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The "Procellarum KREEP Terrane": Implications for mare volcanism and lunar evolution

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Abstract. Geophysical, remote-sensing, and sample data demonstrate that the Procellarum and Imbrium regions of the Moon make up a unique geochemical crustal province (here dubbed the Procellarum KREEP Terrane). Geochemical studies of Imbrium's ejecta and the crustal structure of the Imbrium and Serenitatis basins both suggest that a large portion of the lunar crust in this locale is composed of a material similar in composition to Apollo 15 KREEP basalt. KREEP basalt has about 300 times more uranium and thorium than chondrites, so this implies that a large portion of Moon's heat-producing elements is located within this single crustal province. The spatial distribution of mare volcanism closely parallels the confines of the Procellarum KREEP Terrane and this suggests a causal relationship between the two phenomena. We have modeled the Moon's thermal evolution using a simple thermal conduction model and show that as a result of the high abundance of heat-producing elements that are found in the Procellarum KREEP Terrane, partial melting of the underlying mantle is an inevitable outcome. Specifically, by placing a 10-km KREEP basalt layer at the base of the crust there, our model predicts that mare volcanism should span most of the Moon's history and that the depth of melting should increase with time to a maximum depth of about 600 km. We suggest that the 500-km seismic discontinuity that is observed in the Apollo seismic data may represent this maximum depth of melting. Our model also predicts that the KREEP basalt layer should remain partially molten for a few billion years. Thus the Imbrium impact event most likely excavated into a partially molten KREEP basalt magma chamber. We postulate that the KREEP basalt composition is a by-product of mixing urKREEP with shallow partial melts of the underlying mantle. Since Mg-suite rocks are likely derived from crystallizing KREEP basalt, the provenance of these plutonic rocks is likely to be unique to this region of the Moon.

1. Introduction

The scientific objectives of the Apollo missions were mainly defined by an analysis of Earth-based telescopic observations and photographs that were taken from the Lunar Orbiter spacecraft [e.g., *Wilhelms*, 1993]. With the subsequent Apollo landings our primarily photogeologic knowledge of the Moon was supplemented by the return of surface samples, the deployment of surface geophysical instruments, and the collection of data from orbit. Though the lunar landings increased our knowledge of the Moon's geologic history by many orders of magnitude, the landings provided us with only a limited number of "ground truths" to test theories of lunar origin and its magmatic evolution. Regional observations of the surface chemistry, magnetic field, gravity, and topography were obtained from orbit, but most of these measurements were limited to two equatorial swaths across the lunar surface. The Moon's geologic evolution at the end of the Apollo era thus had to be inferred without having a good knowledge of its global properties.

The analysis of the returned samples quickly revealed that the Moon's evolution was drastically different from that of the Earth. Many samples were found to contain highly elevated concentrations of incompatible elements, and these rocks were given the name "KREEP" because of their high levels of potassium (K), rare earth elements (REE), and phosphorous (P). Though the absolute concentrations of the elements that make up KREEP varied among these samples, their relative concentrations were found to be roughly constant [e.g., *Warren and Wasson*, 1979]. This observation, as well as the revelation that the lunar highlands crust is largely anorthositic in composition [*Wood et al.*, 1970], quickly led to the paradigm that the Moon's crust is a differentiation product of a large "magma ocean" (for a review, see *Warren* [1985]). In this model the last remaining dregs to crystallize from the magma ocean are highly enriched in incompatible elements and are initially sandwiched between the flotation anorthositic crust and mafic mantle cumulates. It was taken as granted by many researchers that this KREEP layer was global in extent.

Observations from the Apollo data hinted that a dichotomy in geologic processes may have existed between the lunar nearside and farside [e.g., *Metzger et al.*, 1973; *Evensen et al.*, 1974; *Wasson and Warren*, 1980; *Shervais and Taylor*, 1986; *Warren and Rasmussen*, 1987]. The Apollo 15 and 16 gamma-ray spectrometers [*Metzger et al.*, 1973; *Bielefeld et al.*, 1976; *Metzger et al.*, 1977; *Metzger*, 1993] proved that the surface distribution of thorium and potassium, and by inference KREEP, could be mapped from orbit. For the equatorial swaths in which gamma-ray data were obtained, these results showed that Oceanus Procellarum and Mare Imbrium (the "western maria")

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were highly enriched in KREEP when compared to the farside highlands and eastern maria. Topographic data obtained from orbit showed that there was a 2-km displacement between the Moon's center of mass and center of figure roughly along the Earth-Moon axis [*Kaula et al.*, 1972], and this suggested that the farside crust was thicker than that of the nearside. It had long been known that the mare basalts erupted preferentially on the lunar nearside, and this observation was generally attributed to the nearside crust being thinner than that of the farside [e.g., *Solomon*, 1975; *Head and Wilson*, 1992].

Though these early data were certainly suggestive, the full extent of the nearside-farside dichotomy would have to wait for data obtained from later polar-orbiting spacecraft. Near-global topography and gravity coverage obtained from the Clementine mission [*Nozette et al.*, 1994] confirmed that the shape of the Moon did indeed possess a nearside-farside dichotomy [*Zuber et al.*, 1994; *Smith et al.*, 1997]. The full topographic expression of the giant South Pole-Aitken impact basin was also determined, and geophysical modeling has since inferred that the crust beneath this basin's floor was significantly thinned as a result of the impact process [*Neumann et al.*, 1996; *Wieczorek and Phillips*, 1998]. If crustal thickness was the only factor controlling the eruption of mare basalts then this basin should certainly have contained far more volcanic flows than are recognized there (for example, see the geologic mapping of *Wilhelms* [1987]). Thus the asymmetric distribution of mare basalts likely depends on other factors besides crustal thickness, such as lateral variations in heat flow and magma production rates [e.g., *Lucey et al.*, 1994; *Smith et al.*, 1997].

The Lunar Prospector mission [*Binder*, 1998] recently obtained global gamma-ray spectra from which long-awaited global elemental concentrations are currently being extracted. These data dramatically show that the surface distribution of incompatible elements is highly concentrated only in the Procellarum and Imbrium regions of the Moon [*Lawrence et al.*, 1998; *Elphic et al.*, 1999]. A few recent studies had come to nearly the same conclusion as we discuss in more detail below. For instance, *Haskin* [1998] assumed that the crust beneath much of Oceanus Procellarum and Mare Imbrium had enhanced concentrations of KREEP. He postulated that the Imbrium impact excavated into this incompatible element-rich province and showed that the surface distribution of thorium was consistent with having an origin as ejecta from this basin. In another study, *Wieczorek and Phillips* [1999] showed that only basins that formed within the Procellarum and Imbrium regions of the Moon had undergone significant postimpact structural modification. They attributed this observation to higher crustal temperatures there and/or voluminous KREEP basalt volcanism (KREEP basalt is distinct from the mare basalts in both composition and density).

In this paper we first argue (as do *Jolliff et al.* [this issue], *Korotev* [this issue], and *Haskin et al.* [this issue]) that the Procellarum and Imbrium region of the Moon is a unique geochemical crustal province. We show that the remote-sensing, sample, and geophysical data all support the conclusion that the bulk of the crust in this region is highly enhanced in heat-producing and incompatible elements. Specifically, we argue (as does *Korotev* [this issue]) that a large portion of the crust in this region has a composition similar to that of Apollo 15 KREEP basalt (or the intrusive or differentiated equivalent). On the basis of the unique properties of this region of the Moon we dub this province the "Procellarum KREEP Terrane," or PKT for short.

Using a simple thermal conduction model, we next show that the dramatic enhancement of heat-producing elements within the Procellarum KREEP Terrane has had a significant influence on the thermal and magmatic evolution of this region. We note that more than 60% of mare volcanism by area is found to occur within the confines of the PKT, and we show that this is likely to be a natural consequence of the high concentrations of heat-producing elements that are found there. By placing a 10-km layer of KREEP basalt at the base of the crust in this region, our model results show that this material would remain molten for a few billion years and that it would additionally heat and partially melt the underlying mantle. Partial melting of the mantle is found to occur over most of the Moon's history, and the maximum depth of melting is shown to increase with time to a depth of about 600 km. This model result is consistent with the long duration of mare volcanism (from at least 4.2 Ga [*Taylor et al.*, 1983] to about 900 Ma [*Schultz and Spudis*, 1983]) as well as the depth of origin of mare basalts (<540 km [e.g., *Longhi*, 1992]).

