonic structures associated with floor-fractured craters. We will examine each of these features in detail below.

4.2.1. Linear rilles. Linear rilles were found to be associated with four deposits in the Smythii/Marginis region (12% of pond occurrences). The mean volume for these ponds is 300 km³, which is higher than the mean value for deposit volumes in this region. The number of ponds associated with linear rilles in Smythii/Marginis is higher than that for lava pond occurrences in South Pole-Aitken basin (two occurrences, or 4% [Yingst and Head, 1997a]), an observation attributable to the poor resolution and viewing angle of the images available for the previous study. In contrast, linear rilles occurred in ~35% of deposits examined in Orientale basin [Gaddis, 1981; Yingst and Head, 1997a], generally being found in deposits with the highest volumes. Linear and arcuate rilles have been interpreted to be linked to impact basin structure [Mason et al., 1976] and the emplacement of mare deposits in the basins via flexural deformation related to the
mare deposit load [Solomon and Head, 1980]. However, some linear rilles have been interpreted on the basis of their nonarcuate shape and their association with volcanic deposits to be the surface manifestation of a dike injected to near-surface depths [Head and Wilson, 1993]. None of the five linear rilles noted in this study follow the contours of an impact structure related to a mare deposit that would suggest a flexural origin [e.g., Solomon and Head, 1980]. For these linear rilles, then, the current best interpretation appears to be one in which graben form in response to local stresses produced by the near-surface emplacement of a dike propagated from depth.

4.2.2. Dark-halo craters. Dark-halo impact craters have been previously mapped in various non-mare units in the northeast portion of Marginis basin and in southern Smythii basin [Schultz and Spudis, 1979]. In general, these craters lie within the central portion of Smythii basin, as well as within the highly degraded pre-Nectarian [Wilhelms, 1987] basins Lomonosov-Fleming and Al-Khwarizmi/King north and east of Smythii. The presence of these dark-halo crater clusters is indicative of the existence of cryptomaria in this region [Schultz and Spudis, 1979; Hawke and Bell, 1981; Bell and Hawke, 1984], suggesting that the onset of mare volcanism might have been earlier than is currently indicated by the inferred ages of the known surface mare deposits [Schultz and Spudis, 1979, 1983; Hawke et al., 1985]. If cryptomaria is present in the Smythii/Marginis region as suggested, the total mare volume is greater than that suggested by the surface deposits. We may thus use the size and distribution of dark-halo craters both to identify cryptomare and to estimate the volumetric significance of cryptomare deposits [Schultz and Spudis, 1979, 1983; Hawke and Bell, 1981; Bell and Hawke, 1984; Antonenko et al., 1995]. For example, if we assume that the central portion of Smythii basin was originally filled with cryptomare material to a depth of -500 m (a depth similar to that of the present Mare Smythii), this yields a cryptomare volume of approximately 70,000 km$^3$, significantly more than either Mare Smythii or Mare Marginis. In fact, the total volume of all the mare deposits in Smythii basin analyzed in this study (18,390 km$^3$) represents only 27% of this potential cryptomare volume estimate.

As another example of the potential volumetric significance of cryptomare material, let us examine the pre-Nectarian basins Lomonosov-Fleming and Al-Khwarizmi/King. We note the areal extent of cryptomare in Lomonosov-Fleming to be roughly 65,000 km$^2$, and in Al-Khwarizmi/King to be about 80,000 km$^2$ based upon estimates by Schultz and Spudis [1983]. If we then assume that the estimated area of cryptomare for these older basins represents a fill of -500 m depth, we calculate a cryptomare volume of about 32,300 km$^3$ in Lomonosov-Fleming basin and 40,000 km$^3$ in Al-Khwarizmi/King basin. These values bracket the cryptomare volume estimated for Smythii basin but are more than twice the total volume of mare material currently visible on the surface of Smythii. These first-order calculations demonstrate the possibility that cryptomare material may be a significant component in the eastern limb Smythii and Marginis basins.

What could have obscured these potential cryptomare deposits? One possibility is that cryptomare material was obscured through local mixing by subsequent impacts. Impactors that form craters deep enough to excavate the underlying highlands may have contributed to mare soil contamination through deposition of ejecta containing highland material. This process has undoubtedly occurred. Because the areal extent of proposed cryptomare regions associated with southern Smythii, Lomonosov-Fleming and Al-Khwarizmi/King are commensurate with regions such as Maria Smythii and Orientale [Schultz and Spudis, 1983], it is likely that local mixing processes from small (<10 km diameter) craters would not be solely sufficient to obscure local cryptomaria. However, several young (Imbrian-aged) craters with diameters > 25 km lie in and around these older basins. These may have deposited significant amounts of ejecta which obscured earlier (pre-Imbrian) mare deposits. Craters such as Langeman and Lobachevsky in Al-Khwarizmi/King, Chang...
Heng and Lomonosov in Lomonosov-Fleming, and Haldane in Smythii are good candidates for this process.

