Large volumes of mare basalts are present on the surface of the moon, located preferentially in large impact basins. Mechanisms relating impact basins and mare basalt eruptions have previously been suggested: lunar impacts removed low-density material that may have inhibited eruption, and created cracks for fluid flow [Icarus 139 (1999) 246], and lunar basins have long been described as catchments for magma (e.g., [Rev. Geophys. Space Phys. 18 (1980) 107] and references therein). We present a new model for melt creation under near side lunar basins that is triggered by the impacts themselves. Magma can be produced in two stages. First, crater excavation depressurizes underlying material such that it may melt in-situ. Second, the cratered lithosphere rises isostatically, warping isotherms at the lithosphere–asthenosphere boundary which may initiate convection, in which adiabatic melting can occur. The first stage produces by far the largest volume of melt, but convective melting can continue for up to 350 Ma. We propose that giant impacts account for a large portion of the volume and longevity of mare basalt volcanism, as well as for several compositional groups, including high alumina, high titanium, KREEP-rich, and picritic magmas.

Keywords: Moon; impact basin; mare basalt; mantle convection; volcanism

1. Introduction

Large volumes of mare basalts are visible on the surface of the Moon, preferentially lying in large impact basins. The processes of formation and eruption of these basalts are not well established, though it is recognized that eruptions peaked in the Imbrian Period (3.85–3.2 Ga), and that the majority of the basalts lie on the near side. Solomon [3] suggested that crustal thickness explains why so little basalt erupted onto the far side: the density of liquid basalt at pressures corresponding to the bottom of the crust are greater than that of the crustal rocks, so eruption is suppressed by gravitational density stability. Evidence from the Apollo 15 and 16 missions indicated that the nearside crust is thinner than the farside crust. Researchers have suggested that tectonic stresses and magma buoyancy or hydrostatic head should lead magma to stall beneath the lunar crust, preferentially extrude to the surface on the nearside, but cool in dikes on the farside of the moon (e.g. [2–6]).

The existence of a relationship between impact basins and mare basalt eruptions has been suggested by many investigators. In particular, large impacts are
thought to have removed enough of the low-density crustal lid that heavier, liquid mare basalts could erupt. Zhong et al. [7] further suggested that heating late cumulates at depth in the Moon would result in a mode 1 eruption, which could cause the preferential appearance of mare basalts on the near side.

Wieczorek and Zuber [8] reanalyzed the relative buoyancy of liquid mare basalts and lunar crustal compositions, and suggested that buoyancy, rather than hydrostatic head, is the controlling factor in eruption. They find that all lunar basaltic liquids are less dense than the lunar lower crust. If impacts excavate the upper crust, therefore, all magmas may erupt at the impact site without consideration of hydrostatic head or composition. In some cases, where upper crust is thought to still be present, super-liquidus temperatures are required to enable eruption in this scenario.

Wieczorek and Phillips [9] coin the term “Procellarum KREEP terrane” for a region including the Procellarum and Serenitatis basins that is highly enriched in incompatible and heat-producing elements. This KREEP-rich terrane was previously suggested by Haskin [10]. Wieczorek and Phillips suggest that the crust in this area is compositionally similar to the Apollo 15 KREEP basalt, and find that the impact basins in this region are highly modified, consistent with viscous relaxation enabled by the anomalous heat production, or by voluminous volcanism. Wieczorek and Phillips find that heat from the enhanced KREEP layer would melt the Moon from the bottom of the crust downward to a maximum depth of about 600 km. They further predict that this molten region should persist for several billion years. The Imbrium basin, therefore, excavated largely molten material. They find that the long duration of the molten zone is consistent with the duration of mare basalt volcanism, as well as the depths of origin. Questions persist, however, on what prevented much more voluminous volcanism, if such a huge volume of material was completely molten, and on the persistence of an ancient anorthositic crust through such an immense melting event. The existence of a completely molten zone is also inconsistent with the correlations of depths of origin with types of magma [11]. The small volumes of melt produced over a wide range of depths are more consistent with adiabatic decompression melting than with stationary conductive heating.