We believe that the first-order results of our thermal model offer a framework in which to comprehend a wide range of hitherto unrelated lunar observations: (1) The fact that the maria are primarily located on the lunar nearside is a natural consequence of the high abundance of heat-producing elements in the crust of the Procellarum KREEP Terrane. (2) The fact that the measured heat flow is larger at the Apollo 15 site than at the Apollo 17 site is a direct consequence of this site being closer to the center of the PKT [e.g., *Langseth et al.*, 1976; *Warren and Rasmussen*, 1987]. (3) The fact that the Apollo 15 KREEP basalts are indistinguishable in absolute age from the Imbrium impact is consistent with the Imbrium bolide impacting a molten KREEP basalt magma chamber [*Ryder*, 1994]. (4) The reason that an igneous protolith to the Low-K Fra Mauro ("LKFM") mafic impact-melt breccias has never been found is that this impact melt is a mixture, of which one component was initially molten [*Spudis et al.*, 1991]. (5) The reason that the Mg-suite rocks are both KREEP and magnesian rich is that urKREEP melted and mixed with the underlying magnesian-rich magma ocean cumulates. (6) The large range of crystallization ages of the Mg- and alkali-suite rocks is due to the slow crystallization of a subcrustal KREEP basalt magma chamber [e.g., *Snyder et al.*, 1995a, b]. (7) The 500-km seismic discontinuity observed beneath the nearside crust is consistent with representing the maximum depth of melting of the mare source.

In the following section we discuss the evidence that demonstrates the Procellarum and Imbrium regions of the Moon are part of a unique KREEP-rich geochemical province. We then go on to explore how the high abundance of heat-producing elements in this province affected the Moon's thermal evolution. Finally, we discuss some of the far-reaching consequences that these model results have for interpreting the geologic history of the Moon.

2. Evidence for the Existence of the Procellarum KREEP Terrane

In this section we present evidence from Lunar Prospector gamma-ray data, Apollo samples, and models of the geophysical structure of lunar impact basins to show that the Procellarum and Imbrium regions of the Moon are a unique KREEP-rich geochemical province. Furthermore, from these studies we infer that a substantial portion of the crust in this province is composed of a material similar in composition to Apollo 15 KREEP basalt.

Since this postulate is crucial to understanding the type of thermal model that we are to present in section 3, we discuss the supporting evidence for this premise even though some of what follows has already been detailed in the lunar literature. Though we do not specifically address why KREEP is concentrated in a single province, we agree with the suggestion of *Warren and Wasson* [1979] that this is likely to be a consequence of asymmetric crystallization of a near-global magma ocean.

2.1. Remote-Sensing Evidence

Global gamma-ray data from the Lunar Prospector mission [*Lawrence et al.*, 1998; *Elphic et al.*, 1999] show dramatically that the surface abundance of KREEP is highly enriched in a single province that encompasses Oceanus Procellarum, Mare Imbrium, Mare Frigorus, and the western portion of Mare Serenitatis. Figure 1 shows the areal extent of this province, here defined by a contour of approximately 3 ppm thorium (see also *Haskin et al.* [this issue] and *Jolliff et al.* [this issue]). A question of major importance to any lunar thermal model is whether the enhancement of radioactive elements in this province is solely surficial or whether the underlying crust is enriched as well. Given that the confines of this province closely follow mare/highland contacts (particularly in western Procellarum), it would seem reasonable to suspect that the elevated concentrations of incompatible elements there are due to the presence of KREEP-rich basaltic flows. However, since mare samples from the Apollo 12, 14, and 15 sites (which lie within this terrane) all have thorium concentrations of about 2 ppm or less [e.g., *Korotev*, 1998], this does not seem to be a probable scenario.

Out of the Apollo landing sites in the Procellarum KREEP terrane, Apollo 12 is the only mare site and thus probably comes closest to approximating the geology of the rest of Oceanus Procellarum. The basalts returned from this site were found to possess thorium concentrations of only 1 ppm or less, whereas the soils from this site were found to vary from about 3 to 10 ppm. It has long been argued that the composition of the soils at this site is due to mixing local basaltic material with an "exotic" KREEP component that is unrelated to the local basalts [e.g., *Hubbard and Gast*, 1971; *Hubbard et al.*, 1971; *McKay et al.*, 1971; *Meyer et al.*, 1971; *Evensen et al.*, 1974; *Jolliff et al.*, this issue]. Given that the mare basalts in this locale are relatively thin (<0.5 km [*De Hon*, 1979]) we agree with the suggestion of *Wasson and Baedecker* [1972] that the high abundance of incompatible elements in these soils is likely derived from vertical mixing with an underlying KREEP-rich crust.

As an alternative explanation, one could argue that the Apollo 12 site is not typical of Oceanus Procellarum and that the majority of basalts in this region are in fact KREEP basalts (~12 ppm Th). Volcanic KREEP basalts, however, are rare in the Apollo sample collection. Only two samples larger than 1 cm and a number of smaller clasts have been found at the Apollo 15 site, and only one breccia from the Apollo 17 site was found to contain a few KREEP basalt clasts. Furthermore, there are no recognized KREEP basalt lunar meteorites. If KREEP basalts did indeed make up the vast majority of basaltic flows in Oceanus Procellarum, then it is probabilistically unreasonable that there are so few of these rocks in the Apollo and lunar meteorite collection. These observations support the view that the majority of mare flows in Oceanus Procellarum are not KREEP basalts, but rather that they are likely to be similar in composition to the more ordinary mare basalts.

2.2. Evidence from the Imbrium Impact

On the basis of the Apollo gamma-ray data, *Haskin* [1998] suggested that the Imbrium basin may have formed within a

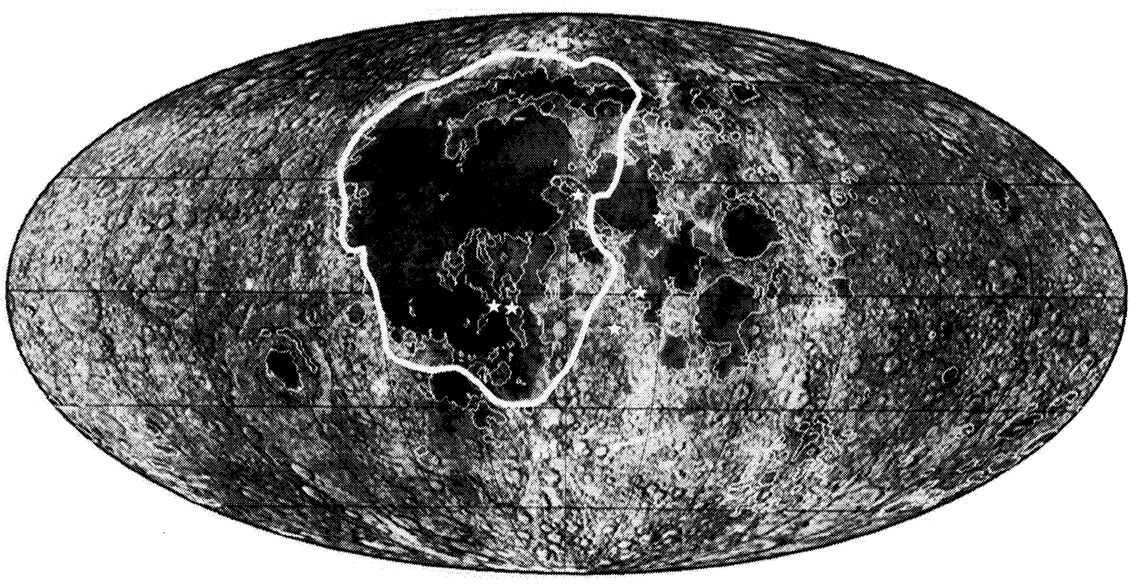

Figure 1. Areal extent of the Procellarum KREEP Terrane (thick line) overlaid on a lunar shaded relief map. The confines of the PKT are here defined by the 3-ppm thorium contour (using the empirical calibration of *Gillis et al.* [1999]). Thorium concentrations inside of this province average about 5 ppm but are locally as high as 9 ppm. The surrounding highlands, in contrast, are typically lower than 1 ppm. Stars represent the locations of the Apollo landing sites, and the thin lines represent the mare/highland contacts as mapped by *Wilhelms* [1987]. The surface area of the Procellarum KREEP terrane is about 16% of the Moon's total surface area.

region of crust that was significantly enhanced in KREEP. Assuming this to be the case, the ejecta from this basin would be rich in incompatible elements as well. Using modern impact scaling laws, he modeled the total thickness and composition of Imbrium's ejecta as a function of range from this basin's rim. Results from this modeling showed that the distribution of thorium outside of the Procellarum region of the Moon (using the limited Apollo data) could be explained solely as having an origin as Imbrium ejecta. Global Lunar Prospector thorium concentrations have confirmed this prediction [*Lawrence et al.*, 1998] and have additionally shown that other large basins (such as Serenitatis, Crisium, and South Pole-Aitken) have not excavated thorium-rich material from depth in the lunar crust (see also *Jolliff et al.* [this issue]).

