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Key Points:

- To understand the localized long-lasting volcanism of the Moon, we developed a numerical model of magmatism in the convecting mantle
- Partially molten plumes ascend from the deep mantle to the uppermost level to cause the calculated volcanic activity
- Localized radioactive enrichment in the uppermost mantle maintains the plume activity for billions of years to cause volcanism

Supporting Information:

Supporting Information may be found in the online version of this article.

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Long-Lasting Volcanism of the Moon Aided by the Switch in Dominant Mechanisms of Magma Ascent: Role of Localized Radioactive Enrichment in a Numerical Model of Magmatism and Mantle Convection

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Abstract Significant volcanic activity continued for billions of years since 3.5–4 Gyr ago in the Procellarum KREEP Terrane (PKT) of the Moon, but not so significant outside the PKT. To understand this volcanic history, we developed a 2-D numerical model of magmatism and mantle convection; the effects of the PKT on lunar evolution are considered by initially imposing a region of localized radioactive enrichment. The calculated volcanism is driven by two different mechanisms. Early volcanism occurs when magma generated in the deep mantle by internal heating ascends to the surface as partially molten plumes. The basaltic blobs in the uppermost mantle formed by this magmatism then sink into the deep mantle, triggering further plumes that cause a resurgence of volcanism in its later history. Our model suggests that later plumes caused by sinking basaltic blobs are the cause of the long-lasting volcanism in the PKT.

Plain Language Summary Geological observations of the Moon have revealed that active volcanism took place at 3–4 Gyr ago and continued for billions of years in the Procellarum KREEP Terrane (PKT), a region enriched in radioactive heat-producing elements (HPEs). Here, we calculated a 2-D model of magmatism and mantle convection, considering a locally HPE-enriched area (EA) beneath the crust to model this localized long-lasting volcanism in the PKT. Localized volcanism continues for billions of years owing to ascending partially molten plumes in our model. However, the mechanism of ascent of plumes changed with time. The early volcanic activity is caused by the ascent of partially molten plumes that are generated by strong internal heating at the base of the mantle. These plumes are globally distributed. The compositionally dense basaltic blobs in the uppermost mantle outside the EA, formed by the earlier magmatism, sink into the deep mantle, triggering counterflows that induce further partially molten plumes. These plumes cause volcanism above the EA from 1.2 to 2.8 Gyr. Our model suggests that the volcanism in the later stage caused by sinking basaltic blobs plays a crucial role in the long-lasting volcanism in the PKT.

1. Introduction

Understanding the volcanism of the Moon has been a long-standing issue in studies of lunar mantle evolution (e.g., Arai et al., 2008; Haupt et al., 2024; Head et al., 2023; Kirk & Stevenson, 1989; Solomon & Chaiken, 1976). Geological observations have revealed that volcanism has been particularly active in the Procellarum KREEP Terrane (PKT) located on the nearside, which is enriched in radioactive heat-producing elements (HPEs) (e.g., Jolliff et al., 2000; Kamata et al., 2013; Lawrence et al., 2000; Wieczorek & Phillips, 2000). Volcanism has occurred in other regions but has been less active (e.g., Shearer et al., 2006). In the PKT, volcanic activity continued for billions of years (e.g., Cho et al., 2012; Giguere et al., 2022; Kato et al., 2017; Morota et al., 2011) with a peak at 3.5–4 Gyr ago (e.g., Che et al., 2021; Hiesinger et al., 2000, 2003; Hurwitz et al., 2013; Nagaoka et al., 2020; Su et al., 2022; Whitten & Head, 2015). This period of active volcanism coincides with the time of the radial expansion of the Moon: the Moon expanded globally by 0.5–5 km until around 3.8 Gyr ago (e.g., Andrews-Hanna et al., 2013, 2014; Liang & Andrews-Hanna, 2022; Nishiyama et al., 2024; Sawada et al., 2016) and then contracted globally over time (Frueh et al., 2023; Yue et al., 2017), by around 1 km or less in the past 1 Gyr (e.g.,

Clark et al., 2017; Klimczak, 2015; Watters et al., 2010, 2015). To understand the long-lasting history of lunar volcanism, which mostly occurs in the PKT, we used a 2-D polar rectangular model of magmatism in the convecting lunar mantle (U et al., 2023b).

