Large shield volcanoes on the Moon

Paul D. Spudis,† Patrick J. McGovern,† and Walter S. Kiefer†

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[1] The volcanic style of the Moon has long been understood to consist almost exclusively of flood basalts erupted from fissures along with minor pyroclastic activity; large central vent shield volcanoes that characterize basaltic volcanism on the other terrestrial planets appeared to be absent. Small (few kilometers diameter) central vent constructs have long been recognized in the lunar maria and often are found clustered in fields throughout the lunar maria. New global topographic data from the LOLA and LROC instruments on LRO reveal that almost all of these volcanic complexes on the Moon occur on large, regional topographic rises in the lunar maria, tens to hundreds of kilometers in extent and between several hundred to several thousand meters high. We propose that these topographic swells are shield volcanoes and are the lunar equivalents of the large basaltic shields found on the Earth, Venus, and Mars. The newly recognized lunar shields are found peripheral to the large, deeply flooded impact basins Imbrium and Serenitatis, suggesting a genetic relation to those features. Loading of the lithosphere by these basalt-filled basins may be responsible for inducing a combination of flexural and membrane stress, inducing a pressure distribution on vertically oriented dikes favorable to magma ascent. This condition would occur in a zone annular to the large circular loads produced by the basins, where the shield volcanoes occur.


1. Introduction

[2] Although basaltic volcanism is a common process on the terrestrial planets, it manifests itself with differing styles and intensities on different planetary bodies. Volcanism on the Moon is manifested largely by the presence of extensive plains of basaltic lava. The dark smooth lunar maria are composed of basaltic lava flows that were largely emplaced through fissure-fed, flood-style eruptions. This style of volcanism is also common on the other terrestrial planets; both Venus and Mars show vast plains made of basaltic lava in addition to their massive, central-vent shields. On Earth, continental flood basalts eruptions are considered the best analog for mare volcanism, with high-effusion rate flows of fluid lava creating vast plateaus of basalt [e.g., Swanson and Wright, 1978].

[3] Central vent, shield-building volcanism is common on Earth, Venus, Mars, and Io and may also have occurred on Mercury. Small shield and dome volcanoes have been observed and mapped on the Moon for many years, but typically are very small (2–10 km diameter) and occur in groups or clusters within selected areas of the maria [McCausley, 1964; Greeley, 1971; Guest, 1971; Whitford-Stark and Head, 1977]. The vast bulk of lunar volcanic deposits is flood lavas, in which large volumes of magma are erupted rapidly from fissures and spread out as sheets on the surface [e.g., Head, 1976]. The Moon seems to lack the very large shield volcanoes [BVSP, 1981; Head and Wilson, 1991] that typify some of the mountains of Earth, Mars, and Venus [Pike, 1978; BVSP, 1981; Plescia, 2004; Herrick et al., 2005]. Or does it?

[4] Shield volcanoes are positive-relief, central vent structures that are broader than they are high, so they have relatively low, average positive slopes [Whitford-Stark, 1975]. The term was first coined to characterize the shape of certain lava constructs on Earth made up principally of low viscosity, basaltic lava that builds up a broad, shield-shaped construct. The bulk of the volcano is made of lava flows, although pyroclastic activity may occur in minor amounts, particularly during late stage eruptions. Many shield volcanoes display a summit crater (caldera) resulting from collapse of the surface over a drained or depleted magma chamber, but some shields do not have a summit crater [Whitford-Stark, 1975; Herrick et al., 2005] and such is not required for the edifice to be classified as a shield volcano. Shield volcanoes typically have both radial and circumferential fissure zones, which serve as pathways for magma to get to the surface and erupt a continuing supply of lava. Parasitical cone and dome building often occurs near the summit and on the flanks of such features during the latter stages of shield growth [e.g., McDonald and Abbott, 1970].

[5] Until recently, regional topographic information for the Moon was sparse and non-contiguous. Nonetheless, substantial regional slopes were evident in the Clementine
global altimetry [Zuber et al., 1994] for the relatively “flat” maria of the Moon. Such slopes tend to conform to the configuration of the containing impact basins, but it was noted that some igneous centers in the maria occur on topographic rises. Specifically, the lunar Marius Hills complex was found to occur on the summit of a broad, gentle topographic swell, leading to the supposition that this complex might be the lunar manifestation of a basaltic shield volcano, a couple of hundred kilometers across and several hundred meters high [Spudis, 1996]. The Clementine topography was of low resolution and could not resolve features within the lunar maria with high precision. However, new global data from the LRO laser altimeter [Smith et al., 2010] give us a high-resolution view of lunar topography. Moreover, global stereo images from the LRO camera have been processed into a global topographic map [Scholten et al., 2012]. Thus, it is an appropriate time to re-visit the topographic character of volcanic complexes in the lunar maria and address the question: Do shield volcanoes exist on the Moon?

2. Data Sources and Approach

The new global topographic map of the Moon obtained by the Lunar Reconnaissance Orbiter (LRO) is the principal source of topographic information used in this study. The GLD100 global map [Scholten et al., 2012] is a stereo-model based on LRO Camera Wide Angle stereo image data. It has a resolution of 100 m/pixel, covers the Moon between ±79° latitude, and has been determined to have vertical accuracy of about 18 m compared to the LRO laser altimeter data set. The laser altimeter data [Smith et al., 2010] complete the global topographic maps for latitudes greater than 79° [Scholten et al., 2012]. The features studied in this paper all fall within the boundaries of the GLD100 map and have dimensions of hundreds of kilometers and heights greater than 1000 m, much larger than the scale of this high-resolution topographic data.