The ejecta modeling of *Haskin et al.* [1998] showed that a significant quantity of Imbrium ejecta should be present at each Apollo site. Using the assumption that this ejecta would be enriched in thorium, they argued (as have *Evensen et al.* [1974], *Tera et al.* [1974], *Reid et al.* [1977], and *Schaeffer and Schaeffer* [1977]) that this ejecta was represented in the Apollo sample collection by the mafic impact-melt breccias (i.e., the Fra Mauro and very-high alumina "basalts"). *Korotev* [this issue] used this assertion to investigate the composition of the Imbrium target. He showed that the composition of most thorium-rich impact-melt breccias can be described primarily as a mixture of three components: (1) a material very similar in composition to that of Apollo 15 KREEP basalt, (2) typical feldspathic upper crustal material, and (3) highly forsteritic olivine. The proportions of these components vary for each landing site (as well as within sites), but average values are 58% KREEP basalt, 29% feldspathic upper crust, and 13% forsteritic olivine. The large proportion of KREEP basalt modeled in these impact melts is generally consistent with previous studies that used different assumptions about the end-member compositions [e.g., *Reid et al.*, 1977; *Ryder*, 1979].

The model results of *Haskin et al.* [1998] and *Korotev* [this issue] imply that a large portion of the Imbrium target was composed of KREEP basalt (or the intrusive or differentiated equivalents). Though the exact percentage of KREEP basalt in this locale is hard to quantify, *Korotev* [this issue] argues that more than half of the crust (~30 km) was likely composed of this composition. Since the abundances of uranium, thorium and potassium in KREEP basalt are about 300 times that of chondrites, this implies that a large portion of the Moon's incompatible elements is sequestered in this region of the crust (see also *Jolliff et al.* [this issue]).

2.3. Geophysical Evidence from Lunar Basins

Using gravity and topography data obtained from orbit, it is possible to infer variations in the thickness of the Moon's crust. Results from such studies show that the lunar Moho (i.e., the crust-mantle interface) is uplifted by tens of kilometers beneath the young multiring basins [e.g., *Wise and Yates*, 1970; *Bratt et al.*, 1985; *Neumann et al.*, 1996; *Wieczorek and Phillips*, 1998; *Arkani-Hamed*, 1998; *Wieczorek and Phillips*, 1999]. This behavior has commonly been attributed to the vast quantity of material that is excavated during the impact event and the subsequent rebound of the crater floor. *Wieczorek and Phillips* [1999] used this observation to argue that most of the lunar basins formed in accordance with the premise of proportional scaling. Specifically, they found that the depth/diameter ratio of the excavation cavity for most of the young basins was equal to ~0.1 independent of crater size [e.g., *Croft*, 1980; *Melosh*, 1989]. Imbrium and Serenitatis, however, were both found to possess anomalously shallow excavation depths.

The favored hypothesis of *Wieczorek and Phillips* [1999] for the anomalous crustal structure of Imbrium and Serenitatis was that these two basins formed in accordance with proportional scaling but that they were somehow modified subsequent to, or during, the basin-forming event. One suggestion was that if the crust beneath these basins was hotter than more typical regions of the Moon, then higher rates of viscous relaxation could have acted to reduce their reconstructed excavation depths. Since these basins are found to reside within the Procellarum KREEP Terrane (see Figure 1), higher crustal temperatures would be expected as a result of the high abundance of radioactive elements that are found there. It was additionally suggested that the presence of volcanic flows in these basins having the density of the lower crust (3.0-3.1 g cm^{-3}) could have biased the crustal structure inversions. The only lunar volcanic rock in the sample collection that was identified as having a density close to that of the lower crust was KREEP basalt (mare basalts have densities of about 3.3-3.4 g cm^{-3}). Voluminous KREEP basalt volcanism occurring in the Imbrium or Serenitatis basin is consistent with the model of *Korotev* [this issue] (see section 2.2) that suggests KREEP basalt is a major chemical component in the Procellarum KREEP Terrane. In addition, KREEP basalts (though they are rare) have been returned from the Apollo 15 and 17 sites which lie at the edge of the Imbrium and Serenitatis basins, respectively.

Wieczorek and Phillips [1999] also showed that many basins are out of isostatic equilibrium even if the excess load of the mare basalts is removed. This superisostatic configuration is consistent with the presence of "mascon" basins that do not possess associated mare volcanic products [e.g., *Neumann et al.*, 1996; *Konopliv et al.*, 1998]. A few basins in the *Wieczorek and Phillips* [1999] study (Imbrium, Serenitatis, Grimaldi, and Humorum), however, were found to be in a nearly isostatic state prior to the onset of mare volcanism. Each of these basins lies within, or on the edge of, the Procellarum KREEP Terrane as outlined in Figure 1. If the crust in this region of the Moon was hotter than typical (as suggested above), then these nonisostatic stresses would have been able to relax by accelerated rates of viscous creep.

Finally, if the Imbrium basin did indeed form in accordance with proportional scaling, and the dual-layer crustal thickness model of *Wieczorek and Phillips* [1998] is valid, then we can

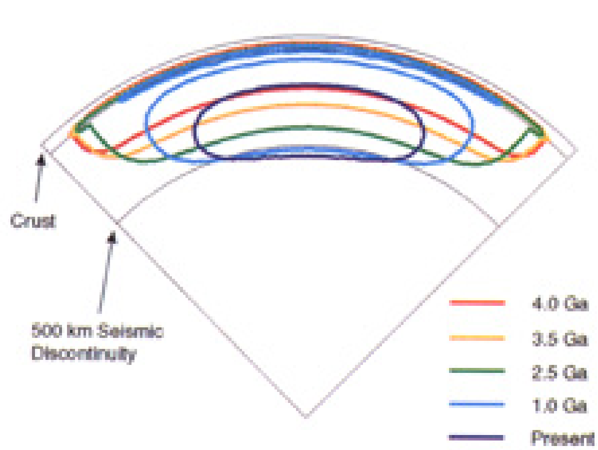

Plate 1. Plot showing the maximum spatial extent of melting beneath the Procellarum KREEP Terrane at times of 4, 3.5, 2.5, 1 and 0 billion years before the present. This cross section spans ±45° of latitude, whereas the PKT covers ±40°.

compute the proportions of upper crustal, lower crustal, and mantle materials that should have been excavated in this event. Using an excavation cavity diameter of 744 km [*Wieczorek and Phillips*, 1999], Imbrium's primary ejecta should be composed of approximately 42% upper feldspathic crust, 49% lower noritic crust, and 9% mantle. These proportions are comparable to the three end-member compositions of the mafic impact melts as determined by *Korotev* [this issue] if we assume that the mantle is composed of forsteritic olivine and that the lower crust in the Procellarum KREEP Terrane is composed of KREEP basalt. In concordance with the estimate of *Korotev* [this issue], this first-order result suggests that about 40 km of the Imbrium target (i.e., the thickness of the lower crust in this region of the Moon [*Wieczorek and Phillips*, 1998]) may have been composed of a material similar in composition to Apollo 15 KREEP basalt.

3. Implications for Mare Volcanism

From a simple inspection of Figure 1 it is clear that the distribution of mare basalts is correlated with the surface enhancement of KREEP. In fact, more than 60% of the mare basalts by area are found to occur within the confines of the Procellarum KREEP Terrane (which only makes up about 16% of the Moon's surface area), and a significant portion of the remainder are found to border this terrane. It is thus a reasonable question to ask if the enhancement of heat-producing elements in the Procellarum KREEP Terrane significantly influenced the thermal history of this region. Some researchers have suspected or suggested that this region of the Moon may have been hotter than typical, or perhaps even partially molten [e.g., *Spudis et al.*, 1984a; *Spudis et al.*, 1991; *Ryder*, 1994; *Haskin*, 1998]. In this section we present thermal modeling results which support this view and which additionally show that the enhancement of KREEP in this terrane likely caused the underlying mantle to partially melt over much of the Moon's history.

Because many lunar thermal models have been crafted in the attempt to explain mare volcanism [e.g., *Basaltic Volcanism Study Project (BVSP)*, 1981; *Kirk and Stevenson*, 1989; *Hess and Parmentier*, 1995; *Alley and Parmentier*, 1998], if one is to favor one model over another then it should satisfy more of the available constraints. Five constraints that we believe a successful thermal model for the Moon should satisfy are the following:

1. The bulk of lunar volcanism has occurred primarily on the nearside. This cannot simply be a consequence of regional variations in crustal thickness since the South Pole-Aitken basin (whose crust is thinner than that beneath Oceanus Procellarum [e.g., *Neumann et al.*, 1996; *Wieczorek and Phillips*, 1998]) does not contain appreciable quantities of mare basalt.