The processes modeled here rely on mass excavation by giant impacts. First, crater formation reduces pressure beneath the crater by an amount equal to the lithostatic pressure of the material excavated. This lowering of pressure effectively moves the solidus deeper into the lunar material, in some cases causing it to cross the selenotherm and create melt. The evolution of the shape of craters, from the first seconds after impact through isostatic rebound at thousands of years after impact, is controversial, but because material excavated from a crater is thrown farther than the outer radius of inward flow, depressurization will result regardless of the magnitude of lateral crater collapse. If the mass has been ejected from the crater, then a mass deficiency exists for the lithosphere in the region of the crater, and our modeled processes would proceed.

A newly excavated crater is not in isostatic equilibrium. The thinned lithosphere rises and forms a dome in the lithosphere–asthenosphere boundary. This dome creates a horizontal temperature gradient at its edge, providing a driving force for convection. The dome may form virtually instantaneously, when the lithosphere behaves as a liquid due to the intense shock of the impact, or it may rebound isostatically over approximately the next $10^4$ years.

We present a new model for melt creation under near side lunar basins that is triggered by the impacts themselves. Magma can be produced in two stages. First, crater excavation depressurizes underlying material such that it may melt in-situ. Second, the cratered lithosphere rises isostatically, warping isotherms at the lithosphere–asthenosphere boundary and may initiate convection, in which adiabatic melting can occur (Fig. 1). This model is similar to a model for impact-produced volcanism on the Earth proposed by Jones et al. [12], though that model includes only the depressurization from excavation, and no later convection.

Because the surface of the planet is generally different compositionally from its interior, shock melts are compositionally distinct from mantle melts, and so are not included in melt volumes produced in this study. However, shock melt may mix with rising decompression melts to produce a range of mare basalt compositions. Shock melt from a large lunar basin is likely to contain aluminous, high titanium, and KREEP components, and thus may be an important contributor to the observed range of mare basalt compositions.
We find, based on analysis of mare basalt and picritic glass multiple saturation points, that a lithospheric thickness of 200 km and a mantle potential temperature of 1450 °C are the most realistic model for the Moon between 4.0 and 3.5 Ga. This impact model is consistent with gravity models of the lunar crust under the large basins and with petrologically constrained models of temperature and depth of mare basalt origin. Specifically, we attempt to explain the volume and longevity of mare basalt volcanism, along with the pattern of eruptions of varying compositions through time. There are a number of distinct basalt compositions from the Moon, including high-alumina basalts, basalts containing the hypothesized late-stage magma ocean liquid rich in potassium, rare earth elements, and phosphorus (KREEP), and basalts anomalously high in titanium. The aluminous and KREEP basalts appear, from returned samples, to have erupted earlier than the high-Ti basalts and picritic glasses, which in turn erupted before the low-Ti basalts and picritic glasses.

2. Models and methods

The models begin with a new crater excavated in the lunar crust, with initial rebound complete. The depth profile of model craters is from O’Keefe and Ahrens [13]. The maximum depth of excavation equals 0.15R, where R is the crater’s radius of excavation [13,14].

The complex crater depth profile, \(D_C\), of O’Keefe and Ahrens [13] is closely fit by the following expression, where \(R\) is the radius of excavation of the crater, and \(r\) is the radial distance from the center of the crater, all in kilometers.

\[
D_C = R \left( \frac{0.2r^3}{R^3} + \frac{0.022}{R} + 0.1 - 0.22 \right)
\]

When integrated, this profile yields the following expression for excavated crater volume, \(V\):

\[
V_C = 0.11\pi R^3
\]

This excavated volume is matched by the total ejecta volume obtained by integrating the following expression for ejecta blanket thickness, \(D_E\), as a function of distance from the crater’s outer edge:

\[
D_E = \frac{0.34R^4}{r^3}
\]

This crater depth profile is used in both the in-situ decompression melting model and the formation of
the lithospheric dome for subsequent convective melting. The volumes are in agreement with the ejecta law given in Housen et al. [14] and crater profiles are roughly in agreement with Cintala and Grieve [15] and Melosh [16].

The maximum depth of excavation is also the maximum isostatic uplift of the bottom of the lithosphere. On the moon, gravity modeling over lunar basins shows domes in the crust (mascons) with heights that are 5–24% of the excavation radii of the craters above them [1,17,18], implying initial excavation depths greater than 0.15R in some cases. These domes have been preserved for four billion years, and attest to the cold, rigid nature of the lunar upper lithosphere at the time of eruption of the mare basalts.