Small volcanic features in the maria have been mapped for many years [e.g., McCauley, 1964; Guest, 1971; Wilhelms and McCauley, 1971] and we have used this previous mapping to locate clusters of such small features in relation to our larger landforms. Specifically, we have used the landmark classification and map of Guest and Murray [1976] to show the correspondence of small volcanic features with our larger shields. Guest and Murray [1976] recognized several distinct landforms, including ridges, domes, pits, and cones. Low domes (small shields) are smooth, convex-shaped positive relief landforms with side slopes of 2–3° and sizes of a few kilometers diameter [Guest and Murray, 1976]; some of these features have summit craters while others do not. Steep domes have more prominent topography and are comparable in size to shields, but have steeper sides with slopes of 7–20° [Guest and Murray, 1976]. Cones are small features (2-3 km across), often occur on top of a broader shield (over 40 of these are found in the Marius Hills) or aligned along a linear vent system, and tend to have steep sides (>20°). Collapse craters (or pits) are common throughout the maria and many are found in association with the other landforms; they tend to be small (a few kilometers across or less) and shallow (tens to hundreds meters). Many collapse pits are associated with sinuous rilles, a common feature of these eruptive centers and are interpreted as vent systems and their associated lava channels and tube systems.

We have used the basic classification and mapping of Guest and Murray [1976], including their distinction between shields (or “low domes”) with and without summit pit craters. In the feature maps presented in this paper, we recognize low domes (shields), with and without summit pits, cones, collapse pits, sinuous rilles, and chains of cinder cones (interpreted as fissure vents) [Guest and Murray, 1976]. Additionally, we mapped eruptive vent centers where recognized (indicated by an irregular crater or landform associated with dark mantling materials). We have plotted the locations of the most prominent features in these areas on a shaded relief base (created from the GLD100 topographic map) [Scholten et al., 2012] for each proposed shield volcano. Associated topographic profiles of each shield volcano were extracted from the GLD100 database using the profiling tool of the Quickmap LROC global basemap (http://target.lroc.asu.edu/da/qmap.html).

3. Topography of Volcanic Complexes in the Lunar Maria

The new global topographic map of the Moon reveals many new relationships on the lunar surface. Although these data validate the conventional wisdom that mare deposits occupy low-lying areas of the Moon, several broad topographic highs are found in both the eastern and western near-side maria (Figure 1). These topographic bulges are tens to hundreds of kilometers across and from 600 to over 2200 m high. We have identified six major and two minor topographic swells (Table 1) that occur within the near-side lunar maria. Interestingly, all of these rises correspond to high concentrations of small (kilometer-scale) volcanic features as mapped over the entire near-side by Guest and Murray [1976], although it appears that the styles of eruption and nature of the dominant landform varies by location. The correspondence of volcanic landforms with topography not only encompasses such long-familiar mare volcanic “complexes” as Mons Rümker, the Marius Hills, and the Aristarchus plateau, but also includes some lesser known eruptive centers, such as Hortensiuss and Cauchy. Because our new interpretation of these areas is so radical, we here describe the geology of each volcanic center and its regional geological and topographic setting.

3.1. Marius Hills

This complex has long been known as a center of intense volcanic activity (Figure 2), displaying over 300 small cones and domes and numerous sinuous rilles and collapse pits [McCauley, 1967, 1968; Greeley, 1971; Guest, 1971; Weitz and Head, 1999; Heather et al. 2003]. The Marius Hills volcanic complex (Figures 3 and 4) occurs within Oceanus Procellarum, the most extensive maria on the Moon and the site of some of the youngest lunar lava flows [Schultz and Spudis, 1983; Hiesinger et al., 2003]. The cones and domes range in plan from a few kilometers to almost 20 km across and from 200 to over 600 m in height. Numerous sinuous rilles are found in the area, emanating from irregular or elongate source vents [McCauley, 1967, 1968; Greeley, 1971; Guest, 1971]. Pit craters and
Figure 1. Hemisphere view of topographic data from GLD100 [Scholten et al., 2012] for the near-side of the Moon, centered on 0°, 20°W showing location of proposed large lunar shield volcanoes. Color map has contour interval of ~250 m. Outlines of shield boundaries are approximate.

Table 1. Large Shield Volcanoes on the Moon*

<table>
<thead>
<tr>
<th>Shield</th>
<th>Summit</th>
<th>Age (Ga)</th>
<th>Diameter (km)</th>
<th>Height (km)</th>
<th>Average slope (°)</th>
<th>Volume (km³)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rümker</td>
<td>41°N, 59°W</td>
<td>&gt;3.4</td>
<td>66</td>
<td>1.2</td>
<td>2.4</td>
<td>1,400</td>
<td>Built on highland block</td>
</tr>
<tr>
<td>Gardner</td>
<td>16.1°N,</td>
<td>&gt;3.8</td>
<td>70</td>
<td>1.6</td>
<td>2.6</td>
<td>2,100</td>
<td>Similar to Rümker; on northern flank of Cauchy shield</td>
</tr>
<tr>
<td>Prinz</td>
<td>26°N, 43°W</td>
<td>3.4–3.6</td>
<td>166</td>
<td>0.8*</td>
<td>0.5</td>
<td>5,800</td>
<td>Built on highland block</td>
</tr>
<tr>
<td>Aristarchus</td>
<td>25.4°N, 50°W</td>
<td>~3.8</td>
<td>240</td>
<td>2.0*</td>
<td>0.9</td>
<td>30,100</td>
<td>Partly developed; built on highland block</td>
</tr>
<tr>
<td>Kepler</td>
<td>8°N, 38°W</td>
<td>2.1–3.6</td>
<td>270</td>
<td>0.6*</td>
<td>0.3</td>
<td>12,300</td>
<td>Very low slopes; few volcanic features</td>
</tr>
<tr>
<td>Hortensius</td>
<td>13°N, 29°W</td>
<td>3.1–3.5</td>
<td>300</td>
<td>1.2</td>
<td>0.4</td>
<td>28,300</td>
<td>Asymmetric; built on Montes Carpatus</td>
</tr>
<tr>
<td>Marius</td>
<td>14°N, 52°W</td>
<td>1.1–3.3</td>
<td>330</td>
<td>2.2</td>
<td>0.8</td>
<td>62,700</td>
<td>Fully developed shield</td>
</tr>
<tr>
<td>Hills</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cauchy</td>
<td>8°N, 35°E</td>
<td>3.6–3.7</td>
<td>560</td>
<td>1.8</td>
<td>0.1</td>
<td>148,000</td>
<td>Largest shield</td>
</tr>
</tbody>
</table>