2. The spatial distribution of KREEP and mare volcanism is highly correlated. Furthermore, the elevated levels of KREEP that are observed from orbit are not solely due to the maria in this region.

3. Basaltic eruptions within the Procellarum KREEP Terrane have occurred over much of the Moon's history. The oldest mare samples here date to about 4.2 Ga [*Taylor et al.*, 1983], whereas ages based on crater counts show that volcanism continued up until at least 900 ± 300 Ma [*Schultz and Spudis*, 1983].

4. Even though most of the Moon's volcanism has occurred within the confines of the PKT (>60% by area), mare volcanism has occurred far from this province to a lesser extent as well. In contrast to the Procellarum region, though, this volcanism occurred almost exclusively during the Imbrian period.

5. On the basis of the assumption that the mare basalts were multiply saturated in both olivine and pyroxene at the time they formed, these volcanic rocks must all be derived from depths <250 km below the surface [e.g., *Longhi*, 1992]. If the mare basalts are genetically related to the picritic glasses, then the mare source must be <540 km below the surface [e.g., *Longhi*, 1992].

Though previous lunar thermal models have been able to account for one or more of the above constraints, prior to the Lunar Prospector mission none has attempted to explain what we think is the most important: The spatial correlation between mare volcanism and KREEP implies a genetic relationship between the two.

3.1. Thermal Modeling

Pre-Lunar Prospector thermal models [e.g., *BVSP*, 1981; *Kirk and Stevenson*, 1989; *Hess and Parmentier*, 1995; *Alley and Parmentier*, 1998] have all assumed that the distribution of heat-producing elements in the Moon varied only as a function of radius. In a recent attempt to explain the asymmetric distribution of KREEP and mare volcanism, *Zhong et al.* [1999] have shown that a degree-1 upwelling of KREEP- and ilmenite-rich magma ocean cumulates may have occurred early in the Moon's history. In contrast to these models we place a large portion of the Moon's heat-producing elements within a single crustal province, namely, the Procellarum KREEP Terrane (approximated here as a spherical cap with an angular radius of 40°).

The composition of the mafic impact-melt breccias suggests that the Imbrium target was largely composed of a material similar in composition to Apollo 15 KREEP basalt (see sections 2.2 and 2.3). Since the final dregs of magma ocean crystallization are believed to be initially sandwiched between the crustal and mantle cumulates [e.g., *Warren*, 1985], we place a layer of KREEP basalt at the base of the crust within the PKT. As was discussed in section 2, this terrane likely contains the equivalent of a layer of KREEP basalt about 30 or 40 km thick. As a conservative estimate, though, we only place a 10-km layer of this material within this terrane and note that a more realistic value would only act to enhance the thermal perturbations in our model. We acknowledge that some KREEP basalts would have erupted to the surface or intruded the overlying crust as dikes and plutons, and we investigate these effects below by redistributing the heat sources in the PKT over various crustal levels.

In modeling the thermal history of the Moon we have used a simple thermal conduction model. Though many have argued that convection may be an important form of heat transport in the lunar interior [e.g., *Cassen et al.*, 1979; *Schubert et al.*, 1979; *Turcotte et al.*, 1979; *Schubert and Stevenson*, 1980; *Hess and Parmentier*, 1995; *Alley and Parmentier*, 1998], we note that there is not unanimous agreement upon this premise [e.g., *Kirk and Stevenson*, 1989; *Pritchard and Stevenson*, 1999]. There are several reasons that lead us to suspect that heat may not have been transported primarily by thermal convection for much of the Moon's post magma ocean history. First, if most of the Moon melted during its formation (as modern giant-impact models seem to imply [e.g., *Stevenson*, 1987; *Cameron*, 1997; *Ida et al.*, 1997; *Pritchard and Stevenson*, 2000], then as this magma ocean crystallized, most of the incompatible and heat-producing elements would have been concentrated into a KREEP-rich layer sandwiched between the flotation anorthositic cumulates of the crust and mafic cumulates of the mantle [e.g., *Warren and Wasson*, 1979; *Warren*, 1985]. If this KREEP-rich material did not remix with the underlying mafic cumulates then the mantle

Table 1. General Thermal Model Parameters

Parameter	Value
Radial grid spacing	10 km
Time step	0.1 Myr
Crustal thickness	60 km
Surface temperature	-20°C
Angular radius of the PKT	40°
Thickness of KREEP basalt layer	10 km
Specific heat	1200 J kg^{-1} K^{-1}

would have been extremely depleted in heat-producing elements, lessening the possibility (or vigor) of mantle convection. Second, if most of the heat-producing elements were sequestered into the crust and/or upper mantle, then the deep lunar interior would have been heated from the top down. In contrast to the scenario of internal heating this would lessen the possibility for large-scale convective instabilities to develop. Third, it is possible that crystallization of a magma ocean could have led to a density-stratified mantle that is stable against convection (the models of *Herbert* [1980], *Spera* [1992], *Hess and Parmentier* [1995], and *Alley and Parmentier* [1998] all incorporate or predict some aspect of this assumption). Finally, density stratification within the upper mantle due to phase changes between plagioclase, spinel, and garnet could possibly hinder the vigor of convection.

Though these reasons do not prove that convection was not an important process for the mantle, they do nonetheless illustrate that the assumption of a purely conductive thermal model may be a plausible first-order approximation for much of the Moon's post magma ocean thermal evolution. In fact, it is not our aim to present a definitive lunar thermal model but rather to show that there are dramatic consequences to sequestering a large amount of heat-producing elements in a single crustal province.

As in the approach of all other thermal models, we are forced to make a variety of assumptions in order to simplify the numerical computations. We solve the spherical axially symmetric time-dependent conduction equation with heat sources using the method of finite differences (see the appendix for details). The axis of symmetry in our formulation passes through the center of the spherical cap containing the KREEP basalt layer. For simplicity, the thermophysical properties (thermal conductivity, specific heat, latent heat, density, and composition) were assumed to be independent of pressure and temperature within each of three distinct zones representing the mantle, feldspathic crust, and KREEP basalt layer. The latent heat of fusion for each zone was approximated by computing a mineralogic norm of its bulk composition and summing the mass-weighted latent heats of its respective minerals. Table 1 lists the general parameters of our model, and Table 2 lists the parameters specific to the distinct compositional zones.

We simulate melting in our model using the pressure-dependent solidus and liquidus of *Ringwood* [1976] (which is based on the bulk Moon composition of *Taylor and Jakes* [1974]; see Figure 2) for both the mantle and feldspathic crust. The low-pressure solidus and liquidus of KREEP basalt are significantly lower than that of the feldspathic crust and mantle and were taken from *Rutherford et al.* [1996]. We note that the liquidus temperature of KREEP basalt approximately corresponds to that of the mantle solidus. When the temperature exceeded the material's solidus in our model, but was below its liquidus, it was assumed that the incremental change in melt fraction and temperature were linearly related to the amount of heat that was added to the material (see the appendix). Though we tracked the amount of partial melt that was present for each node with time, we did not simulate the advection of heat that would occur if this magma was extracted and transported away from its source. Since the volume of erupted mare basalts is small (<1% the volume of the crust [*Head and Wilson*, 1992]), neglecting this form of heat transport should not significantly compromise our results.

The initial temperature conditions of our model were taken to approximate the final stages of magma ocean crystallization. It was assumed that the mantle had completely crystallized but that the residual urKREEP layer was still molten. Excluding urKREEP, the lunar magma ocean should completely crystallize in about 100-200 Myr [e.g., *Solomon and Longhi*, 1977; *Minear and Fletcher*, 1978]). Within this time period the mantle cumulates should also have overturned to achieve gravitational stability [*Hess and Parmentier*, 1995]. We start our thermal conduction model 4.5 billion years before the present and set the entire feldspathic crust to its solidus and the KREEP basalt layer to its liquidus. The top of the mantle was set to its solidus, and the temperature with depth in the mantle was assumed to follow an adiabat (increasing by about 40 K to the center of the Moon; see Figure 2). Using the values of the heat-producing elements that are listed in Table 2, our model Moon has a current bulk uranium concentration of 24 ppb, which is similar to estimates of the Earth's primitive mantle (~18 ppb [e.g., *Taylor*, 1992]) and about twice that of ordinary chondrites.