In the lunar mantle beneath the crater, we assume mantle potential temperatures of either 1350 or 1450 °C. These temperatures are consistent with the experimentally determined multiple saturation points of mare basalts and picritic glasses (Fig. 2; multiple saturation points are the approximate pressure and temperature at which the magma separated from its source). The picritic glasses imply a mantle potential temperature of about 1450 °C, and the boundary between the source depths of the glasses and the basalts may imply a thermal lithosphere about 200–250 km thick, consistent with lunar thermal evolution models [19].

Note that with these starting conditions, as implied by petrologic evidence, sluggish convection would be...
expected to be in place before impact, as well as some adiabatic mantle melting. Because the mantle potential temperature creates extant melt in the mantle before impact, any impact that affects the thickness of the lithosphere will create more melt. The smallest crater size that is likely to cause significant thinning of the lithosphere is about 50 km in diameter, but specifying a smallest size requires more detailed knowledge of the mechanics of the lithosphere than we know about the Moon.

2.1. Shock melting

Shock melting from the energy of impact was not modeled in this study. A 300-km radius crater makes about $10^5$ km$^3$ of shock melt, and a 100-km radius crater about $10^4$ km$^3$ of melt, from the results of Pierazzo et al. [20] and Tonks and Melosh [21]. Because the surface of the planet is different compositionally from its interior, shock melts are generally compositionally distinct from adiabatic melts, and would not be mistaken for them, and so are not included in total volumes of basaltic melt produced by impacts in this study. Another effect of shock melt may be important, though: its ability to mix with rising decompression melts to produce a range of mare basalt compositions. Shock melt from a large lunar basin is likely to contain aluminous, high titanium, and KREEP components, and thus may be an important contributor through melt mixing to the observed range of mare basalt compositions.

2.2. Stage 1: In-situ decompression melting

For simplicity, a conductive selenotherm through the lithosphere culminating at an adiabatic asthenosphere is assumed to remain fixed in the lunar material after excavation (Fig. 1). Excavation of the crater reduces pressure in the material beneath the crater by an amount equal to the lithostatic pressure of the material excavated. The greatest pressure release is over the interval from $0.2R$ to $0.4R$, and pressure release declines to zero at the crater rim (and can increase outside the rim due to new deposits). This lowering of pressure effectively moves the solidus deeper into the Moon, in some cases causing it to cross the selenotherm and create melt. A general equation for in-situ decompression melt was created by integrating the area of intersection between the solidus and selenotherm around the axis of symmetry of the crater. The parameters used in the equations are given in Table 1.

A critical parameter is $\beta$, the melt fraction per degree between solidus and liquidus. This can be calculated using $C_p$, the heat capacity of the silicate by integrating the area of intersection between the solidus and selenotherm around the axis of symmetry of the crater. The parameters used in the equations are given in Table 1.