Various numerical models have been developed to account for the long-lasting volcanism of the Moon (e.g., Breuer & Moore, 2015; Solomon & Head, 1979; Wilson & Head, 2003). Some numerical models, in which the surface is covered by the HPE-enriched crust or a regolith layer as a blanket layer, suggest that magma persists in the uppermost mantle for 1–2 Gyr (e.g., Konrad & Spohn, 1997; Spohn et al., 2001; Ziethe et al., 2009). This partially molten region, however, extends globally and is unlikely to have caused the localized volcanism at the PKT. Some earlier researchers suggest that the localized volcanism is caused by one or some hot plumes growing from a layer of ilmenite-bearing cumulates (IBC) enriched in HPEs that developed on the core-mantle boundary (CMB) in the lunar early history (e.g., Stegman et al., 2003; de Vries et al., 2010; W. Zhang, Zhang, & Li, 2022; N. Zhang et al., 2013a, 2017; W. Zhang et al., 2023) owing to the magma ocean and mantle overturn (e.g., Hess & Parmentier, 1995; Ringwood & Kesson, 1976). Although the plume enriched in the IBC component ascends by thermal buoyancy in these models, it is unclear if the large excess density of the IBC component, as expected from earlier models of mantle overturn, allows the ascent of such a compositionally dense plume in the Moon (e.g., Le Bars & Davaille, 2004; H. Li et al., 2019; Yu et al., 2019; Zhao et al., 2019). On the other hand, Wiczorek and Phillips (2000) and Laneuville et al. (2013, 2018) assumed a locally HPE-enriched area (called EA hereinafter) at the top of the mantle in the initial condition and found that the area remained partially molten for more than 3 Gyr. Various mechanisms have been suggested for the EA formation: differences in crystallization rates between the hemispheres of the lunar magma ocean (Loper & Werner, 2002; Wasson & Warren, 1980); the residual layer enriched in KREEP (K, rare earth elements, and P-rich material) after the mantle overturn (e.g., Parmentier et al., 2002; N. Zhang, Ding, et al., 2022); impact-induced convection (e.g., Rolf et al., 2017; Zhu et al., 2019; N. Zhang, Ding, et al., 2022; Jones et al., 2022). In the models of the subsequent thermal evolution of the mantle, however, the extraction of HPEs from the partially molten area by segregating magma (Cassen et al., 1979; Cassen & Reynolds, 1973) is neglected, and HPEs remain in the uppermost mantle throughout the calculated 3 Gyr. Ogawa (2018a) found that partially molten regions solidify within 2 Gyr when HPE-extraction by magmatism is considered.

To understand the mantle evolution of the Moon, constrained by its volcanic history as well as radial expansion/contraction history, we extend the 2-D model of magmatism and mantle convection that we have developed (U et al., 2023b). In this model, magma is generated by decompression melting and internal heating, then migrates upward in partially molten regions as a permeable flow through the coexisting matrix; thermal, compositional, and melt buoyancy drive mantle convection in a Newtonian fluid with temperature-dependent viscosity; heat, mass, and incompatible HPEs are transported by magma. In our earlier study, we initiated the calculation with a compositionally stratified mantle, enriched with the IBC component at the base of the mantle (U et al., 2023b) following earlier studies of the magma ocean: during the last stage of crystallization of the lunar magma ocean, a dense IBC layer enriched in KREEP develops at the top of the mantle and subsequently sinks to the CMB, a process called mantle overturn (e.g., Alley & Parmentier, 1998; Boukaré et al., 2018; Hess & Parmentier, 1995). In addition to this initial stratification, we also consider a localized enrichment of HPEs beneath the crust to model the PKT and investigate whether this EA induces a localized volcanism similar to that observed at the PKT (Figure 1a).