*Age estimates taken from the literature (see text). Asterisk indicates that non-shield impact topography was deleted from estimate of edifice height. Measurements of diameter and height were made on the LROC-LOLA Digital Terrain Model GLD 100 [Scholten et al., 2012]. Volumes are computed using an approximation of a simple conical segment of radius (1/2D) and height shown.

Figure 2. The Marius Hills shield. At left, topographic image shows abrupt boundary at northern edge of shield (arrows). This boundary is clearly seen in the Kaguya high-definition television view (right, top and bottom) of the edge of the Marius Hills shield. Kaguya view is looking south while flying over about 18°N, 52°W. Field of view is about 200 km.
collapse features are common, including a recently discovered skylight within an apparent lava tube [Haruyama et al., 2009]. The cones and domes of the Marius Hills do not appear to be compositionally distinct from either the surrounding mare plains or the inter-volcano plains that make up the surface of the Marius Hills construct [Weitz and Head, 1999; Heather et al., 2003; Besse et al., 2011]. However, the decimeter-scale surface texture of the domes indicates that at least some of the constructs are rougher than the average mare surface, possibly indicating that clinkery aa lava, pasty eruptive spatter, and/or interbedded pyroclastics make up at least some of these edifices [Campbell et al., 2009; Lawrence et al., 2013].

[11] The Marius Hills complex occurs on an elongated, elliptical topographic rise approximately 330 km in extent. It is broadly shaped like a shield, with the summit near 14°N, 52°W, about 40 km northwest of the crater Marius, and it rises about 2.2 km above the surrounding mare plain (Table 1). Images from the orbiting Kaguya HDTV imager clearly show the shield-like morphology of the structure (Figure 2) and it is also evident in topographic profiles taken from the new global DTM (Figure 5). The cones, domes, and rilles that make up the volcanic complex are all superposed on the shield in a manner similar to the numerous late-stage cones and eruptive vents of the Mauna Kea shield on the island of Hawaii [MacDonald and Abbott, 1970]. On the basis of the broad, low-relief shape of this topographic bulge seen in Clementine topography, the Marius Hills were proposed to be the lunar equivalent of a basaltic shield volcano by Spudis [1996].

[12] The precise age of the Marius Hills construct is uncertain, but most workers agree that it is relatively young. It was mapped as Eratosthenian in age by McCauley [1967] and Wilhelms and McCauley [1971]. Whitford-Stark and Head [1980] mapped the lava flows of Oceanus Procellarum

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**Figure 3.** Volcanic features of the Marius Hills. Left, sinuous rilles and vent craters on western half of shield. Center, collapse pits (caldera?) near summit of Marius Hills shield. Right, Pit craters, sinuous rilles and low domes near eastern edge of Marius Hills shield.

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**Figure 4.** Volcanic features map of Marius Hills, simplified from McCauley [1968] and Guest and Murray [1976]. Symbols for volcanic landforms used here are the same as on all subsequent maps (landform definition and classification from Guest and Murray [1976]; see text for discussion).
and found that the surface flows around the Marius Hills included lavas from the uppermost sequence of flows, the Sharp and Hermann Formations; more recent studies have mapped these flows on the surface of the Marius Hills shield and have estimated ages of 2.5 and 3.3 Ga for the two principal flow series [Heather and Dunkin, 2002; Heather et al., 2003]. Most recently, Huang et al. [2011] propose very young ages for the surface lavas of the Marius Hills, ranging from 0.8 to 1.1 Ga.

The free-air gravity anomaly at the Marius Hills is about 200–250 km across [Konopliv et al., 2001] and requires the presence of a significant volume of dense subsurface material, which likely takes the form of either a laccolithic intrusion or the filling by basalt of the impact-induced pore space in the uppermost crust. Kiefer [2013] models this gravity feature as being produced by two dense, subsurface bodies: a northern one 160–180 km across, corresponding to the Marius Hills bulge (including most of the domes and cones within its boundary) and a smaller one 100–140 km in extent, south of the main part of the Marius Hills shield. The two structures are connected by a narrower line of dense material. The geophysical evidence of a large, dense subsurface feature centered under the topographic bulge is consistent with the Marius Hills being made up of a single subsurface magmatic system and supports our interpretation of it as a large basaltic shield volcano. If the anomaly is caused by dispersed intrusive in the crustal pore space rather than being concentrated in a classical magma chamber, this would inhibit the crustal collapse needed to form a large, significant summit caldera (although a chain of collapse features (center image, Figure 3) near the summit is evident). Because impact-induced porosity is probably widespread in the upper crust across the Moon, this may provide an explanation for the general absence of calderas on the lunar volcanic complexes described here.