Another possibility is that cryptomaria were blanketed by emplacement of extensive ejecta deposits through basin formation. It is believed that the effects associated with the Orientale basin-forming event may have reached ~1500 km away into South Pole-Aitken basin [Wilhelms, 1987; Head et al., 1993]. This suggests that basin-forming events have potentially extensive effects. Thus nearby basins that formed subsequent to Smythii, Lomonosov-Fleming, and Al-Khwarizmi/King may have been in close enough proximity to have contributed to the obscuring of any previously emplaced cryptomaria. For the case of Smythii basin, there are several Nectarian-aged craters which might have contributed mantling ejecta; among these are Al-Biruni and Hubble in Marginis basin, Lomonosov in Lomonosov-Fleming basin, and most significantly, Neper and Jansky craters within the outer Smythii basin ring. In addition, the minimal difference in Al/Si ratios between the western furrowed plains within Smythii and the adjacent highlands [Andre et al., 1977], particularly in the region southeast of Crisium [Clark and Hawke, 1987], indicates a chemically homogeneous stratigraphic layer, the most likely source for which is Crisium ejecta [Andre et al., 1977]. Thus the formation of Crisium basin is also a candidate event for obscuration of cryptomaria. For Lomonosov-Fleming and Al-Khwarizmi/King, the formation of Smythii and Crisium basins subsequent to the formation of these two basins [Wilhelms, 1987] would have contributed ejecta material to mantle cryptomaria in both basins. Due to the extreme age of Lomonosov-Fleming and Al-Khwarizmi/King [Wilhelms, 1987], several Nectarian basins are also potential candidates for contributing to the mantling of cryptomaria material, such as Mendeleev and Moscovienne. On the basis of the above evidence, we conclude that the existence of dark-halo impact craters associated with Smythii and Marginis basins represent cryptomaria buried primarily by the mantling ejecta of younger impact events. Thus the significance of early volcanic activity may have been underestimated [e.g., Schultz and Spudis, 1979, 1983; Hawke et al., 1985; Head and Wilson, 1992], and estimates of the onset of volcanism and the total volume of mare deposits for this region should both be regarded as minimum values pending constraints on cryptomare volume and stratigraphy.

4.2.3. Floor-fractured craters. Eight ponds (21% of occurrences) on the floor of Smythii basin occur in floor-fractured craters (e.g., Doyle and Camoens, Figure 7). Ponds occurring in floor-fractured craters are similar in texture, albedo, and volume to other lava ponds in the region, suggesting a similar emplacement mechanism. However, it is also observed that ponds within floor-fractured craters tend to occur in the young late Imbrian-aged craters that ring Mare Smythii. Craters of this age, which are abundant on the floor of Smythii basin, formed concurrently with the most voluminous period of mare emplacement.

We previously noted two general models for the formation of floor-fractured craters: (1) igneous intrusion and (2) viscous relaxation. Both models depend strongly on the history of thermal activity for a given region. The igneous intrusion model [Schultz, 1976; Brennan, 1975] involves the shallow (a few hundreds to a few thousands of meters from the surface) injection of a laccolith or sill beneath a crater, which drives crater modification through floor uplift [Wichman, 1993; Wichman and Schultz, 1995]. According to the model, such laccoliths would serve as reservoirs for magma, which then move to the surface through the resulting fractures. Currently, however, no features such as large shields have been observed that would suggest derivation of the associated lava ponds through this type of low-pressure, low-effusion rate mechanism. Thus, if floor-fractured craters are formed from the shallow igneous intrusion of a laccolith, that laccolith does not appear to be the subsequent reservoir for the pond. Rather, both pond and laccolith (if such exists) must have originated from a deeper source that would provide the high driving pressures consistent with morphologies like those in this region, as suggested by Wichman and Schultz [1995]. Indeed, such a scenario of laccolith emplacement manifesting itself as a system of fractures would be consistent with the range of features predicted by Wilson and Head [1996], for surface manifestations of magma propagated through dikes fed from sub-crustal source regions. However, the hydrostatic arguments [e.g., Solomon, 1975] invoked by Wichman and Schultz [1995] in order to emplace the relatively dense laccolith high into the lower density crust may not be fully applicable. Magma transport solely by hydrostatic rise may be overly simplified in terms of the ability of a dike to remain open through 20-50 km of crust [Head and Wilson, 1992]. In addition, for formation of the associated pond to occur, a mechanism other than hydrostatic rise must then be employed which overcomes both the density barrier presented by the lunar crust, and the decrease in driving pressure resulting from laccolith emplacement. Finally, it must be noted that a model of laccolithic intrusion necessarily requires a large number of dikes in the lunar crust. Since this model assumes the magma column supporting these laccoliths extends into the mantle, the propagation of conduits into the overlying crust is probably not affected by the formation of a crater on the surface, at least 20 km above. Thus, for every dike that actually emplaces a laccolith in proximity to a crater to produce floor uplift, there must be several that either do not reach near-surface levels, or are not emplaced under a crater. Unless the density of dikes in the crust (a number which is currently very poorly constrained) is very high, the sheer number of floor-fractured craters in this region seems to favor a regional, rather than a local origin.