2. Model Descriptions

We used a 2-D annular model of mantle convection and magmatism, developed by U et al. (2023b). The finite difference numerical code calculates the energy, mass and momentum conservation equations of mantle convection in $R = [(r, \theta) | 385 \text{ km} \leq r \leq 1735 \text{ km}, 0 \leq \theta \leq \pi]$ on a mesh with 128 (radial direction) times 256 (lateral direction) mesh points, under the Boussinesq approximation (See Text S1-1 in Supporting Information S1 for the detail of the basic equations). To model the blanket effect of the crust (Ziethe et al., 2009), we reduced thermal diffusivity by a factor of two within a 35 km-thick blanket layer at the top of the mantle. The core is modeled as a heat bath with a uniform temperature (see Equations S20, S21 in Text S1-1 in Supporting Information S1). The vertical sidewalls are insulated, while the surface boundary is fixed at 270 K; all boundaries are shear stress-free and impermeable to both magma and matrix.

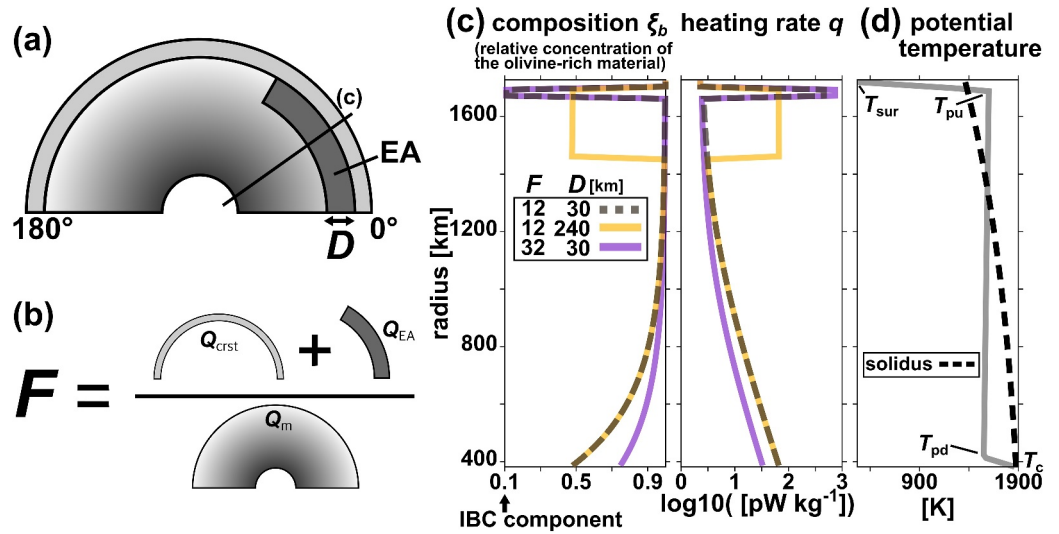


Figure 1. (a, b) An illustration of the initial distribution of heat-producing elements (HPEs). The distribution is laterally uniform except for the enriched area “EA.” In (b) F is the ratio of the total amount of HPEs in the crust Q_{crst} and the EA Q_{EA} to that in the mantle Q_m (see Texts S1-2 in Supporting Information S1). Also shown are (c) the initial distribution of the composition and heating rate along the line in (a); (d) the initial distribution of the potential temperature ($T_{\text{sur}} = 270$ K, $T_{\text{pu}} = 1,600$ K, $T_{\text{pd}} = 1,550$ K, and $T_c = 1,875$ K are assumed in this figure).

The convecting material is a binary eutectic system ($A_2B_{1-\xi}$) composed of an olivine-rich material (A_1B_0) with a density of $3,300 \text{ kg m}^{-3}$ and an IBC material (A_0B_1) with a density of $3,745 \text{ kg m}^{-3}$ (e.g., Elkins-Tanton et al., 2011; Liang et al., 2024; Snyder et al., 1992). The concentration of the olivine component is denoted by ξ , then Equation 1 below would give the concentration of the olivine-rich material in the partially molten mantle. The eutectic composition is at $\xi_e = 0.1$ ($A_{0.1}B_{0.9}$). We adopted this simple binary material as a model of mantle materials to mimic variation in density distribution in the mantle caused by segregation of compositionally dense basaltic materials by magmatism; a detailed petrologic simulation of the mantle is beyond the scope of this study. The bulk composition ξ , that is, the relative concentration of olivine-rich material, is calculated from the composition of the solid phase ξ_s and that of the liquid phase ξ_l written as