3.2. Results

For the specific parameters listed in Tables 1 and 2, Plate 1 shows the maximum spatial extent of melting that occurs in our

Table 2. Parameters Associated With the Distinct Compositional Zones

	Feldspathic Crust	KREEP basalt	Mantle
Thermal conductivity, W m^{-1} K^{-1}	2	2	3
Latent heat of fusion, J kg^{-1}	5×10^5	5×10^5	6.8×10^5
Density, kg m^{-3}	2900	3100	3400
U	0.14 ppm*	3.4 ppm†	6.8 ppb‡
Th	0.53 ppm*	12.4 ppm†	25 ppb‡
K, ppm	480*	5100†	17‡
Solidus	*Ringwood* [1976]	1296 K§	*Ringwood* [1976]
Liquidus	*Ringwood* [1976]	1448 K§	*Ringwood* [1976]

**Metzger et al.* [1977].
†*Korotev* [this issue].
‡*Warren and Wasson* [1979], assuming U = Th / 3.7, and K = 2500 U.
§*Rutherford et al.* [1996].

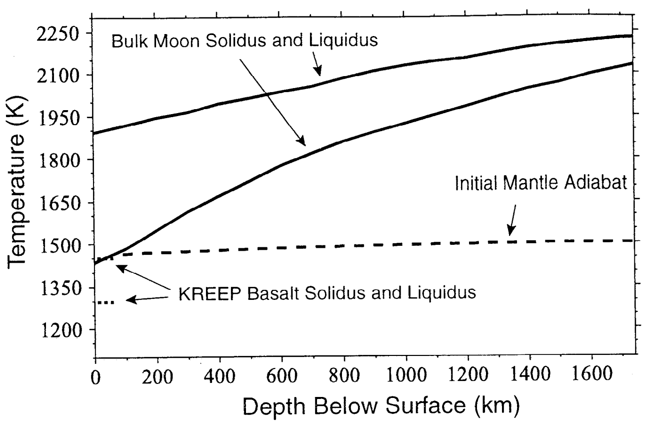

Figure 2. Plot of the bulk Moon liquidus and solidus of *Ringwood* [1976] and the KREEP basalt liquidus and solidus of *Rutherford et al.* [1996]. Also shown is the initial adiabatic temperature profile of the mantle.

thermal conduction model for a variety of times before the present. As is seen, melting in the lunar mantle occurs only directly beneath the Procellarum KREEP Terrane. This type of behavior occurs because the KREEP basalt layer in this region becomes heated above its liquidus (see Figure 3) and a portion of this heat is conducted downward into the underlying mantle. This general result is consistent with the observation that mare basalts are largely confined to the Procellarum KREEP Terrane.

The fraction of material that is melted beneath the PKT is plotted in Figure 4 as a function of depth for a variety of times before the present. The first feature that should be noted is that partial melting in the mantle occurs almost immediately after the model is started (~4.5 Ga) and continues to a lesser extent into the present. Melting initially occurs directly beneath the KREEP basalt layer and increases with depth as time progresses. Unfortunately, this behavior of our model cannot be adequately tested since the mare basalt samples are far from being representative of the Moon both in terms of their composition [e.g., *Pieters*, 1978] and age [e.g., *Head and Wilson*, 1992; *Nyquist and Shih*, 1992]. The maximum depth of melting predicted from this model is ~600 km below the surface, consistent with the petrologically determined maximum depth of melting [e.g., *Longhi*, 1992]. Figure 4 also illustrates that the KREEP basalt layer should remain partially molten for much of the Moon's history. This model result is due to both the high concentration of heat-producing elements that are present in this material and the fact that the solidus of KREEP basalt is significantly lower than that of the underlying mantle.

Figure 5 plots the modeled present day surface heat flux as a function of distance from the center of the Procellarum KREEP Terrane. As expected, the heat flux is found to be substantially higher within the confines of the PKT (~34 mW m^{-2}) as opposed to far from this region (~11 mW m^{-2}). Also plotted in Figure 5 are the surface heat flow measurements from the Apollo 15 and 17 sites (gray boxes) at their approximate distance from the center of the PKT. The corrected values due to a possible mare/highlands heat flow enhancement are shown as well (dotted boxes) [*Warren and Rasmussen*, 1987]. The measured values are seen to be consistent with the results of our model. In particular, the higher measured heat flow at the Apollo 15 site is shown to be a direct consequence of this site lying closer to the Procellarum KREEP Terrane. This is not a new interpretation for it has long been argued that the higher concentration of heat-producing elements at the Apollo 15 site (as observed from orbit) could have affected the heat flow measurements there [e.g., *Langseth et al.*, 1976].

Unfortunately, both of the Apollo sites for which we posses heat flow data are located close to the boundary of the Procellarum KREEP Terrane. The measured heat flux at these sites is thus presumably due to some combination of both the enhanced heat sources in the crust there and the background contribution from the mantle. Though the Apollo 17 heat flow measurement does not appear to be greatly affected by its proximity to the PKT in Figure 5, the subsurface extent of the KREEP basalt layer in this region may be greater than is alluded to in Figure 1. As a consequence of the unfavorable locations of these heat flow measurements, the Apollo measurements do not place a significant constraint on the bulk concentration of heat-producing elements in the Moon.

We plot the amount of melt that is generated per year as a function of time in Figure 6. The amount of magma that is generated within the Moon is shown to decrease exponentially with time. This curve is very similar to that predicted by the thermal conduction model of *Kirk and Stevenson* [1989], which used a different set of assumptions than was employed here. For comparison, we also show the measured mare basaltic production rate of *Head and Wilson* [1992]. The observed volcanic record generally mirrors the magmatic production of our model; however, our model generates almost 60 times more magma than has apparently reached the surface in the form of mare flows (see Table 3). This result seems to imply that only a small fraction of the magma produced in the mantle ultimately makes its way to the surface. The remainder of this magma may be sequestered as dikes and plutons in the crust [e.g., *Dvorak and Phillips*, 1978; *Head and Wilson*, 1992], may have sunk to the deep lunar interior [*Delano*, 1990], or may never have been extracted from its mantle source.

As a test of the sensitivity of our model results to our initial conditions, we have investigated a number of permutations of the above model. First, even if the KREEP basalt layer was initially sandwiched between the crust and mantle, some of this magma likely intruded the overlying crust as dikes and plutons or erupted onto the surface. We investigated this phenomenon by spreading the heat sources of our initial 10-km layer of KREEP basalt over a larger thickness of the lower portion of the crust. Our results show (see Table 3) that as the heat production of the lower crustal KREEP basalt layer is diluted into the overlying crust that both

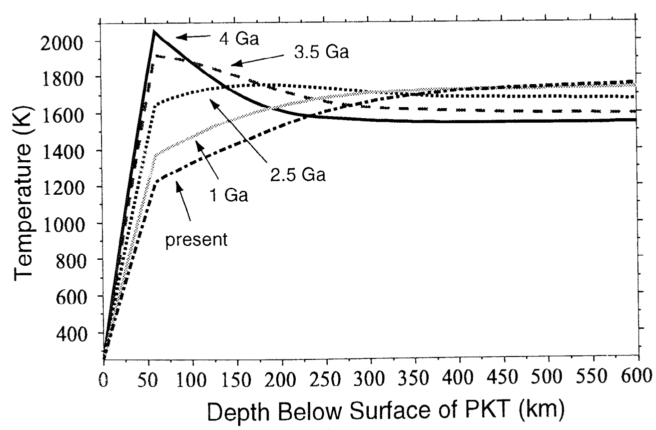

Figure 3. Plot showing the temperature beneath the center of the PKT at times of 4, 3.5, 2.5, 1, and 0 billion years before the present.

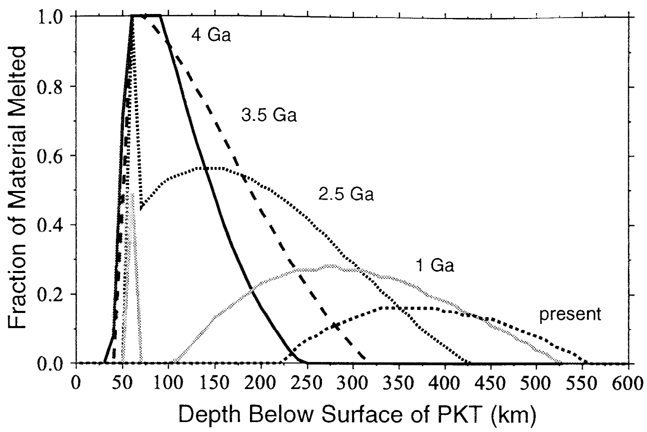

Figure 4. Plot showing the fraction of material melted beneath the center of the PKT at times of 4, 3.5, 2.5, 1, and 0 billion years before the present.