In contrast to this model, the relaxation model employs local crustal heating during mare emplacement to lower local viscosity. This allows viscous relaxation to occur at a faster rate than in other regions [Danes, 1965; Baldwin, 1968; Hall et al., 1981]. This model has the benefit of not requiring shallow emplacement of high-density material into a low-density (brecciated) crust, making it more consistent with morphological and geophysical constraints. However, the proximity of floor-fractured craters to Mare Smythii suggests that the volcanism associated with Mare Smythii could have been the source of heat required to produce lower crustal viscosity in the surrounding area. The concentration of volcanic deposits in the Smythii basin center suggests that the heat flux might have been strongest here, so that viscous relaxation might have been more prominent. However, this does not explain why Mare Orientale has a larger volume than Mare Smythii [Head, 1982] and would presumably have had an even greater effect on the local crust, yet has no floor-fractured craters.

Wichman and Schultz [1995] note in their observations a positive correlation between the diameter of a floor-fractured crater and the extent of crater uplift. This suggests the process
involved must be a regional phenomenon or must explain why craters alone are affected. A model which we suggest as a derivative of the above models is one in which thermal conditions create a favorable environment for both viscous relaxation and volcanism. Because significant mantle uplift is indicated in Smythii basin [Neumann et al., 1996], this may have provided a mechanism for relatively near-surface crustal heating and thus regional viscous relaxation. Neumann et al. [1996] state that there is a decrease in the relief of the lunar Moho (uplift of the mantle) with increasing basin age, such that the oldest basins are the most isostatically compensated. Specifically, older pre-Nectarian basins, such as South Pole-Aitken, Facetuditatis, Australe, and Tranquillitatis, are isostatically compensated but the younger Smythii is not [Neumann et al., 1996], suggesting that it was only in the later part of the pre-Nectarian period that the lunar lithosphere was strong enough to maintain a high state of stress. Thus older basins would have no evidence of localized (crater) relaxation because at that time lateral movement of crustal material would have made long wavelength relaxation of the entire basin possible. By contrast, during the later stages of the pre-Nectarian, the crust would have been sufficiently strong to resist isostatic compensation. Consequently, conduction through the crust of the heat provided by the uplifted mantle in Smythii would have yielded a relatively thermally mobile basin center. This would have presumably been a conducive environment for local relaxation. Such local relaxation of relatively young craters would not be evident in younger basins such as Crisium [Wilhelms, 1987], because basin floor craters would have been buried by subsequent voluminous mare emplacement episodes. Thus, the lack of floor-fractured craters in Crisium would be a consequence of the volcanic activity that emplaced Mare Crisium [Head et al., 1978]. It should be noted that recent analyses of Oriental basin using Clementine images [Head et al., 1997] suggest that Kopff crater may very well be a floor-fractured crater. If such is the case, it would be of significant finding in terms of determining the nature of the thermal environment in which floor-fractured craters form, since the young Oriental basin would have a very different thermal structure than the older basins.

Finally, this model does not depend on hydrostatic rise as a mechanism for fracture formation or mare extrusion. Magma in the above model passes directly from reservoir to surface through dikes (the mechanism known to transport magma through brittle crust), held open by a state of overpressurization in the source region instead of hydrostatically [Head and Wilson, 1992]. Such reasoning also removes the obstacle of finding a mechanism to decipher why ponds are associated with floor-fractured craters. Instead, this model decouples the mechanism responsible for the formation of floor-fractured craters (local relaxation due to heating from below) with that for the emplacement of mare deposits (overpressurized reservoirs generated by the same heating). We therefore postulate that uplift of the lunar mantle provided sufficient heat to viscously relax the crust in a local sense, forming floor-fractured craters, and yielded an accessible source of magma for lava pond emplacement.

4.3. Modes of Occurrence

Ponds in Smythii/Marginis show a preference for deposition in impact craters, as shown in Figures 5 and 6. Occurrences of ponds in craters accounted for 70% of all individual eruptive deposits. A similar preference for pond occurrence in impact structures was observed for deposits in the Orientale, Mendel-Rydberg, and South Pole-Aitken basins. In these regions, approximately 70% of all ponds were found in craters, superimposed basins, or within the low-lying Orientale basin ring [Yingst and Head, 1997a]. For all regions, ponds found in craters and superimposed basins generally displayed higher average volumes than those lying in intercrater highlands. Many of the largest lava ponds (e.g., the Apollo deposits in South Pole-Aitken, Kiess and Helmhert-Kao craters in Smythii) lie within the deepest basins and the largest craters. Together, these observations suggest both a higher total volume of extrusion, and a higher frequency of eruptive events for topographic depressions and lows than for highland regions. These observations are consistent with the model of Head and Wilson [1992], where local-scale variations in crustal thickness affect mare extrusion. Further implications of this model are discussed in more detail later.