$$\xi = (1 - \phi)\xi_s + \phi\xi_l, \quad (1)$$

where ϕ is the melt-content. The solidus and liquidus temperatures are calculated from the phase diagram assumed for the binary eutectic system (see Equations S9–S11 and S17–S19 in Text S1-1 in Supporting Information S1). The convecting material contains HPEs that decay with time. Magmatism is modeled by the generation and upward migration of magma as a permeable flow (e.g., McKenzie, 1984; Miller et al., 2014). The flow is driven by the buoyancy of magma, and the relative velocity between the velocity of melt $\mathbf{u}$ and that of matrix $\mathbf{U}$ is calculated from

$$\mathbf{u}^* - \mathbf{U}^* = -Mg^* \frac{\phi^2}{\phi_0^3} (\rho_s^* - \rho_l^*) \mathbf{e}_r, \quad (2)$$

where the asterisks stand for normalized quantities; $\mathbf{e}_r$ is the unit vector in the radial direction. $\rho_s - \rho_l$ is the density difference between solid and melt phases calculated as

$$\rho_s^* - \rho_l^* = \beta(\xi_l - \xi_s) + \frac{\Delta V_l}{V_0} [1 + \beta(1 - \xi_l)] \quad (3)$$

(See Equations S6 and S7 in Supporting Information S1). Here, g is the gravitational acceleration; $\phi_0 = 0.05$ is the reference melt-content; $\Delta V_l/V_0$ is the amount of density reduction (see Equation S11 in Text S1-1 in Supporting Information S1). We assumed the reference permeability M to be

$$M \equiv \frac{k_{\phi_0} \rho_0 g_0 L}{\kappa \eta_{\text{melt}}} \simeq 100, \quad (4)$$

where the parameter values are listed in Table S1 in Supporting Information S1. $\beta = 0.135$ is a constant that expresses the sensitivity of density to the bulk composition (see Text S1-1 in Supporting Information S1). In most cases, we assumed that the crustal density is higher than the density of magma for simplicity. However, the density of the lunar crust is lower than that of basaltic magma (e.g., Head & Wilson, 2017, 2020; Morota et al., 2009; Taguchi et al., 2017; Wilson & Head, 2017), and Lourenço et al. (2018) suggest that the thermal history of the Moon can substantially depend on the crustal density inversion. To clarify the effect of the density inversion, we increased the value of β to 0.253 in the blanket layer in a case (Case crst-F12-D30): the crustal density is $2,550 \text{ kg m}^{-3}$ (Wieczorek et al., 2013) which is lower than the density of the basaltic magma at this value of β .

To simulate the localized enrichment of HPEs in the mantle beneath the PKT, we assumed the EA in $[(r, \theta) | 1700 - D \text{ km} \leq r \leq 1700 \text{ km}, 0 \leq \theta \leq 1/3 \pi]$ that is enriched in HPEs and the IBC component, as illustrated in Figure 1a. We examined the dependence of calculated volcanism on the thickness of the EA “ D ” and on the ratio of the total amount of HPEs in the crust and the EA to the mantle “ F ” (Figures 1a and 1b); note that HPE-content in the crust is constant for simplicity (see Text S1-2 in Supporting Information S1). The searched range of “ D ” (10–240 km) is estimated from the thickness of the KREEP layer that remains after the mantle overturn (e.g., Charlier et al., 2018; Laneuville et al., 2013; Schwinger & Breuer, 2022; N. Zhang, Ding, et al., 2022). We also take the searched range of “ F ” (4–32) from previous estimates of Th concentrations in the enriched area (e.g., Laneuville et al., 2018; Wieczorek & Phillips, 2000); the upper value of the Th concentration is around the average surface composition shown by orbital gamma-ray spectroscopy data (e.g., Taylor & Wieczorek, 2014).