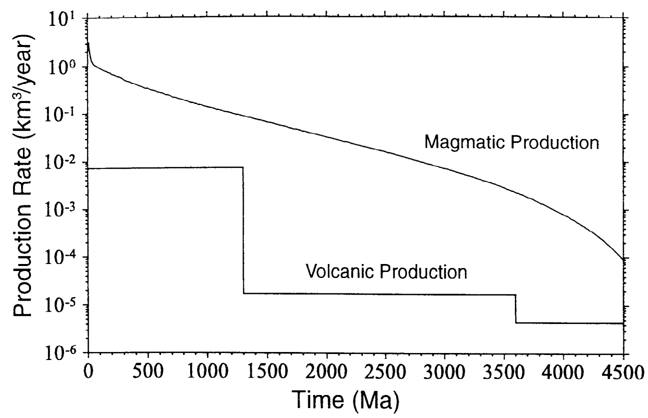

Figure 6. Plot showing the modeled lunar magmatic production rate as a function of time. The estimated basaltic production rate is taken from *Head and Wilson* [1992].

the maximum depth and duration of melting in the mantle decrease. However, even if the heat production associated with 10 km of KREEP basalt is spread evenly over the entire crust, melting in the mantle will still reach a depth of 180 km and span a duration of about 1.5 Gyr. Second, we investigated a model in which the initial mantle temperature was uniformly set to 1000 K (about 500 K cooler than our initial model). Melting in this model reached a depth of 470 km and lasted for about 3.5 Gyr. Lastly, we ran a model that did not possess any heat-producing elements in the mantle. Melting in this model reached 340 km and lasted about 4 Gyr. It thus seems inescapable to us that if the Procellarum KREEP Terrane contains the equivalent of a 10-km layer of KREEP basalt, and if convection is not an important process for heat transport within the Moon, then melting in the mantle directly beneath this terrane must have occurred. The maximum depth and duration of this melting depends upon the initial temperature conditions and the vertical distribution of heat sources in the crust of this region.

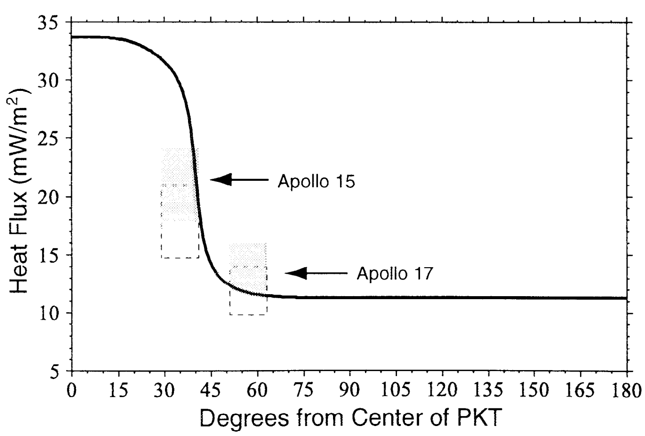

Figure 5. Plot showing the modeled present-day surface heat flux as a function of distance from the center of the Procellarum KREEP Terrane. Also shown are the Apollo heat flow measurements (gray boxes) at their approximate distance from the center of the PKT. The estimated heat flux at these sites after correction for a possible enhancement due to their location at a mare/highlands boundary [*Warren and Rasmussen*, 1987] is given by the dotted boxes.

Our model successfully explains the following aspects of mare volcanism:

1. The enhancement of heat-producing elements in a single crustal terrane leads to widespread partial melting of the mantle directly beneath this region. This explains the observed spatial correlation between the distribution of volcanic resurfacing and heat-producing elements.

2. Our model predicts that basaltic volcanism within the Procellarum KREEP Terrane should have occurred for most of the Moon's history, consistent with the dated samples [*Taylor et al.*, 1983] and photogeologic mapping [*Schultz and Spudis*, 1983].

3. The maximum depth of melting in our model is consistent with the maximum depth of origin of the mare source using the assumption that the basaltic and picritic magmas were multiply saturated with olivine and pyroxene in their source [e.g., *Longhi*, 1992].

Our model does not directly address the origin of mare basalts that lie outside of the Procellarum KREEP Terrane. We do note, though, that by increasing the heat production or initial temperature of the mantle (both by not unrealistic values) that our model is able to partially melt portions of the farside mantle as well. In contrast to the above results, however, near-surface melting does not occur. For instance, by setting the initial temperature of the entire mantle to its solidus, we found that the farside mantle would melt only at depths >300 km below the surface. Although this model result could be taken as being consistent with the geologic observations, melting of the farside mantle is not an inevitable outcome as is the melting that occurs directly beneath the PKT. Aspects of other thermal models (such as those of *Hess and Parmentier* [1995] and *Alley and Parmentier* [1998]) may be necessary in order to explain the origin of farside mare volcanism.

3.3. Caveats and Unresolved Issues

Even though our model appears to be consistent with most of the first-order constraints imposed by the samples and photogeology, we recognize that there are several important issues and potential problems that will eventually have to be addressed. Below are three of the most important.

1. The mare basalts in the Apollo sample collection do not appear to be heavily contaminated by a KREEP component, yet

Table 3. Maximum Depth and Duration of Melting for Various Model Permutations

Model	Mantle Heat Production?	Depth of Melting, km	Duration of Melting, Gyr	Total Melt Generated / Volume Mare Basalts*
10 km of KREEP basalt at the base of the crust	yes	560	4.5	56
10 km of KREEP basalt spread over the lower 20 km of the crust	yes	480	4.0	35
10 km of KREEP basalt spread over the lower 30 km of the crust	yes	400	3.5	22
10 km of KREEP basalt spread over the lower 40 km of the crust	yes	330	2.5	14
10 km of KREEP basalt spread over the lower 50 km of the crust	yes	260	2.0	8
10 km of KREEP basalt spread over the entire crust	yes	180	1.5	5
10 km KREEP basalt at the base of the crust; initial mantle temperature = 1000 K	yes	470	3.5	21
10 km of KREEP basalt at the base of the crust	no	340	4.0	41

*Total volume of mare basalts assumed to be 1.0×10^7 km^3 [*Head and Wilson*, 1992].

our model predicts (as do others, such as *Solomon and Longhi* [1977]) that they were transported though a partially molten KREEP basalt or urKREEP layer. It may be possible to crystallize this KREEP-rich layer in our model more quickly by including convection within the mantle and the temperature dependence of specific heat and thermal conductivity within the crust. In addition, it is possible that our postulated KREEP basalt layer may have varied in thickness throughout the Procellarum KREEP Terrane, and that it may even have been absent in places. Alternatively, KREEP basalt may have been present in the crust in the form of numerous bladed dikes, as opposed to a single giant "sill."

2. The fact that the Imbrium basin is supporting a mascon that is likely to be solely due to uncompensated mare basalts [*Wieczorek and Phillips*, 1999] may be at odds with the crust beneath this region being hotter than typical. This observation may require that the crust in this region cooled faster than is predicted by our model.

3. The bulk moon solidus and liquidus of *Ringwood* [1976] may not be totally appropriate for the Moon's interior given that the Moon did indeed differentiate. A mantle overturn following crystallization of a magma ocean [e.g., *Herbert*, 1980; *Hess and Parmentier*, 1995] may additionally have brought refractory magnesium cumulates to the top of the mantle, increasing the liquidus and solidus temperatures in comparison to that of *Ringwood* [1976]. This would act to reduce the magnitude of melting as compared with our present model.

4. Global Implications of the Procellarum KREEP Terrane

The fact that a single region of the Moon's crust is enriched in heat-producing elements and that massive amounts of melting likely occurred beneath this province has many far-reaching implications for the Moon's geologic evolution. We first show that the Imbrium impact likely excavated into a KREEP basalt magma chamber and that the Apollo 15 KREEP basalts were consequently extruded into this basin. Second, we suggest a possible explanation for the paradox of why the highly evolved KREEP-rich rocks contain a high abundance of magnesium. Finally, we suggest that the 500-km seismic discontinuity may represent the maximum depth of melting of the mare source.

4.1. Nature of the Imbrium Impact and the Search for LKFM

The composition of what is likely Imbrium ejecta [e.g., *Haskin et al.*, 1998] implies that a substantial portion of the crust beneath the Procellarum and Imbrium regions of the Moon is composed of a material similar in composition to Apollo 15 KREEP basalt [*Korotev*, this issue]. Since KREEP basalt is highly enriched in incompatible elements, we have showed that this material would remain partially to fully molten for much of the Moon's history. The Imbrium impact thus likely excavated into an extensive KREEP basalt magma chamber that was present beneath what we now recognize as the Procellarum KREEP Terrane. Immediately following this impact event, KREEP basaltic lavas from this denuded magma chamber would have flowed into Imbrium's excavation cavity. Typical mare basaltic volcanism that occurred millions of years later would have subsequently covered these KREEP basalt flows.