<table>
<thead>
<tr>
<th>Table 1 Parameters used in melt volume calculations</th>
</tr>
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<tbody>
<tr>
<td><strong>Constants used in melt volume calculations</strong></td>
</tr>
<tr>
<td>$s$ slope of solidus $0.75 ^\circ$/km</td>
</tr>
<tr>
<td>$h_o$ maximum crater km</td>
</tr>
<tr>
<td>$R$ radius of excavation of the crater 50, 100, 200, 300 km</td>
</tr>
<tr>
<td>$\delta_o$ solidus temperature at 1 atmosphere 1150$^\circ$</td>
</tr>
<tr>
<td>$T_p$ mantle potential temperature 1350 or 1450 $^\circ$C</td>
</tr>
<tr>
<td>$d$ lithospheric thickness 200, 250, 300, or 350 km</td>
</tr>
<tr>
<td>$a$ slope of the adiabat 0.17$/^\circ$/km</td>
</tr>
<tr>
<td>$C_p$ heat capacity of silicates 1256.1 J/kg</td>
</tr>
<tr>
<td>$H_f$ heat of fusion of silicates 418,700 J/kg</td>
</tr>
<tr>
<td>$\beta$ frac. of melt produced $0.3$ wt.$/^\circ$/</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Variables used in in-situ melt volume calculations</th>
</tr>
</thead>
<tbody>
<tr>
<td>$z$ depth, 0 at original land surface km</td>
</tr>
<tr>
<td>$r$ radial distance from crater center km</td>
</tr>
<tr>
<td>$D_{\delta}(r)$ crater excavation depth km</td>
</tr>
<tr>
<td>$T_{A}(z)$ geotherm in the lithosphere $^\circ$</td>
</tr>
<tr>
<td>$T_{B}(z)$ geotherm in mantle $^\circ$</td>
</tr>
<tr>
<td>$T_{C}(z)$ solidus after crater excavation $^\circ$</td>
</tr>
<tr>
<td>$z_u$ where $T_A$ and $T_C$ cross: km</td>
</tr>
<tr>
<td>$z_l$ where $T_B$ and $T_C$ cross: km</td>
</tr>
<tr>
<td>$F$ volume of melt km$^3$</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Constants used in numerical modeling of convection</th>
</tr>
</thead>
<tbody>
<tr>
<td>$h$ height and width of model box 400 km</td>
</tr>
<tr>
<td>$\rho$ reference mantle density 3300 kg/m$^3$</td>
</tr>
<tr>
<td>$\Delta \rho$ difference between high-Ti cumulate and mantle densities $-400$ kg/m$^3$</td>
</tr>
<tr>
<td>$\Delta T$ temperature across the model box 1350 or 1450 $^\circ$C</td>
</tr>
<tr>
<td>$\eta_{o}$ reference viscosity $10^{20}$ Pas</td>
</tr>
<tr>
<td>$\alpha$ thermal expansivity $3 \times 10^{-5}$ $^\circ$/</td>
</tr>
<tr>
<td>$\sigma_{o}$ reference stress $1.6 \times 10^{13}$ Pa</td>
</tr>
<tr>
<td>$\kappa$ thermal diffusivity $10^{-6}$ m$^2$/s</td>
</tr>
<tr>
<td>$v$ number of elements in 100</td>
</tr>
<tr>
<td>$Ra$ Rayleigh number $1.4 \times 10^5$</td>
</tr>
<tr>
<td>$Ra_{comp}$ compositional Rayleigh number $-3 \times 10^5$</td>
</tr>
</tbody>
</table>
cumulates, and $H_0$, the heat of fusion of the silicate cumulates, as follows:

$$\beta = \frac{df}{dT} = \frac{C_p}{H_f} = 0.3 \text{ wt\%}$$  \hspace{1cm} (4)

Based on this $df/dT$ and the maximum decompression created by any impact modeling, the maximum melt percent of a parcel of mantle in the in-situ melting model is 6%.

The initial selenotherm (at the beginning of a calculation) is modeled as a linear profile through the lithosphere from the surface to the adiabatic potential temperature at the base of the lithosphere, and as an adiabat through the mantle. The following expressions describe the initial selenotherm in the lithosphere, which is conductive, and in the mantle, where it is adiabatic:

Lithospheric_selenotherm = $T_\Lambda = z \left( \frac{T_p}{d} + a \right)$  \hspace{1cm} (5)

Mantle_selenotherm_{(adiabatic)} = $T_B = az + T_p$  \hspace{1cm} (6)

(Shortly after the calculation begins, the initial kink diffuses to a more realistic error-function-like form.).

A solidus for mantle material can be expressed as follows, and subsequently as it appears following excavation of the crater:

Solidus = $sz + s_o$  \hspace{1cm} (7)

Solidus_following_crater_excavation = $T_C = s(z + D_C) + s_o$  \hspace{1cm} (8)

where the expression for the crater profile, $D_C$, is given by Eq. (1) above.