4.4 Areal Distribution

Average volume distribution, frequency, and spacing of lava ponds in various regions yield important constraints on the different characteristics and geometry of magma reservoirs associated with these mare deposits. Typical volumes for lava ponds in the limb and farside regions are within the range of 195 to 860 km$^3$ [Yingst and Head, 1997a; this study]. Adopting a geometry of 100 x 100 x 0.25 km for the dikes feeding these flows [Head and Wilson, 1992], and assuming that the total volume (dike plus pond) typically represents about 1% of the total volume of the reservoir [Blake, 1981], then each pond could potentially represent an eruption from a reservoir with a volume of ~270,000-340,000 km$^3$. If we further assume an ideal spherical magma reservoir, such a volume yields a diameter of approximately 80-90 km. This diameter range is similar to the average range of nearest-neighbor distances for all basins examined. Note that a difference of ~600 km$^3$ in mean volume (such as that found between the mean volumes of the limb areas and South Pole-Aitken) translates into only a 10 km difference in the calculated reservoir radius. The presence and spacing of lava ponds may thus provide an indication of the frequency and geometry of magma reservoirs at depth.

While average nearest-neighbor distance values are an important factor in determining reservoir geometry, nearest-neighbor distances for individual ponds may yield information regarding pond clustering associated with the parent source regions. For example, if we accept the above estimates for reservoir diameters as reasonable, the fact that many ponds are separated from one another by distances of less than 100 km (such as the Doyle-CamOens Haldane region; Figures 2a, 7) suggests that such clusters may represent a population of ponds derived from one reservoir. In other cases, ponds are separated by much larger distances. For example, the pond northwest of Erro crater (pond 2 in Figure 2a) is separated from the next nearest pond by ~150 km. Individual ponds such as these are candidates for separate reservoirs. In order to assess the relevance of nearest-neighbor distances in understanding the sequence of emplacement for these deposits, spectral and age characterizations, as well as constraints on the extent of possible cryptomaria, are required as a next step.

The general distribution of ponds in Smythii and Marginis basins appears to be related to basin degradation state. In Smythii, ponds are concentrated within the central portion of the basin, while in Marginis pond distribution is more diffuse.
This correlation suggests a connection between pond occurrence and basin age. One plausible explanation is that differences in the thermal regime of each basin influenced the concentration of magma reservoirs. Thus, because Smythii is younger (and thus uncompensated [Neumann et al., 1996]), heat was focused where mantle uplift occurred, namely, in the central basin. This is where the greatest number of reservoirs formed, or alternatively, where diapirs were able to penetrate to a shallower depth. Conversely, Marginis, an older compensated basin [Neumann et al., 1996], would have had no such concentration of a heat source in the basin center, so that reservoir distribution, and thus associated pond concentration, would be more diffuse. It is also possible that the observed pond distribution is due to mantle heterogeneities that relate to mechanisms not currently well constrained.

4.5. Crustal Thickness Relationships

It is clear that the areal and volumetric mean values for ponds in Smythii and Marginis basins are more comparable to the younger Orientale basin than to South Pole-Aitken basin. In fact, the mean value for volume of magma extruded in an individual eruptive episode in the three smaller basins is nearly equal (~195-250 km³), while that for South Pole-Aitken is more than 3 times larger. In addition, as previously mentioned, there are many more relatively low-volume ponds in the limb basins than in South Pole-Aitken basin. There are some hypotheses which might explain the greater mean volumes of individual eruptive events on the farside.

Some South Pole-Aitken lava ponds might be the products of multiple flows, and thus yield a larger mean volume estimate for individual eruptive episodes than reality dictates. Although this possibility cannot be discounted, there is currently no evidence (e.g., variations in albedo, different crater densities, other characteristics discussed above) to suggest that this is the case in a general sense. High resolution multispectral imaging data is currently being utilized to further test this possibility (e.g., Clementine multispectral image data [Yingst and Head, 1997b]), but preliminary data support the interpretation that the vast majority of the basins represent distinct episodes.

It is possible that, given a relatively homogeneous distribution of dikes within the crust below these regions, the older South Pole-Aitken basin was volcanically active longer and thus sampled more large volume eruptions over time. However, it has been suggested that a lower thermal gradient existed for this region at the time of basin formation [Solomon et al., 1982; Neumann et al., 1996], which would have resulted in fewer magma sources, with shorter cooling times and possibly lower degrees of partial melting. This would have served to decrease the number and volume of eruptions within the farside basin compared to those on the lunar limbs. Another possibility is that factors such as the number of tappable magma reservoirs, or the degree of partial melting associated with these reservoirs, are variable on a global scale. These latter factors cannot currently be established or ruled out.