In the initial condition, we assumed that the mantle is compositionally stratified to simulate the stratification of the lunar mantle inferred from previous studies of the magma ocean and mantle overturn (e.g., Hess & Parmentier, 1995; Zhao et al., 2019). The contents of HPEs and the IBC component increase exponentially with depth while the content of the olivine-rich material decreases with depth (Figure 1c). The average of the internal heating rate over the entire mantle is fixed at $q_0 = 14.7 \text{ pW kg}^{-1}$ at 4.4 Gyr ago as estimated from Table S2 in Text S1-1 in Supporting Information S1 for all the cases listed in Table S3 in Supporting Information S1. The potential temperature distribution in the initial condition (Figure 1d) is also taken from earlier models of the lunar mantle overturn (e.g., Boukaré et al., 2018; Yu et al., 2019) and the evolution model (U et al., 2022). The details of the initial condition are described in Text S1-2 in Supporting Information S1. Note that the uppermost mantle is partially molten in the initial condition (Figure 1d). This region persists for around 600 Myr (see Figure 2 below). However, the melt-content is small, up to only 1%, and this region does not affect the overall dynamics of the calculated mantle as we will describe below.

3. Results

In Figures 2 and 3 as well as the animation (Movie S1), we present the evolution of the mantle calculated in Case F12-D30 where $F = 12$ and $D = 30 \text{ km}$ (see also figures in Supporting Information S1).

Magma is generated in the deep mantle owing to the strong internal heating assumed in the initial condition (Figure 1c) and migrates upward as partially molten plumes to reach the uppermost mantle within the first 0.7 Gyr (Figures 2b and 2c; see also Figure S1a in Supporting Information S1); melt buoyancy is the main driving force of this flow. Magma rises to the depth level of 100 km beneath the EA in $[0 \leq \theta \leq 1/3 \pi]$, which is shallower than beneath the area outside the EA (Figures 3a and 3b; see also Figures S1b and S1c in Supporting Information S1). This is because the lithosphere is thinned beneath the EA owing to the strong heating by HPEs in the EA (Figures 2a and 2c; see also Figure S2 in Supporting Information S1). The magma transports HPEs and the basaltic component to the uppermost mantle, forming basaltic blobs (see the arrows in Figure 2d for 0.76 Gyr). A deeper part of the compositionally dense basaltic blobs eventually founders to the CMB as magma accumulates and the blobs grow (see the arrow in Figure 2d for 1.12 Gyr; see also the description in Text S2-1 in Supporting Information S1). The descending basaltic blobs trigger further ascent of partially molten plumes from the deep mantle driven by melt buoyancy, as depicted by the arrows in Figure 2b. Accordingly, the upward magma flux