The maximum interval of melting between the solidus and selenotherm is found over the area of greatest excavation. As the radius increases toward the edge of the crater the solidus effectively moves up, following the crater floor, and the melting interval decreases (Fig. 1). The volume, $F$, of in-situ decompression melt is proportional to the area between the solidus and the selenotherm:

$$F = 2\pi \beta \left[ \int_0^a R \int_0^R r(T_B - T_C) dr dz + \int_a^d \int_0^R r(T_A - T_C) dr dz \right]$$  \hspace{1cm} (9)

where $T_A(z)$ is the conductive lithospheric temperature profile, $T_B(z)$ is a mantle adiabat, $T_C(z)$ is the solidus following excavation (this takes into consideration the crater profile, $D_C(r)$, $z_1$ is the intersection of the solidus and the adiabat, and $z_d$ is the intersection between the solidus and the conductive selenotherm.

2.3. Stage 2: Adiabatic melting in convection currents

Once the lithospheric dome is formed, mantle material flows upward in the center of the dome, and out, and down at the edges (Fig. 3). This convection is calculated numerically using a spherical axisymmetric version of the finite element code ConMan [22], called NewScam, using a non-Newtonian viscosity law composed such that a 100 °C change in temperature or a 3.3 GPa change in pressure create a factor of ten change in strain rate:

$$\eta_{nonN} = \eta_0 \exp \left( \frac{E + vz}{T + T_o} - \frac{E + vz_0}{1 + T_o} \right)$$  \hspace{1cm} (10)

where $\eta_0$, $\tau_0$, $z_0$, and $T_0$ are reference values for viscosity, deviatoric stress, depth, and temperature, respectively; $E$ is the activation energy, and $v$ is the reference volume. If the second invariant of the stress tensor is non-zero, then viscosity is recalculated such that a factor of 2.2 in deviatoric stress creates a factor of ten change in strain rate.

To first order, initial conditions model a weakly convecting lunar mantle, with a conductive thermal boundary layer overlying an adiabatic interior. The resulting melt is calculated with a post-processing routine (see Table 1).

Thermal convection is governed by the Rayleigh number:

$$Ra = \frac{\rho g \Delta Th^3}{\eta_0 \kappa}$$  \hspace{1cm} (11)

where $\kappa$ is thermal expansivity, $\kappa$ is thermal diffusivity, $\rho$ is a characteristic density, $\Delta T$ is the temperature change over the model box, $h$ is a characteristic length and $\eta_0$ is a reference viscosity. The compositional Rayleigh number controls convection due to the negative buoyancy of the high-Ti cumulate

$$Ra_{comp} = \frac{\Delta \rho gh^3}{\eta_0 \kappa} = Ra \left[ \frac{\Delta \rho}{(\rho \Delta T)} \right]$$  \hspace{1cm} (12)
where $\Delta \rho$ is the difference in densities between the compositions. A starting condition for each model run was created by approximating an erfc cooling law to make a cooled lithosphere of the required depth with a lithospheric dome with maximum uplift equal to 0.15 times the crater excavation radius.

For each of the models, the top 15 elements (60 km) are anorthosite crust, which is modeled with an initial viscosity 100 times that of the mantle. To test how changes in lithospheric geometry (isostatic uplift and later convection) affects the hypothesized high-titanium cumulate, it is included in the numerical model (for further description of this model, see Ref. [11]). After the anorthosite crust, the next layer is three elements (12 km) of KREEP. This is probably a significant overestimate of the thickness of the KREEP layer, but numerical considerations require that at least three elements be used in each material. The initial viscosity of the KREEP layer is 100 times less than the mantle. Below the KREEP layer are 5 elements (20 km) of high-Ti cumulate. The remainder of the model box is made up of 77 elements (308 km) of silicate mantle. The KREEP and high-Ti layers are removed within the radius of excavation of the crater in the models with crater radii greater than 100 km, because the KREEP and high-Ti layers would be significantly disturbed or melted by craters of these sizes. For these large craters, the high-Ti cumulate is retained outside the crater radius.

The boundary conditions for each model are the same: free-slip in the horizontal along the top boundary, free-slip in the vertical along the side boundaries, and a flow-through bottom boundary.

### 3. Results

In-situ decompression melt is the largest contributor to melt created by lunar impacts, constituting between 98 and 100% of the melt calculated in these models. In almost every case, the mantle potential temperature and selenotherm used in these models would create extant mantle melt before crater excavation. The total volume of this existing melt is calculated under the region that the crater will later excavate, and it is included in the total in-situ decompression melt, making up between 0% and 54% of the total in-situ melt volume.
Three main parameters control the volume of melt produced by in-situ melting: the slope of the sublithospheric selenotherm (here defined as the slope of the mantle adiabat), the mantle potential temperature, and the slope of the mantle solidus. Reasonable changes within these parameters change our results by less than an order of magnitude.