Finally, the average erupted volume may be a function of variations in crustal thickness of different lunar regions [e.g., Head and Wilson, 1992; Robinson et al., 1992]. Estimates of lunar crustal thickness (T_c) derived from altimetry data obtained by the Clementine laser altimeter [Zuber et al., 1994] show that there is a very close correlation between the average volume of lava ponds and regions of topographic lows associated with thin crust (compare Figures 2a and 2b). This relationship is shown for the farside in Figure 13, which displays the volume of mare material observed in South Pole-Aitken basin as a function of the estimated crustal thickness (shown in increments of 5 km). Volume has been normalized to the total area within South Pole-Aitken basin associated with each 5 km increment of crustal thickness. Thus the volume at each increment of T_c represents the effective thickness of mare fill that would exist if the total volume of mare material lying at that thickness value were spread in a uniform layer across the corresponding area. For the example of South Pole-Aitken basin, the bulk of mare material and the majority of ponds occur where the crust is thinnest (T_c < 50 km; only 40% of the basin area). Areas of thicker crust (50-70 km) contain a smaller number of ponds and a very small total volume [Yingst and Head, 1997a]. Smythii and Marginis basins also follow this general trend. In general terms, regions of major maria (Mare Smythii and Mare Marginis) occur in areas of thinnest crust, while ponds tend to occur in regions that have thicker crust. Overall, it can be seen in Figure 14 that most ponds occur in regions where crustal thickness is estimated to be < 50 km. About 61% of the total volume of ponds analyzed occurs within just 31% of the region, the area corresponding to T_c < 50 km. Only 39% occur in the remaining areas of thicker crust. In terms of eruption frequency for mare ponds, 21 eruptive occurrences (62%) lie at T_c < 50 km, while 13 (38%) occur at values of 50 km or above. In addition, we can analyze the contribution of the major maria (associated with crust 40 km thick or less) to the number of eruptive events, by estimating the number of eruptions these deposits represent. We assume a typical volume range for individual eruptive phases in the Smythii/Marginis region of 190-270 km³, as suggested by the volume data enumerated earlier. Dividing these averages by the combined volume of Mare Smythii and Mare Marginis yields ~200-300 potential

![Figure 13. Total volume of lava extruded compared to crustal thickness for South Pole-Aitken basin. In order to avoid sampling bias, the total volume of mare material has been normalized to the amount of surface area occupied by each estimated crustal thickness value. Thus the numerical value at each increment of crustal thickness represents the thickness of mare fill that would exist if the total volume of mare material lying at each thickness was spread in a uniform layer across the area corresponding to that crustal thickness value. This figure shows the inverse correlation between the volume of mare material and the thickness of the lunar crust.](image-url)
eruptive phases that these regions may represent. Thus, the highest frequency of eruptive episodes appears to have occurred in regions of the thinnest crust.

In terms of maria in specific basin regions, we have observed that the total extruded mare volume is a function of the thickness of the intervening crustal column. Determining the role of crustal thickness in mechanisms of mare transport, however, is ultimately a global issue. Thus a useful exercise is to consider estimates of the total lunar mare volume as a function of $T_c$, where the large maria whose volumes are known are included. Although the number of individual eruptive episodes cannot be estimated because of the reasons stated in our approach, an estimate of the total volume of mare material deposited at each crustal thickness value across the Moon is possible. This is shown in Figure 15, where the estimated total volume of mare material for deposits on the lunar limbs and farside, as well as the nearside contiguous mare regions for which volume estimates are available, is plotted against the estimated crustal thickness (displayed in increments of 5 km). These regions include South Pole-Aitken and Orientale basins [Yingst and Head, 1997a], Crisium, Humorum, Nectaris, Imbrium, and Serenitatis nearside basins [Solomon and Head, 1980], and Smythii and Marginis basins (this study). As before, volume has been normalized to the total area associated with each 5 km bin of crustal thickness so that effective mare thickness for each $T_c$ value is displayed. As shown in this figure, the total volume of mare material is inversely related to crustal thickness in that areas of thinnest corresponding crust show the greatest amount of mare material on the surface. Specifically, 66% of the total mare volume measured occurs on only 12% of the lunar surface, the area represented by $T_c < 40$ km. In addition, it is seen that regions on the nearside which are characterized by the large volumes of the contiguous maria also have the thinnest crust. As stated above, the number of eruptive occurrences corresponding to the total mare volume cannot currently be determined. This is to be expected, however, if crustal thickness is directly related to eruption volume. The number of occurrences becomes more difficult to judge as $T_c$ decreases because the increased volume of deposits effectively obliterates our ability to discern individual flows.

We conclude, on the basis of the very strong local and global correlations observed between the magnitude and frequency of magma eruption and crustal thickness differences, that it is likely that the ability of magma to reach the surface from subcrustal reservoirs is directly related to the thickness of the intervening crustal column. This connection has been observed in other regions of the Moon [e.g., Robinson et al., 1992], and has been suggested by Head and Wilson [1992] to be the result of diapirs stalled under crustal columns of varying height. A schematic representation of this model for the limb and farside basins is shown in Figure 16. Assuming a relatively homogeneous distribution of magma sources throughout the lunar mantle, mare emplacement in this model is dependent on the level of overpressurization which these source regions reach upon stalling at a boundary defined by the low density lunar highland crust. This overpressurization would, in turn, drive the propagation of dikes from depth. For dikes at equal levels of overpressurization, those dikes emplaced into regions of thinner crust reach the surface more readily than those propagating into thicker crust. This model implies that the height of the overlying crustal column is pivotal in mare transport and distribution. The amount of this intervening crust determines whether dikes propagated from overpressurized source regions are able to extrude onto the surface (at areas of thinnest crust) or must stall and freeze (at regions where the crust is thicker). A high degree of correlation between crustal thickness and the number and magnitude of individual eruptive events is required. The observations discussed here for Smythii/Marginis are consistent with both the analyses of other lava ponds on the lunar limbs and farside, and with the predictions implied by this model. This model also predicts that, because dikes must be driven from depths of at least the base of the crust, each episode of mare emplacement is likely to be associated with a high effusion rate and thus large volumes of basalt. Emplacement episodes would be evidenced by relatively smooth deposits showing a lack of features associated with shallow source regions (e.g., calderas, large shield volcanoes). Again, these predictions are consistent with what we have observed for lava ponds in Smythii/Marginis and on the lunar limbs and farside [Yingst and Head, 1994, 1996, 1997a].