The predominant volcanic landform of the Marius Hills is the relatively steep-sided cone (Figure 4) [see also Figure 5. Topographic profiles of the Marius Hills shield; all topographic information for the features described in this paper is taken from the LROC stereo DTM GLD 100 [Scholten et al., 2012]. The typical broad shield profile is evident, showing a feature about 330 km across and 2.2 km high.

Figure 5. Topographic profiles of the Marius Hills shield; all topographic information for the features described in this paper is taken from the LROC stereo DTM GLD 100 [Scholten et al., 2012]. The typical broad shield profile is evident, showing a feature about 330 km across and 2.2 km high.

Figure 6. Rümker, a volcanic shield in northern Oceanus Procellarum. This relatively small feature consists of a broad shield and several overlapping low shields. WAC view of Rümker structure (left); low-sun view of the overlap of shields of the Rümker shield (right).
McCauley, 1968; Guest and Murray, 1976], most of which are found at the summits of broad, low shields. Both collapse pits and sinuous rilles are also common [e.g., Greeley, 1971] as are larger collapse pits near the summit (Figure 3). Small shield-like volcanoes without capping cones are much less common here, although they are abundant at some of the other complexes we discuss (see below). The dominance of certain landforms and the paucity of others at different complexes probably indicate differing styles of the predominant eruption and evolutionary paths of the various complexes [Whitford-Stark and Head, 1977]. The near-exclusive presence of cones at Marius Hills suggests a protracted volcanic evolution, with the eruption of many, relatively volatile-rich, partly crystallized magmas (resulting in the production of abundant spatter, degassing, and clinkery aa lava flows) [Weitz and Head, 1999; Heather et al., 2003].

3.2. Rümker

[15] The Rümker complex (Figure 6) in northern Oceanus Procellarum (~70 km in extent, centered at 40°N, 58°W) was recognized as a volcanic center early in lunar geological studies [McCauley, 1968; Guest, 1971; Scott and Eggleton, 1973; Smith, 1973, 1974]. It consists of a broadly elevated cluster of more than a dozen (up to 30 according to Smith [1974]) blister-like landforms (Figure 7), built on top of a kipuka of Imbrium basin ejecta, the Fra Mauro Formation [Guest, 1971; Scott and Eggleton, 1973]. The thin mare

Figure 7. Volcanic features map of Rümker. Low shields, probably of basaltic composition, predominate.

Figure 8. Topographic profiles of the Rümker shield, showing a construct 66 km across and about 1.2 km high.
flows of northern Procellarum lap up and partly cover the lower portions of the edifice, suggesting that the complex pre-dates the ~3.4 Ga old surface mare basalts in this region; very young mare basalts (<1.5 Ga old) lap over the complex to the northeast [Hiesinger et al., 2003]. In profile, Rümker displays the bulbous shape of an elongate shield (Figure 8), with a typical relief of 1000–1200 m above the surrounding mare surface.

In some ways, Mons Rümker appears to be a miniature version of the Marius Hills shield, but much smaller and less well developed [Smith, 1974]. However, here at Rümker, the low shield is the dominant landform (Figure 7). Some of the overlapping shield constructs may be thinly mantled uplands, but the characteristic shield shape of these features argues instead that they are small, central volcanoes, similar to basaltic shields found elsewhere on the Moon and the other terrestrial planets [e.g., Greeley, 1976, 1982]. A large crater at the north end of the complex may be a collapse pit (Figure 6). There is no evidence for sinuous rilles or other vent structures, although such features could be covered by the younger mare basalts of the surrounding plain.

3.3. Prinz

The Prinz volcanic complex (Figures 9 and 10; ~150 km in extent; 26°N 43°W) is built upon a block of highlands material (Montes Harbinger) that is probably related to the Imbrium basin [Strain and El-Baz, 1977]. Unlike Rümker (but similar to its neighbor, Aristarchus), the Prinz complex is notable as the source region for several sinuous rilles (lava channels) that supply the mare deposits north and west of the plateau (Figure 10). Some of the dome-like features in the Prinz area could be volcanic constructs, particularly one that appears to be a breached

**Figure 9.** Volcanic vents and channels of the Prinz shield (~150 km in extent; 26°N 43°W). Collapse pits and rilles near rim of crater Prinz (left); caldera-like pit and vent area near middle of structure (right).

**Figure 10.** Volcanic features map of Prinz shield. Pit craters and sinuous rilles are the dominant landform.
cone that might have served as a lava source (arrow in Figure 9) [see also “E” in Figure 1 of Strain and El Baz, 1977]. The Prinz complex as a whole displays relief of about 600 m (Figure 11); this low value, in conjunction with the exposure of many highlands units near and within the construct as well as the principal manifestation of rilles as the main landform here, suggests incomplete development as a volcanic complex. Prinz appears to have been a significant eruptive center, but most of its lava products were supplied to the surrounding mare plains of Oceanus Procellarum and Mare Imbrium. Although Prinz is adjacent to the Aristarchus plateau, there is no obvious direct connection between the two complexes; each features similar landforms, but the relative importance of the various types differ. The basalts of the Prinz shield appear to have ages between 3.4 and 3.6 Ga [Strain and El-Baz, 1977; Zisk et al., 1977], indicating an Imbrian age for the construct.