The results change slightly with changes in the slope of the sublithospheric selenotherm. The slope of the mantle adiabat used in these models is 0.17 °C/km, and other references use values within 0.02 °C/km of this value. Each 0.01 °C/km change in the adiabat changes the volume of the in-situ melt produced by about 5% (of course, lowering the slope of the adiabat lessens the volume of melt produced).

Mantle potential temperature changes the results more significantly. At 1400 °C, in-situ melt totals are about 60% of their 1450 °C values; dropping the mantle potential temperature to 1350 °C reduces in-situ totals by another 40% from their 1400 °C values, and dropping the temperature to 1300 °C drops totals another 30%. These differences are reflected in Figs. 4 and 5.

The slope of the mantle solidus is a larger question, since it depends upon the composition of the cumulates being melted, and this is a matter of debate. In this model we use 0.7 °C/km, in agreement with estimates for the green glass mantle source, but other researchers have used values as low as 0.4 °C/km. Each 0.1 decrease in the solidus slope increases in-situ melt volumes by about 40%.

In-situ melting occurs instantaneously, in a geological sense, after the impact. The longevity of magmatism is entirely the province of later convective melting under the crater. The dome, with a horizontal temperature gradient across its rim, provides a driving force for convection (Fig. 3). Rising mantle material in the dome’s center may melt adiabatically. In terrestrial models, stress-dependent rheology is important in modeling this process: instabilities eventually drop off the rim of the lithospheric dome, and nascent instabilities quickly build up stress while deforming, thereby lowering their viscosity and enabling downward flow [23]. This feedback process speeds the fall of instabilities significantly when compared to simple temperature-dependent rheology. On the Moon, velocities and stresses are low. For the temperature dependence used in the viscosity laws of this study, no instabilities drop off the crater rims, and slow convection continues long after it would have ceased on the Earth. The highest extent of mantle melting, in the case of the largest impact and the longest convective movement, is 20%. Shallower lithospheric domes create less driving force for convection in the mantle, and therefore lead to less convective melt. If the height of the lithospheric dome is reduced from 0.15R to 0.05R, then melt volumes are reduced by about 60%.

Although convective melts total only 1% or 2% of the total melt produced by the cratering process, convective melting continues for many millions, and in some cases, hundreds of millions of years after initial impact (Figs. 4 and 5). When the mantle potential temperature is 1350 °C, the largest crater modeled (R = 300 km) produces melt for about 150
Ma after impact. When the mantle potential temperature is 1450 \(^{\circ}\)C, however, melting continues for about 330 Ma after impact.

The maximum mantle velocity in any of the numerical convection runs is about 1 mm/year. In the case of the largest impact and the longest convective movement, some mantle materials are melted by 20% at the conclusion of the process. This is the maximum melting extent.

The impact-generated mafic mantle melts originate from between 190 and 390 km depth when the mantle potential temperature is 1350 \(^{\circ}\)C, and from 150 and 560 km for a mantle potential temperature of 1450 \(^{\circ}\)C (Fig. 2). Fig. 5 shows that volumes of mare basalt are closely approximated by this model, assuming that 10% of the melt produced erupts.

None of the numerical convection models mobilized any solid KREEP or high-Ti cumulate. We conclude that these materials, if they remained at 60–80 km depth at the time of the late heavy bombardment, were incorporated into mare basalts and picritic glasses only through assimilation by existing magmas, or by mixing after having been melted by the initial impact.

4. Discussion

The data of Nyquist and Shih show that the record of eruptions peaked in the Imbrian Period (3.85–3.2 Ga) and lasted to at least 3.0 Ga [24]. Given the range of time over which the large basins formed, melt production in these models (up to 350 Ma for one basin) would last from before 4.0 Ga to after 3.5 Ga. Though by far the largest volume of basalt is produced by the initial in-situ decompression melting, later eruptions will cover the initial eruptions, so dates are more likely to be obtained from these later eruptions.