![Figure 14. Total volume of lava extruded compared to crustal thickness for Smythii/Marginis basins, normalized to amount of surface area as in Figure 13.](image1)

![Figure 15. Crustal thickness versus the total volume of mare material for the Moon, normalized to amount of surface area as in Figure 13. Regions were chosen on the basis of availability of volume estimates. These regions include South Pole-Aitken and Orientale basins [Yingst and Head, 1997a], Crisium, Humorum, Nectaris, Imbrium and Serenitatis nearside basins [Solomon and Head, 1980], and Smythii/Marginis basins (this study).](image2)
Pole-Aitken and Australe basins may have played a role in explained by several hypotheses. Dated as older Late Imbrian in age. This observation may be mantle deposits. In South Pole-Aitken and Australe, by Marginis have a large population of deposits classified either as Late Imbrian, or undivided Late Imbrian/Eratosthenian dark Late Imbrian (3.80 - 3.20 Ga) in Orientale basin.) Smythii and the base of the Late Imbrian period, no ponds can be older than mare deposits in this region are somewhat younger in between emplacement periods for these flows occurring in Head, 1994]. Grimaldi basin is 170 km in diameter. In this comparison to ponds on the farside [Wilhelms et al., 1979; 4.6. Stratigraphy

The local stratigraphic profile for the Smythii/Marginis region (shown in Figure 10) suggests that, since the deposits in this region are all relatively close in age, there may have been a period of active volcanism in this region around 3.85-3.60 Ga. However, almost 200 myr appears to have elapsed between the emplacement of the pond in Ibn Yunes crater and the portion of Mare Marginis due south of that crater (region III in Figure 2a). Similar observations were made for mare deposits in Grimaldi basin northeast of Orientale [Yingst and Head, 1994]. Grimaldi basin is 170 km in diameter. In this relatively small area, three flows of varying compositions are represented, dated at 3.12, 2.79, and 2.49 Ga [Greeley et al., 1993; Williams et al., 1995]. The large periods of quiescence between emplacement periods for these flows occurring in relatively small areas suggests that (1) parent source regions have a very long cooling time (200-750 Ma) or (2) magma reservoirs are being emptied and then replenished from depth. On the basis of our present understanding, we favor the interpretation that these deposits represent the sequential emplacement of material derived from different source regions at depth, whose formation was separated in time by several hundred million years.

The age-frequency distribution of lava ponds in Smythii/Marginis displayed in Figure 10 suggests that the mare deposits in this region are somewhat younger in comparison to ponds on the farside [Wilhelms et al., 1979; Hiesinger et al., 1996]. In Smythii/Marginis, only five ponds, all of which are within the older Marginis basin, are dated as older Late Imbrian. (Because the formation of Orientale defines the base of the Late Imbrian period, no ponds can be older than Late Imbrian (3.80 - 3.20 Ga) in Orientale basin.) Smythii and Marginis have a large population of deposits classified either as Late Imbrian, or undivided Late Imbrian/Eratosthenian dark mantle deposits. In South Pole-Aitken and Australe, by comparison, a large percentage of the subdivided ponds are dated as older Late Imbrian in age. This observation may be explained by several hypotheses.

As has been suggested above, the extreme age of South Pole-Aitken and Australe basins may have played a role in mare deposit distribution, in that mare deposits would have had a longer time to accumulate in an older basin. Specifically, ponds in South Pole-Aitken and Australe were emplaced throughout the Late Imbrian period [Wilhelms and El-Baz, 1977; Wilhelms et al., 1979; Hiesinger et al., 1996], while ponds in Smythii/Marginis and Orientale were emplaced during the upper Late Imbrian and Eratosthenian periods [Wilhelms and El-Baz, 1977; Scott et al., 1977; Hiesinger et al., 1997]. However, the volcanic activity in all regions have the same apparent starting point. Only the number of deposits emplaced during each period are different. The formation of the farside and eastern limb basins predates the apparent onset of mare volcanism. In addition, ejecta deposits from the Orientale basin have emplaced a stratigraphic datum at ~ 3.8 Ga, burying any pre-Orientale mare material at least in South Pole-Aitken and areas in the proximity of Orientale [Head et al., 1993]. Thus post-Orientale mare emplacement in South Pole-Aitken and Orientale appears to have occurred over similar periods in the case of these basins.