3.4. Kepler

[18] The Kepler volcanic complex (~270 km; 7°N, 38°W) is newly identified in this work, although its presence peripheral to the Imbrium basin is predicted by the work of McGovern and Litherland [2011]. It consists of a very low relief topographic rise near and south of the younger superposed, unrelated impact crater Kepler. Highland materials in this area are facies of Imbrium basin ejecta, primarily the knobby Alpes Formation [Wilhelms and McCauley, 1971]. This region is not commonly thought of as a mare volcanic complex, but close examination reveals that sinuous rilles, irregular volcanic craters and associated dark mantle (pyroclastic) materials occur throughout the area (Figures 12 and 13). Mare lavas on the Kepler shield have not been dated directly, but two mare units 3.6 and 2.1 Ga make up part of the western and southern edges of the shield [Hiesinger et al., 2003]. The Kepler feature is similar in developmental state to the Prinz structure described above; it is built on top of highlands material, mostly ejecta from the Imbrium basin, which is thinly covered by a veneer of basaltic lava and pyroclastic deposits. Although relatively low in overall relief (~0.6 km in height; Figure 14), this rise is too tall and wide to be attributed exclusively to impact causes, either from the crater Kepler or the regional Imbrium basin back slope topography and thus we classify it as a shield (Table 1). Rough topography near the summit of the Kepler shield (partially obscured by impact ejecta from the crater Kepler) could be remnants of additional igneous activity.

Figure 11. Topographic profiles of the Prinz shield, about 160 km across and slightly less than 1 km high.

Figure 12. Volcanic features of the Kepler shield (~270 km; 7°N, 38°W). Collapse pits and rilles (arrow; left) near Maestlin R; elongate vent and associated dark pyroclastics (arrow) NE of Encke (center); sinuous rille complex (arrow) on eastern shield edge (right).
3.5. Hortensius

[19] The Hortensius-Tobias Mayer area (approximately 150 by 350 km; 12°N, 27°W; Figures 15 and 16) has been long known for its high concentration of small volcanic landforms, including small shields, rilles, and cones [Shoemaker, 1962; Schmitt et al., 1967; Smith, 1973; Schultz, 1976; Phillips, 1989; Wood, 2003; Wöhler et al., 2006]. The northern edge of this structure contains a series of vents and pyroclastic deposits that are associated with the eruption of the famous late Imbrium flows, the long, striking lobate lava flows that cover Mare Imbrium [Schaber, 1973; Schaber et al., 1976]. This shield is similar to both the Prinz and Kepler shields in that altimetry data show that it is built upon the main ring/rim of the Imbrium basin, but both volcanic development and the diversity of landforms at Hortensius are much greater than in either Prinz or Kepler. Imbrium ejecta crop out on the surface as knobby Alpes Fm. [Wilhelms and McCauley, 1971] along with the occasional basin massif, but volcanic shields and structures are abundant and clearly distinguishable from the highlands units upon which they are built (Figure 15). The small shields of Hortensius are well-developed miniature volcanoes [Schultz, 1976; Head and Gifford, 1980; Wöhler et al., 2006], similar to basaltic shields that are found in

Figure 13. Volcanic features map of Kepler shield. Pit craters, sinuous rilles, and pyroclastic vents are the principal features of this shield.

Figure 14. Topographic profiles of the Kepler shield. Feature is about 270 km across and 500–600 m high.
“plains volcanic terrains” elsewhere on the Moon and on other terrestrial planets [Greeley, 1976; 1982]. Linear vents, lines of spatter cones, and discontinuous pyroclastic deposits (Figures 15 and 16) [see also Gustafson et al., 2012] are also abundant, particularly at the northern edge of the shield, the probable source vents for the spectacular late Imbrium lava flows [Schaber, 1973; Schaber et al., 1976].

[20] The Hortensius shield is one of the largest identified in this study, extending almost 350 km along its NW/SE axis (Figure 16). Topographic profiles show that the shield is about 1200 m or less in total height (Figure 17 and Table 1), making it a very low relief structure. Clusters of small shields are found mostly along the margins of the shield, particularly in the southwest (Figure 16). The eastern margin of the shield is partly masked by ejecta from the crater Copernicus. Mare basalts on the western and southern edge of the shield have estimated ages of 3.5 and 3.1 Ga, respectively [Hiesinger et al., 2003].

3.6. Cauchy

[21] The largest of the newly detected volcanic shields is in eastern Mare Tranquillitatis, centered near the crater Cauchy (~560 km; 8°N, 35°E; Figures 18 and 19). This area has long been known as a locus of small volcanic features, including numerous cones, low shields, and sinuous rilles [e.g., Wilhelms, 1972; Guest and Murray, 1976; Wood, 2003; Wöhler et al., 2006]. Topographic data reveal that the eastern half of Mare Tranquillitatis is a broad, low rise about 560 km across and over 1.8 km high (Figure 20). This shield has been the source of multiple flows and eruptive events emplaced between 3.6 and 3.7 Ga [Hiesinger et al., 2000; Rajmon and Spudis, 2004]. The existence of this volcanic shield may explain the apparent lack of topographic evidence for the putative Tranquillitatis impact basin, for which there is clear morphological evidence [Wilhelms and McCauley, 1971; Wilhelms, 1987]—the Cauchy shield is built on the floor of the Tranquillitatis basin, creating a topographic high in its eastern half, where a low would be expected.