Under the impact basins the selenotherm is perturbed significantly, due to isostatic uplift and rise of the hot mantle (Fig. 2). The model depth range of melting for the 1450 \(^{\circ}\)C mantle, from 150 to 560 km, is consistent with the depth of formation of all but the very shallowest mare basalts. Superheat from deeper magmas, and possibly heat of fusion from crystallization of the same, produces adequate heat to create the shallowest mare basalts, if they were not formed by the initial heat of impact.

South Pole Aitken, the oldest and largest basin on the far side of the moon, contains a relatively small amount of basalt fill [25,26]. Because our preferred mantle potential temperatures for the Moon at this stage in its evolution place shallow mantle cumulates above their solidii even before impact, there are only a few simple explanations for why South Pole-Aiken would contain relatively little volcanism. Either the mantle was composed of different material with a higher solidus; or the lithosphere was thick enough on the far side to keep the mantle beneath its solidus; or the temperature of the mantle was lower on the far side. It seems unlikely that lunar mantle at the depths of green glass genesis on the far side would be significantly different compositionally from the near side, if the mantle formed by crystallization from a magma ocean, which would presumably be well-mixed until late stages of crystallization. Because of the shallow pressure gradient on the Moon, however, the late stages of lunar magma ocean crystallization would occur with a high fraction of solid present, likely higher than the critical crystal fraction, thus preventing large-scale homogenization of the uppermost mantle, perhaps allowing only interstitial fluid flow. This may be a reasonable mechanism for creating a thicker lithosphere on the far side of the Moon than on the near side. Schmitt [27] suggested that the absence of mare basalts at South Pole-Aitken may be related to the region’s depletion in heat-producing KREEP materials. Other studies have also come to the conclusion that there were large-scale inhomogeneities in the thermal structure of the Moon’s upper mantle before about 3.5
Ga [2,9,28]. We therefore suggest that the lithosphere may have been thicker on the far side, as further evidenced by the greater thickness of crust on the far side, and that the near side shallow mantle may have been preferentially warmed by radiogenic sources. If the mantle potential temperature on the far side were \(1350\) °C, then a lithospheric thickness of 450 km would prevent South Pole-Aitken from filling with basalts.

It is also possible that the impact at South Pole Aitken did cause some degree of decompression melting as suggested by this model. The Late Heavy Bombardment that saturated the lunar crust between the suggested age of South Pole Aitken (4.2 Ga) and the end of heavy bombardment (several hundred million years later) may have brecciated and ejected most of the mare basalts fill of the crater, leaving only the noritic roots of the magmatism.

The implications of this model are that the deepest melts from both the in-situ and convective processes are picritic glasses. The mare basalts were formed either by simply melting the lithosphere (the favored scenario that came from study of the returned Apollo samples), or they are the result of picritic glass-like magmas rising from the deeper lunar interior and interacting with the shallow lithosphere. If the KREEP and high-Ti cumulates still lay between 60 and 80 km depth, they would have been melted by initial shock impact in craters with excavation radii larger than 150 km. The resulting dense melt may have percolated downward in the moon [11] to make a hybrid mantle, or it may have mixed with rising decompression mantle melts to create KREEP and high-Ti mare basalts. Similarly, aluminous upper lithosphere and crust may have melted and mixed with rising mafic decompression melts to create the high-alumina mare basalts. The order of eruption, with the most primitive, lowest titanium magmas last, is analogous to Earth processes in which early injections of magma can heat, assimilate and mix with crustal components (here, lithospheric components), and the last eruptions come through a clean magma conduit and so retain their primitive, pristine compositions [29,30].

5. Conclusions

The late heavy bombardment is sufficient to create the volumes and durations of mare basalt eruptions seen on the moon. The process of impact-induced volcanism can create a variety of magmatic compositions, probably including the aluminous, KREEP, high-Ti, and low-Ti glass and basalt compositions. The model implies that deep magmas are picritic glasses, and that mare basalts originated in the lithosphere, which was heated and partly melted by impact processes. Only the shallowest mare basalts could not be produced directly by this process; they require the superheat and possibly heat of fusion of stalled, cooling magmas from deeper in the moon.

Acknowledgements

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