A scenario similar to this was originally proposed by *Ryder* [1994]. He argued that the Apollo 15 KREEP basalts are extrusive volcanic rocks that were derived from the nearby Apennine bench [*Spudis*, 1978] and that these flows were indistinguishable in absolute age from that of the Imbrium impact. On the basis of this evidence he suggested that the Imbrium impact and KREEP basaltic volcanism were somehow genetically related. One explanation that he put forth was that the decrease in pressure associated with the rebound of Imbrium's transient cavity could have caused massive melting, but only if the temperature of the target was already at or near its solidus. Since the adiabatic uplift associated with the Imbrium event would raise the temperature of the underlying mantle by only about 5 K (assuming a transient cavity depth of 85 km and an adiabatic gradient of 0.06 K km^{-1}) this explanation is clearly implausible. Alternatively, he suggested that the subcrustal portion of the Imbrium target may have been partially molten and that KREEP basalt volcanism was triggered by the removal of the overlying crust. He further speculated that since the Imbrium region of the Moon appeared to be enhanced in heat-producing elements (and hence was hotter), impact-induced volcanism may have been unique to the Imbrium impact, consistent with the predictions of our thermal models. We note that if the Serenitatis target was similar to that of Imbrium, then it is possible that the 4.08-Ga age of the Apollo 17 KREEP basalts [*Shih et al.*, 1992] may likewise date the Serenitatis impact.

Our hypothesis that the Imbrium target contained a molten magma chamber is also consistent with the petrology of the mafic impact-melt breccias. From analyses of the lunar samples it was quickly noted that the collection did not include a crystalline igneous rock that had the composition of these melts, specifically, the composition of LKFM (Low-K Fra Mauro basalt, a subset of the melt breccias [*Reid et al.*, 1972]). On the basis of this observation, *Reid et al.* [1977] suggested that the LKFM composition could have been a mixture of other lunar rock types and that about a third of this mixture had a composition similar to that of KREEP basalt. At the time, however, the LKFM composition could not be adequately modeled as a mixture of known lunar materials, and it was commonly assumed that the "missing component" was derived from the lower lunar crust [e.g., *Ryder*, 1979; *McCormick et al.*, 1989]. In the hope of shedding light on the origin and provenance of LKFM, *Spudis et al.* [1991] undertook a detailed chemical study of clastic material found in impact melts that were likely derived from the Imbrium basin. Though they did not find any clasts of the "missing component," they did show that the clasts present were likely derived from plutonic Mg-suite rocks. Absent in these melts were ferroan anorthositic clasts, and this suggested to them that the composition of the Imbrium target may have been atypical for the Moon. As a possible explanation that could account for the absence of lithic precursors to LKFM in these impact melts, *Spudis et al.* [1991] suggested that the "missing component" of LKFM may have been a magma that was present in the lower portions of the crust.

As we discussed in section 2.2, *Korotev* [this issue] has shown that the composition of the mafic impact melts can in fact be adequately described as a mixture of KREEP basalt and other known lunar lithologies. The fact that KREEP basalt clasts are not found in these melt breccias suggests that this component could indeed have been molten at the time of the Imbrium impact, consistent with the suggestion of *Spudis et al.* [1991] and *Ryder* [1994]. Additionally, the fact that the Apollo 15 KREEP basalts were extruded shortly after the Imbrium impact suggests that a KREEP basalt magma chamber could have existed beneath this province at that time. Both of these observations are consistent with the results of our thermal model, which suggests that a KREEP basalt layer within the crust would indeed have been molten at the time of the Imbrium impact.

4.2 Origin of Mg-Suite Rocks

Lunar igneous rocks can be broadly classified into three main groups: ferroan anorthosites, Mg-suite rocks, and mare basalts. The ferroan anorthosites are generally believed to be derived from crystallization of an extensive lunar magma ocean, while the mare basalts are usually interpreted as being derived from partial melting of the magma ocean's mafic cumulates [e.g., *Warren*, 1985]. The relationship of the Mg-suite rocks (i.e., the troctolites, norites, dunites, and KREEP basalts) to a lunar magma ocean, however, is less clear. As a magma ocean crystallizes, the remaining magma should become enriched in incompatible elements, and in this sense the Mg-suite rocks fit nicely into a lunar magma ocean scenario if they are genetically related to the final remaining dregs of crystallization. However, as a magma ocean continues to crystallize, the remaining magma should also become increasingly iron rich, which is clearly not the case for the Mg-suite rocks.

To explain this paradox, many have suggested scenarios in which the magmas parental to the Mg-suite rocks could have been generated by mixing evolved iron-rich magma ocean residuals with mantle-derived Mg-rich melts [e.g., *Longhi and Bourdrea*, 1979; *Norman and Ryder*, 1980; *James*, 1980; *Longhi*, 1980; *Warren and Wasson*, 1980; *James and Flohr*, 1983; *Warren*, 1986]. *Warren* [1988] used such a scenario in an attempt to model quantitatively the composition of the pristine KREEP basalts. By mixing urKREEP with a primitive mantle-derived Mg-rich magma (and allowing this magma to assimilate ferroan anorthositic country rock and to fractionally crystallize), he was able to reproduce the KREEP basalt composition to a good approximation. Furthermore, it has been shown that the composition of most Mg- and alkali-suite rocks can be accounted for by fractionally crystallizing KREEP basalt [e.g., *Snyder et al.*, 1995a, b].

Though the model of *Warren* [1988] was successful at generating a KREEP basalt magma, the source and origin of his mantle-derived Mg-rich magma was never quantitatively addressed. The results of our thermal model suggest to us that the Mg-rich melt may simply be the result of a layer of urKREEP melting the underlying mantle. We offer the following scenario for generating the KREEP basalt composition and other Mg-suite rocks: (1) Crystallization of a magma ocean results in an exceedingly ferroan KREEP-rich layer sandwiched between the flotation crust and mafic cumulates (i.e., urKREEP [*Warren*, 1988]). (2) The mafic cumulate pile is initially gravitationally unstable with dense iron-rich cumulates overlying less dense magnesium cumulates [e.g., *Herbert*, 1980]. This gravitationally unstable configuration overturns on a timescale of a million years, bringing early magnesium-rich cumulates to the top of the mantle [*Hess and Parmentier*, 1995]. (3) The iron-rich urKREEP magmas melt the underlying magnesian-rich cumulates, as we demonstrated in section 3. These magmas mix, giving rise to a magnesian and KREEP-rich magma sandwiched between the crust and mantle having the composition of KREEP basalt [*Warren*, 1988]. (4) Differentiation of this KREEP basalt magma over a prolonged period of time gives rise to the Mg- and alkali-suite rocks that are observed in the Apollo sample collection [e.g., *Snyder et al.*, 1995a, b].

One consequence of our proposed scenario is that the Mg-suite rocks should be locally present only beneath the Procellarum KREEP Terrane (see also *Korotev* [this issue]). Hence we predict that the mafic anomaly associated with the farside South Pole-Aitken basin is not due to the presence of Mg-suite rocks but is rather a result of mafic ferroan rocks derived from the lower crust (see also *Jolliff et al.* [this issue]). Another related consequence is that the "noritic" ejecta surrounding the largest basins [*Spudis et al.*, 1984b] outside of the PKT should have a ferroan affinity as well.

4.3. Nature of the Moon's 500-km Seismic Discontinuity

Seismometers placed on the lunar surface during the Apollo 12, 14, 15, and 16 missions allowed a crude seismic velocity structure of the Moon's interior to be constructed [e.g., *Toksöz et al.*, 1974; *Goins et al.*, 1981; *Nakamura et al.*, 1982]. Using the complete seismic database, *Nakamura et al.* [1982] modeled the seismic velocity of the mantle for three arbitrary constant velocity layers. For the first two layers (down to 500 km) the seismic velocities were shown to decrease slightly with depth, and below 500 km the velocity was found to sharply increase. Though they cautioned that the 500-km seismic discontinuity was not necessarily real and was chosen as a matter of convenience for the modeling, the overall velocity structure suggested to them that the lower mantle (beneath ~500 km) was compositionally distinct from the upper mantle. Recently, *Khan et al.* [1999] have re-analyzed the Apollo seismic data using a more sophisticated Monte Carlo approach. Their results show that the mantle likely possesses a constant velocity down to a depth of 500 km. Directly beneath 500 km the seismic velocity is found to sharply increase, and beneath 600 km the velocity structure becomes more complicated and less well constrained. These results seem to suggest that the upper 500 km of the mantle beneath the Procellarum region do not vary substantially in bulk composition. The sharp increase in velocity that is found at 500 km argues much more persuasively for the existence of a seismic discontinuity at this depth.