[22] The Cauchy shield displays most of the volcanic landforms seen in other lunar shields. The predominant feature is a low shield, usually with a summit pit (Figure 18). The unusual Rimae Cauchy I and II appear to be combinations of linear graben and sinuous rilles in different portions of the features. Rima Cauchy I terminates in a couple of collapse pits (Figure 18), apparently the source vents for the basalts that created the sinuous rille parts of the feature (Figure 19). The Gardner “megadome” shield (described below) occurs on the northern margin of the Cauchy shield, but as with Prinz and Aristarchus, there appears to be no direct genetic connection between the two features.
3.7. Gardner

At the northern end of the Cauchy structure is a smaller feature, identified as the Gardner “megadome” by Wood [2003, 2004]. This small (~70 km; 16°N, 34°E; Figures 21 and 22) topographic blister displays several smaller, overlapping shields and sinuous rilles. Its summit displays a series of irregular depressions that may constitute a caldera complex. A similar sequence of collapse pits are found near the summit of the Marius Hills shield [Guest, 1971; Figure 3], but in this case, the collapse structure covers most of the summit of Gardner (Figure 22). The surface composition of Gardner seems to be less mafic than the surrounding lavas of Mare Tranquillitatis, with ~14–16 wt.% FeO content, values that while broadly basaltic are lower than typical mare material (e.g., the surrounding Mare Tranquillitatis basalts have FeO content of ~20 wt. %). The low FeO content of the surface of Gardner could indicate either the eruption of a less mafic type of volcanic magma (e.g., high-alumina mare basalts) [BVSP, 1981] or that the surface materials are at least partly of highlands

Figure 16. Map of volcanic features of Hortensius shield. After Guest and Murray [1976].

Figure 17. Topographic profile of Hortensius shield showing feature about 300 km across and over 1 km high.
composition and the Gardner shield is the surface expression of predominantly intrusive activity, such as a laccolith [e.g., see Wöhler et al., 2006]. The Gardner shield is relatively small, but shows relief of about 1.6 km (Figure 23), thus, as at Rümker, it has higher than typical average slope compared to other large shield volcanoes. The age of the feature is undetermined, but both its heavily cratered appearance and the apparent superposition of basin radial texture (Figure 21) suggest that it is old, perhaps older than 3.8 Ga.

[24] Based in part on its alignment with the volcanic area near the crater Jansen, Wood et al. [2005] proposed that Gardner is the northern terminus of an elongate quasi-linear, volcano-tectonic structure. We suggest instead that Jansen is part of the Cauchy shield (Figure 19) and that Gardner is a possible parasitic shield of Cauchy, located on its periphery (Figure 1). In terms of size, morphology, and distribution of features, the Gardner structure strongly resembles the Rümker shield, which in turn is a miniature version of the Marius Hills shield. Surface features of the Gardner shield are mostly smooth, overlapping shield-like domes, some of which have collapse pits (Figure 22). Thus, there appears to be a continuous sequence of size in lunar shield volcanoes over at least an order of magnitude.

3.8. Aristarchus

[25] Among the lunar volcanic structures described here, the Aristarchus plateau (~250 km; 25°N, 50°W; Figures 24, 25) seems to be a special case. For the plateau, topographic prominence is caused principally by the uplifting of a structural block associated with the formation of the Imbrium basin and is enhanced only partly by the overplating of erupted lava [Moore, 1967; Zisk et al., 1977; McEwen et al., 1994]. The highland block that makes up the bulk of the plateau shows
clear control by radial structures of the Imbrium basin and ejecta from that event is exposed in the rugged terra of the plateau (Figure 25). However, massive eruption of lava both onto and away from the plateau is indicated by the presence of many large sinuous rilles, including the enormous Vallis Schröteri (165 km long) and numerous other rilles (Figure 26). Some rilles show clear evidence of at least two-phases of eruption (e.g., the highly sinuous rille within the broader, graben-like rille of V. Schröteri; Figure 24), suggesting a prolonged, multi-phased volcanic evolution [e.g., Moore, 1967; Schultz, 1976]. The Aristarchus plateau is also the source of dark red, 49,000 km² regional pyroclastic deposits, five times larger in areal extent than any other on the Moon [Gaddis et al., 2003]. These eruptions of both lava and ash have partly covered the pre-existing plateau but seem to have ended before a significant shield-shaped construct could be built. Zisk et al. [1977] estimate the age of plateau materials to be Orientale-contemporaneous (about 3.8 Ga) but very young lavas are found south of the shield (1–1.5 Ga) [Hiesinger et al., 2003] and at least some of these flows originated from vents on the Aristarchus shield itself (Figure 26).

The dominant landform on the Aristarchus shield is similar to the neighboring Prinz shield, collapse craters, and sinuous rilles (Figure 26), along with the substantial pyroclastic deposits mentioned above. Some of the smaller hills might be volcanic constructs, but the majority appears to be outcrops of the underlying highlands block upon which the plateau lavas have been erupted [Zisk et al., 1977]. The very young mare basalts that lap up onto the plateau [e.g., Hiesinger et al., 2003] partly obscure the relations of the rille termini around the eruptive center. These relations suggest that the Aristarchus plateau is a proto-shield, currently exposed in an arrested state of development whereas the Marius Hills construct (Figure 2) is a fully developed lunar shield volcano. Nevertheless, topographic profiles of the Aristarchus shield (Figure 27) show a broad, shield-like shape, 240 km across and up to 2 km high (Table 1). The blister-like morphology of the Aristarchus shield is also evident in the low sun angle mosaic (Figure 25).