An alternate theory, which we propose here, is that, due to thin crust, some areas of the Moon were filled preferentially during the early stages of lunar volcanism and remained active for several periods, while other nearby regions were filled only during the later stages. Consider for example, Crisium basin, which lies so close to the Smythii/Marginis region that their basin rings overlap [Wilhelms and El-Baz, 1977]. Crisium maria represents three different and comparatively voluminous stages of volcanism [Solomon and Head, 1980], spanning a time period from possibly the earliest stages of volcanic history to less than 3.5 Ga [Boyce et al., 1977; Head et al., 1978]. Crisium basin has been estimated to have a corresponding crustal thickness which varies from 12-25 km, while crust in the Smythii/Marginis region ranges from 25-60 km [Zuber et al., 1994]. If the thickness of the crust determines the likelihood of a magma conduit reaching the surface, as has been suggested above, regions with a relatively thinner crust would have preferentially been filled first, and would have been subject to voluminous mare emplacement for longer periods of time than nearby areas. Basins such as Crisium, Australe and South Pole-Aitken would thus display a wider range of deposits ages because they have been active for longer periods of time due to the thinner crust in that particular region. Thicker crust in surrounding regions (such as Smythii/Marginis in the case of Crisium) would thus cause local mare flooding to be constrained to a later, possibly more active stage. Regions of very thick crust (such as that surrounding Australe) would show no volcanic activity. Wilhelms [1987] dates the deposits within Mare Crisium as relatively younger than those in Marginis basin, and contemporaneous with those in Smythii basin. However, if the volcanic flux on the eastern limb was similar to that calculated for the Moon during the Late Imbrian [Hartmann et al., 1981; Head and Wilson, 1992], it is possible that many Early Imbrian flows emplaced in the lowest lying areas are now obscured because they were covered up by later, more voluminous flows. More precise age determination of the deposits within Crisium and other regions, as well as the testing of the models presented here, awaits crater-frequency dating derived using high-resolution Clementine images.

Finally, it is possible that the mare material observed on the surface is not indicative of the full stratigraphic range of mare emplacement episodes. As has been noted previously, Nectarian-Imbrian furrowed and pitted material is extensive in the region north of Mare Marginis and in a subcircular region
around Mare Smythii [Wilhelms and El-Baz, 1977]. The geomorphology of this material in Smythii basin is particularly striking in that its boundary is very closely defined by the inner ring of the basin, similar to the boundary of volcanic material that fills the central portion of Orientale basin. It has thus been suggested that this region represents basalt that has been reworked by subsequent cratering or other processes [Stewart et al., 1975]. Because this region is similar in Al/Si ratios to surrounding highlands soils and dissimilar to nearby mare deposits, a soil composed solely of highly reworked basalt is not indicated [Andre et al., 1977]. However, there is only a minimal difference in Al/Si ratios between furrowed plains soils and highland soils to the west of Smythii basin, an observation that has been interpreted as indicating a chemically homogeneous ejecta layer emplaced prior to the Late Imbrian period of mare volcanism [Andre et al., 1977; Wilhelms and El-Baz, 1977]. This plains-type material may thus cover older cryptomaria, which might have filled vast portions of the basin in the Early Imbrian or earlier. As previously noted, the presence of features such as dark halo craters is a possible indicator of a scenario such as this [Schultz and Spudis, 1979, 1983; Hawke and Bell, 1981; Bell and Hawke, 1984; Antonenko et al., 1995]. The geochemical data provided by the Apollo X-ray and gamma-ray spectrometers did not directly indicate the presence of any mare material in the light plains region [e.g., Adler et al., 1972; Metzger et al., 1973; Wilhelms and El-Baz, 1977]. However, multispectral data from Mariner 10 has been used to identify a region northeast of Mare Marginis that has a relatively strong spectral signature in the 0.40/0.56 μm ratio and thus is spectrally bluer than the surrounding soils [Robinson et al., 1992]. This area, identified as intermediate in composition between mare and highlands units [Robinson et al., 1992], corresponds to the furrowed plains unit of Wilhelms and El-Baz [1977]. Such spectrally blue units have been previously associated with cryptomare deposits [Metzger and Parker, 1979; Davis, 1980]. Thus the geologic setting of this furrowed plains unit, combined with its multispectral signature, suggests that soils in this region contain a cryptomare component. Examination using higher resolution Clementine multispectral data in order to search for further mineralogic indications of mare material within this plains unit would be a next step in resolving the stratigraphic profile of volcanic activity in Smythii/Marginis.

5. Local-Scale Variations

Thus far, we have presented observations of the characteristics of discrete mare ponds on the Moon's eastern limb and the general trends which these observations follow with respect to other similar localized deposits on the lunar western limb and farside. While the deposits within the Smythii/Marginis study area are consistent with the trends noted, there is one local-scale variation of mare pond characteristics for which there is currently insufficient data to make any firm conclusions, but is noted here in the context of future studies. This variation relates to a difference between the mode of occurrence of ponds in the Smythii basin and Marginis basin.

In Smythii basin, more than 86% of the mare deposits studied occur in relatively young (Imbrian-aged) craters, while in Marginis, less than 50% lie within craters. Thus the difference in frequency of occurrence may be due to the fact that fewer Imbrian craters exist within Marginis than within Smythii, implying a connection between the age of the associated crater and the emplacement mechanism. Head and Wilson [1992] suggest that local thinning of the crust in the vicinity of impact craters creates a favorable environment for mare emplacement. Formation of a crater thins the local crustal regime, but does not affect the overall pressure conditions at subcrustal depths. All things being equal, a dike propagating beneath a crater would thus have a greater chance of intersecting the surface because the local crust was thinner. This implies that young craters (Imbrian-aged), which would be relatively fresh during the period of local volcanism, would be somewhat more likely candidates for mare emplacement than older craters (pre-Nectarian or Nectarian) which would be more degraded and thus more shallow at the time of active mare volcanism. The greater frequency of Imbrian-aged craters in Smythii basin could thus provide more favorable regions for mare extrusion.