One interpretation for the origin of the 500-km discontinuity was that it could be partially the result of a phase transition between spinel and garnet [e.g., *Hood*, 1986; *Hood and Jones*, 1987; *Mueller et al.*, 1988]. The depth at which this phase transition occurs is temperature dependent and should occur between about 400 and 550 km [e.g., *Hood*, 1986]. In addition to this phase change, *Mueller et al.* [1988] also argued that the mantle beneath 500 km depth was likely to be more aluminous than the upper mantle (see also *Hood and Jones* [1987]). On the basis of this observation they suggested that the 500-km discontinuity may correspond to the maximum depth of the lunar magma ocean and that the deeper mantle represented some form of primitive or undifferentiated composition.

Many researchers have suggested that a deep primitive lunar interior could possibly have escaped melting at the time of a lunar magma ocean [e.g., *Taylor*, 1982; *Warren*, 1985; *Hood and Jones*, 1987; *Mueller et al.*, 1988]. The scenario that the 500-km discontinuity could represent this putative maximum depth of melting, however, does not seem very plausible to us. First, if the Moon formed as a result of a "giant impact," it seems hard to avoid that the Moon's initial state would be very hot, and possibly completely molten, as a result of its short accretion time in Earth orbit [e.g., *Stevenson*, 1987; *Cameron*, 1997; *Ida et al.*, 1997; *Pritchard and Stevenson*, 2000]. Second, magma ocean cumulates overlying a "primitive" mantle would be gravitationally unstable and would likely lead to a global mantle overturn and homogenization as the models of *Herbert* [1980], *Spera* [1992], and *Hess and Parmentier* [1995] seem to imply. A putative iron or ilmenite core-forming event would also have enhanced mixing between these two mantle layers. Even if a magma ocean was only 500 km deep, it seems unlikely to us that this compositional discontinuity could have remained unmodified and would still be detectable at the present time.

We offer a new interpretation for the origin of the 500-km discontinuity. On the basis of the premise that the mare basalts and picritic glasses were multiply saturated when they formed, the region of the mare source extended to a depth no greater than about 540 km below the surface [e.g., *Longhi*, 1992]. Consistent with this observation, our thermal model predicts that partial melting of the mantle should also occur to a depth of no more than 600 km (see Table 3). On the basis of this coincidence, as well as the fact that the four Apollo seismic stations lie in or close to the PKT, we suggest that the 500-km seismic discontinuity may in fact represent the maximum depth of melting that occurred in the mare source beneath the Procellarum KREEP Terrane. In this scenario the deeper portions of mantle there that did not have basaltic melts extracted from them would be richer in aluminous phases in comparison to the mantle directly above. This view is consistent with the results of *Mueller et al.* [1988] that suggest the mantle beneath 500 km is indeed more aluminous than the upper mantle. Finally, since our thermal model predicts that most melting in the mantle should occur directly beneath the Procellarum KREEP Terrane, our hypothesis suggests that the 500-km seismic discontinuity may not be global in extent, but may be a feature local to this region of the Moon.

5. Summary and Conclusions

Gamma-ray data obtained from the Lunar Prospector spacecraft show that the surface abundance of KREEP is highly concentrated in a single province that encompasses Oceanus Procellarum and Mare Imbrium [*Lawrence et al.*, 1998; *Elphic et al.*, 1999]. The hypothesis that the entire crust in this region is enriched in incompatible elements [e.g., *Haskin*, 1998; *Jolliff et al.*, this issue] is supported by (1) the anomalous crustal structure of basins that formed in this region [*Wieczorek and Phillips*, 1999], (2) the radial variation in composition of Imbrium's ejecta as determined from orbit [*Haskin*, 1998; *Lawrence et al.*, 1998; *Haskin et al.*, this issue], and (3) the composition of samples that are likely to be representative of Imbrium's ejecta [*Haskin et al.*, 1998; *Korotev*, this issue]. Furthermore, the chemical mixing models of *Korotev* [this issue] and the geophysical studies of *Wieczorek and Phillips* [1999] both support the notion that a large portion of the crust in this region (on the order of 10 km or more) is composed of a material similar in composition to Apollo 15 KREEP basalt.

We investigated the thermal consequences of sequestering a large portion of the Moon's heat-producing elements within a single crustal terrane by utilizing a simple thermal conduction model. By placing the equivalent of a 10-km-thick layer of KREEP basalt within the crust of the Procellarum KREEP Terrane, we have shown that the mantle directly below this province would partially melt over much of the Moon's history. The depth of partial melting was shown to increase with time, and the KREEP basalt layer was found to remain molten for a large fraction of that time. Though the duration and maximum depth of melting depended upon the initial conditions of our model, we have shown that large-scale melting beneath this region is inescapable. The results of our model are consistent with (1) the fact that most mare basalts are located within the confines of the Procellarum KREEP Terrane, (2) the present-day heat flow measurements at the Apollo 15 and 17 landing sites [*Langseth et al.*, 1976], (3) the long duration of mare volcanism [e.g., *Schultz and Spudis*, 1983; *Taylor et al.*, 1983], and (4) the maximum depth of the mare source as inferred from the pressures of multiple saturation of the mare basalts and picritic glasses [e.g., *Longhi*, 1992].

The results of our thermal model also elucidate a number of lunar observations and problems. First, our model predicts that the KREEP basalt layer within the Procellarum KREEP Terrane should have been partially molten when the Imbrium basin formed. This impact event thus likely excavated into a KREEP basalt magma chamber, consistent with the suggestions of *Spudis* *et al.* [1991] and *Ryder* [1994]. Second, our model provides a mechanism for forming the KREEP basalt composition. Following solidification of a lunar magma ocean, a gravitational overturn of the cumulate pile would likely have brought early magnesian cumulates to the upper mantle [e.g., *Herbert*, 1980; *Spera*, 1992; *Hess and Parmentier*, 1995]. We suggest that urKREEP within the Procellarum KREEP Terrane melted and mixed with these cumulates, giving rise to a Mg- and incompatible-rich KREEP basalt magma. Finally, we suggest that the 500-km seismic discontinuity [e.g., *Nakamura et al.*, 1982; *Khan et al.*, 1999] may represent the maximum depth of melting that occurred beneath this region.

Appendix

We solved the spherical time-dependent thermal conduction equation with heat sources

$$\nabla \cdot (k \nabla T) + \rho H = \rho C_p \frac{dT}{dt},$$

where k is the thermal conductivity, ρ is the density, H is the heat production per unit mass, C_p is the specific heat per unit mass at constant pressure, and T is the temperature in degrees Kelvin (all of which depend implicitly upon position). If the thermal conductivity depends only on radial position, then for an axially symmetric system in spherical coordinates the above equation reduces to

$$C_p \frac{dT}{dt} = \frac{2k}{r}\frac{dT}{dr} + k\frac{d^2T}{dr^2} + \frac{k \cot\theta}{r^2}\frac{dT}{d\theta} + \frac{k}{r^2}\frac{d^2T}{d\theta^2} + \frac{dk}{dr}\frac{dT}{dr} + \rho H.$$

Using the divergence theorem, the equation governing the temperature evolution of the origin (r=0) is found to be

$$\frac{k}{\Delta V}\oint_A \frac{dT}{dr} da + \rho H = C_P \frac{dT}{dt},$$

where ΔV is the volume inside of the spherical surface of integration.

Starting with an initial condition, we solved the above equations using the method of finite differences. For one boundary condition the surface was set to a constant temperature (see Table 1). The boundary conditions along the symmetry axis were satisfied by specifying that the heat flow in the θ direction was zero on this axis (i.e., $dT/d\theta = 0$). Our finite difference grid was equally spaced in the radial direction. To speed computations, however, the grid spacing in the θ direction varied as a function of radius such that the grid points were approximately equidistant from each other.

The effects of melting were simulated when the temperature exceeded the material's solidus. At a given depth the amount of heat (i.e., increase in enthalpy) needed to completely melt a given volume of material is approximately

$$h_{\text{melt}} = \rho L_f + \rho C_p (T_l - T_s),$$

where L_f is the latent heat of fusion and T_l and T_s are the liquidus and solidus temperatures of the material, respectively. When the temperature exceeded the solidus, the amount of heat h subsequently added to the material was tracked. The melt fraction m and temperature of the material were then assumed to be linearly related to the material's heat content by

$$m = \frac{h}{h_m}$$

$$T = T_s + m(T_l - T_s).$$

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