Many of the lunar shields have long been recognized as volcanic complexes [McCauley, 1967; Guest, 1971; Guest and Murray, 1976] or eruptive centers [Whitford-Stark and Head, 1977], but their topographic nature (Figure 1 and Table 1) has been only mentioned in passing or has not been known. Both Marius Hills and Rümker were long known to occur on topographic highs and data from the Apollo metric camera demonstrated that the Prinz complex is associated with a fragment of the rim massifs of the Imbrium basin (Montes Harbinger) [Strain and El-Baz, 1977]. Most workers consider the volcanic activity of these eruptive centers to have been minor adjuncts of the main phase of mare volcanism, which was largely characterized by voluminous, fissure-fed, flat-lying eruptions of flood
lavas [e.g., Wilhelms, 1987]. However, Whitford-Stark and Head [1977] suggested that much of the basalt of Oceanus Procellarum may have been emplaced as eruptives from a few volcanic centers, including both the Marius Hills and Aristarchus structures. We concur with this latter interpretation at least in part, as rilles (lava channels) originating on the lunar shields have clearly supplied lava to the surrounding maria.

4. Distribution and Morphometry of Large Lunar Shields

The distribution of these proposed lunar shield volcanoes is decidedly non-random (Figure 1). The largest grouping (Marius, Prinz, Hortensius, Rümker, Aristarchus, and Kepler) is distributed along the southern and western periphery of the Imbrium basin, within the large Procellarum-KREEP terrane [Haskin et al., 2000], with its anomalously high-Th content [Lawrence et al., 2007], while the largest shield (Cauchy) sits in the midst of a cluster of mare basins, including Serenitatis, Nectaris, and Crisium. These shields all lie within predicted annular zones of enhanced magma ascent produced by stresses from loading of mare units inside these basins [Litherland and McGovern, 2009; McGovern and Litherland, 2011]. The mechanism for enhancing ascent of magma stems from a combination of the lithosphere’s flexural and membrane responses to initial basin-filling mare loads that creates favorable principal

Figure 22. Volcanic features map of the Gardner shield. The predominant landform expression is a series of interfingered low shields and the collapse pit at its summit.

Figure 23. Topographic profiles of the Gardner shield.
stress orientations [Anderson, 1936] and tectonic stress gra
dients [Rubin, 1995] for vertical transport of magma in dikes
with orientations radial to the basins.

[29] The sizes of these lunar shield volcanoes are compa-
rable to other broad, low-relief basaltic shields on the Earth,
Mars, and Venus (Figures 28 and 29). For example, the Kali
Mons volcano on Venus has diameter and flank slope distrib-
utions similar to those of the proposed Marius Hills shield
(Figure 29). Typical slopes for the flanks of these shields are
extremely low, on the order of less than 1° (Table 1). How-
ever, the slopes of the two smallest features, Rümker and
Gardner, are about 2.5°; these features also have unique
populations of small overlapping volcanic domes and low
shields. These relations may indicate slightly different pro-
cesses at work in the construction of these smaller shields.
The estimated volumes of the large lunar shields range from
about $10^{12}$ to $10^{14}$ m$^3$ (Table 1). These values fall among
the lower range of estimates of volumes for shield volcanoes on
Mars, which range from $10^{12}$ to more than $10^{15}$ m$^3$ [Plescia,
2004]. It is evident that the size, shape, and volume of these
lunar features (Figures 28 and 29 and Table 1) are compara-
ble in morphometry to unequivocal basaltic shield volcanoes
on other terrestrial planets, including Earth [Pike, 1978],
Mars [Plescia, 2004], and Venus [Herrick et al., 2005].

[30] With the possible exceptions of Gardner and Marius
Hills, most of the newly described lunar shields do not have
summit pits or calderas. However, some shields on the other
terrestrial planets likewise do not have summit craters and
the absence of such does not negate the classification of ei-
ther these or the lunar features as shield volcanoes. Many
shield volcanoes on Venus do not display summit craters
but the existence of these volcanoes is evident by concentra-
tions of eruptive landforms, radiating flows, and broad, low
relief topographic swells. The absence of a caldera is consist-
ent with the mode of mantle-to-surface magma transport
predicted by the models of McGovern and Litherland
[2011]: dikes are aligned with regional (basin-loading)
stresses. Evidence for this is particularly strong at the
Cauchy shield, which is topped by a set of large graben/lava
fissures (Rimae Cauchy I and II) that are radial to the Seren-
itatis basin.

[31] Scenarios for the development of the proposed
Cauchy shield can be constrained by remote sensing and
gеological mapping. Cauchy lacks the strong free-air gravity
high characteristic of flexurally supported shield volcanoes
such as Marius Hills [Kiefer, 2013], but joint analysis of
gravity and topography indicates a local crustal thickness
maximum beneath the structure [Neumann et al., 1996],
consistent with an isostatically compensated rise of material
with the density of crustal rocks. Mapping and interpretation
of multispectral imaging data indicate that the surface basalts
in eastern Mare Tranquilitatis are on the order of several
hundred meters in thickness [Rajmon and Spudis,
2004]. This value is a fraction of the observed 1.8 km relief of
the Cauchy shield. Perhaps the basalts of the Cauchy shield
comprise a thin carapace covering a pre-existing crustal
block (similar to the developmental scenario proposed above
for Aristarchus). Alternatively, the topographic rise could
have been built up by substantial amounts of intrusion and
underplating of moderate-density magmas, augmented by
late-stage eruptions of denser cumulate-rich magmas aided
by favorable stresses in the lithosphere [McGovern and
Litherland, 2011]. The former case would be inconsistent

Figure 24. Volcanic features of the Aristarchus shield (~250 km; 25°N, 50°W). Pit craters and sinuous
rilles of the northern margin of the plateau (left); Cobra Head pit crater source vent and Vallis Schröteri, a
nested sinuous rille indicating multi-phase eruption history (center); collapse pits and sinuous rilles of the
NE section of the plateau (right).