As indicated previously, it is also possible that there are differences in the thermal history of the region that favor emplacement of mare within craters formed around the time of mare extrusion. Superposition relationships suggest that Smythii basin is younger than basins such as Fecunditatis, Australe, and Tranquillitatis [Wilhelms, 1987]. These older basins display a compensation signature which is lacking in the younger Smythii basin [Neumann et al., 1996], indicating that the thermal structure of the Moon evolved so that there was a point in the pre-Nectarian after which cooling of the outer layers of the Moon prevented basins from reaching a fully compensated state. Thus minor crustal thickness differences and increases in mantle uplift caused by the formation of impact craters would have been smoothed out in the older basins. In Smythii basin, where differences in local compensation would still have existed, craters might have provided a more favorable environment for extrusion.

6. Conclusions

Analysis of 41 mare ponds and dark mantle deposits has provided information on associated features, modes of occurrence, and the range and frequency distribution of eruption areas and volumes in Smythii and Marginis basins. On the basis of this analysis, we find the following.

1. The majority of deposits likely to represent single eruptive episodes have areas which lie in the range of 170 to 6575 km², with an average value of 965 km². This translates into a mean area of 720 km² for Smythii basin and 1830 km² for the Marginis basin region.

2. Volumes for these deposits range from 15 to 1045 km³, with a mean value of 195 km³. For Smythii basin the mean value for deposit volume is 190 km³, while for Marginis basin it is 270 km³.

3. Deposits tend to occur preferentially within areas of locally thinned crust (craters). In terms of pond density, concentration of mare deposits appears to be related to the state of preservation of the basin, such that deposits are highly concentrated in the center of the better preserved Smythii basin, while in Marginis basin the distribution of deposits is more diffuse.

4. No definitive morphological evidence, such as large shield volcanoes or collapse calderas, was found for shallow crustal magma reservoirs in association with these ponds.
Features associated with isolated lava ponds (such as linear rilles) are consistent with emplacement and extrusion onto the surface through dikes from deep, perhaps subcrustal reservoirs. In terms of floor-fractured craters, some models suggest that the existence of such craters indicates shallow laccolithic intrusion. However, such a model also requires that both the laccolith and the pond stem from the same deep-seated reservoir.

5. The presence of such indicators as geochemical/multispectral anomalies and abundant dark-halo impact craters in the Smythii/Marginis region suggests the presence of previously deposited cryptomare material. Based upon the distribution of these indicators, such cryptomaria, if it exists, may be a significant volumetric component of the total mare material in the area (e.g., potentially twice the volume of presently exposed mare deposits).

6. Peak volcanism in this region as represented by exposed mare deposits appears to have occurred at around 3.80-3.60 Ga. Because of the possible existence of cryptomare material, as indicated by dark-halo craters, this should be considered as the latest period at which the onset and peak of volcanic activity could have occurred. In a preliminary analysis, Spudis and Hood [1992] have reported evidence for the presence of young mare deposits in Mare Smythii, possibly younger than Apollo 12 mare material (~3.20 Ga), although the location of their crater counts within Smythii were not specified. Further crater size frequency distribution data for Smythii may reveal evidence for young deposits.

We have compared the characteristics of discrete mare deposits on the eastern limb with those of similar deposits on the western limb and fissade in order to put Smythii and Marginis ponds into a global context. On the basis of the observations made, we infer the following.

1. Volumes of eruptive events on the Moon are very large compared to most terrestrial eruptions, and seem to have their best terrestrial analog, geomorphologically and volumetrically, in flood basalts rather than small volume eruptions derived from shallow reservoirs.

2. Nearest-neighbor distances average about 60-100 km, suggesting a constraint for the diameter of source regions to within or below this range. Typical lava pond volumes and nearest-neighbor distances for ponds on the limbs and fissade of the Moon suggest that magma is derived from subcrustal reservoirs <100 km in diameter. Thus many ponds, especially those lying in the furthest rings of their respective basins, are likely to be solitary representatives of their source regions. By the same reasoning, other ponds that lie more tightly spaced, notably those within the central portions of the basin regions, may be members of a pond cluster which originates from a single source region.

3. For each individual basin, the volume and frequency of eruptions is related to the amount of crust through which the magma must pass. Those areas with thinner crust have a greater number and volume of mare occurrences on average relative to regions of thicker crust. Although the number of eruptive events cannot currently be determined in the large contiguous maria for the Moon as a whole, regions of thinner crust are likely to be solitary representatives of their source regions.

Acknowledgments. We gratefully acknowledge the assistance of Irene Antonenko, B. Ray Hawke and an anonymous reviewer who provided reviews of earlier versions of this manuscript, Peter Neivert, who provided photographic support, and funding from NASA Grant NAGW-713 to J.H. from the National Aeronautics and Space Administration Planetary Geology and Geophysics Program.

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