Figure 25. The LROC WAC mosaic at low sun illumination of the Aristarchus region, showing bulbous, blister-like
shape of the topography of the plateau. Such morphology is consistent with the interpretation of the Aristarchus plateau
with the postulated presence of a Tranquillitatis impact basin [e.g., Wilhelms, 1987], while the latter allows it.

The recognition of the existence of large shield volcanoes on the Moon invites a re-examination of scenarios for lunar magma ascent. It has been previously held that the apparent absence of shield volcanoes on the Moon was consistent with buoyancy-controlled ascent of magmas through the lunar crust [e.g., Head and Wilson, 1991]. In this view, the presumed negative buoyancy of basaltic magmas in an anorthositic crust precludes the creation of shallow reservoirs at neutral buoyancy horizons, from which relatively low- and moderate-volume shield-building eruptions could emanate. Instead, high-volume eruptions from dikes long enough to benefit from positive buoyancy deep in the mantle were required; these would produce sheets of flood lavas rather than central-vent edifices [Head and Wilson, 1991]. Although there is reason to question the negative buoyancy of mare basalts magmas [e.g., Wieczorek et al., 2001], our observations that large shields do exist suggest that lunar eruptions probably spanned a range of volumes and mass eruption rates, allowing both shield-building and flood-type eruptions. The shield-building eruptions could come directly from the mantle, driven by basin loading-induced stress in dikes [McGovern and Litherland, 2011], from shallow magma bodies created by intrusion-trapping loading stresses [e.g., Solomon and Head, 1980; Galgana et al., 2011] or filling of pore space in an impact-processed and fractured upper crust [Kiefer, 2013].

Figure 26. Map of the volcanic features of the Aristarchus shield (~250 km; 25°N, 50°W). The principal landform is the sinuous rille and pit crater, many of which have emplaced the surrounding mare deposits. Abundant pyroclastic materials discontinuously cover the plateau lavas, making determination of the ages of surface units uncertain.
Figure 28. Height-diameter relations for shield volcanoes on several terrestrial planets; plotted “trends” are informal and diagrammatic. The newly described lunar structures (stars) fall within the ranges of sizes and heights of large, low-relief basaltic shields on Venus and Mars. Data for Earth (green squares) [Pike, 1978], Mars (red triangles) [Plescia, 2004], and Venus (blue circles) [Herrick et al., 2005].

Figure 29. Cross-sections of topography (solid black line, left y-axis) and slope (red boxes, right y-axis) for proposed lunar shield volcano Marius Hills (left, from LOLA) [Smith et al., 2010] and for the Kali Mons edifice on Venus (right, from Magellan) [Ford and Pettengill, 1992]. Azimuths from feature center are indicated above each profile. Slope measurements are calculated using a least squares fit to a plane over an 80 km wide baseline centered on each point.
in magma supply rate may also be important. Regional variability in magma supply on the Moon is likely but cannot be well constrained observationally. Assuming such supply rate variations do occur, this could contribute to the scatter in h/d observed for the various lunar shields. Temporal variability in magma supply may also be important. For example, volcanic flows from the Marius Hills have been mapped as sources for lava flows in the surrounding regions of Oceanus Procellarum [Whitford-Stark and Head, 1980] in addition to producing the Marius Hills topographic shield. Periods of relatively high magma flux would result in thin, widely distributed lava flows in Oceanus Procellarum, while periods of lower flux would result in a thicker lava sequence and a more localized production of lava in the Marius Hills shield.

[36] A new appreciation of these lunar volcanic complexes as shield volcanoes does not alter our general picture of mare volcanism on the Moon as being dominated by the eruption of high-volume flood lavas. Sheets of lava erupted from linear fissures are still the best explanation for most of the infilling of the mare basins to varying degrees. However, the identification of large shield volcanoes indicates that some significant fraction of erupted magma on the Moon has been emplaced via the mechanism of low volume, prolonged shield-building eruptions. The ages of the mare surfaces associated with these features [Hiesinger et al., 2000; 2003] range over several hundred million years during the era of mare volcanism (Table 1), from as old as 3.8 Ga for the lavas of the Aristarchus shield [Zisk et al., 1977] to flows possibly as young as ~1 Ga on the Marius Hills shield [Huang et al., 2011]. Thus, shield building was a continuous process ‘during the principal epoch of mare volcanism on the Moon (3.9–3.0 Ga) [BVSP, 1981; Wilhelms, 1987] and possibly extending well beyond it, up to as recently as 1 Ga ago [Huang et al., 2011].

5. Conclusions

[37] We correlate large topographic prominences in the lunar maria with concentrations of small volcanic features such as domes, pit craters, small shields, cones, and rilles. We interpret these large, broad topographic features as shield volcanoes, a previously unrecognized style of lunar volcanic activity. Shield building occurred during the main phases of mare volcanism on the Moon, between 3.9 and 3.0 Ga ago. The lunar shields show a variety of development states, ranging from nearly complete shield development (e.g., Marius Hills) to proto-shield, highland block volcanic resurfacing (e.g., Aristarchus). The features studied in this work are comparable in size and shape to basaltic shield volcanoes on other terrestrial planets, supporting our interpretation that they too are volcanic shields. Shields are found proximate to the large, lava-filled impact basins Imbrium and Serenitatis and structural features associated with the shields tend to be radial to these basins. The recognition of shield volcanoes on the Moon affirms that this style of volcanism has been ubiquitous on the terrestrial planets of our Solar System.

References