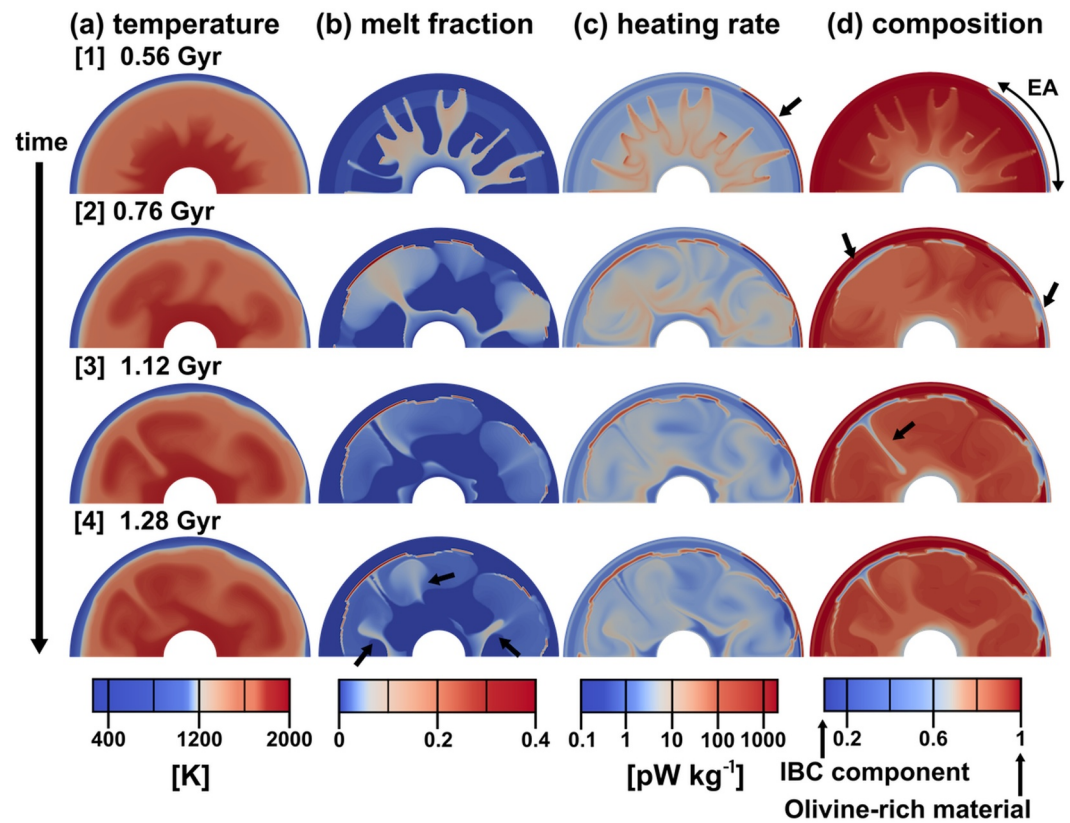


Figure 2. Snapshots of the distributions of (a) temperature, (b) melt-content, (c) internal heating rate, and (d) bulk composition calculated for the reference case ($F = 12$ and $D = 30$ km; Case F12-D30). The elapsed times are indicated in the figure. In (d), the blue color stands for the basaltic composition enriched in the IBC component, while the red color stands for the olivine-rich end-member. The numbers [1]–[4] correspond to those of Figure 3b.

beneath the EA becomes significant again from 1.2 to 2.8 Gyr (Figure 3b). The resurgence of magma flux is not found in our earlier models that are devoid of the EA (see the gray dotted line in Figure 3b). In contrast, the magma flux is negligibly small throughout the calculated history outside the EA due to the thicker lithosphere (Figures 3a and 3b). After 2.8 Gyr, the partially molten region in the mantle shrinks with time due to the decay of HPEs but persists until 4.4 Gyr (Movie S1 as well as Figure S1d in Supporting Information S1). From further numerical experiments with various values of D in the range of 10–240 km and F in the range of 4–32, we found that the localized volcanism continues for more than 1.5 Gyr in the cases with $D < 120$ km and $8 < F < 24$ (see Text S2-2 in Supporting Information S1 including Table S3 and Figures S3, S4 in Supporting Information S1).

The resurgence of magma flux beneath the EA observed in the reference case is more pronounced in Case crst-F12-D30 where the crustal density is lower than the density of the basaltic magma (see Figures 3c, 3d and Figure S5 in Supporting Information S1). The upward magma flux shown in Figure 3c is substantial beneath the entire EA until 3.7 Gyr, and magma enriched in HPEs remains along the crust-mantle boundary even at as late as 3.5 Gyr (Figure S5 in Supporting Information S1). The resurgence of magma flux is so significant in crst-F12-D30 because the lower density of the crust prevents magma from transporting HPEs to the surface; HPEs remain beneath the crust to keep the lithosphere beneath the EA thinner compared with that in the reference case. The thinner lithosphere beneath the EA allows the magma generated by the counterflows to rise higher than that in the reference case; the magma flux increases locally several times when the heads of partially molten plumes ascend to the base of the crust (see Figures 3c and 3d; see also Figure S5 in Supporting Information S1 and Movie S2).

We found that strong internal heating and compositionally high density in the EA, as well as the existence of an initial mantle stratification, are important for the resurgence of magma flux beneath the EA (see Figures S6–S9 in Supporting Information S1). The resurgence of magma flux is not observed in Case noHeat-F12-D30 where the enrichment of HPEs in the EA is not considered (Figures S6a and S7 in Supporting Information S1). The

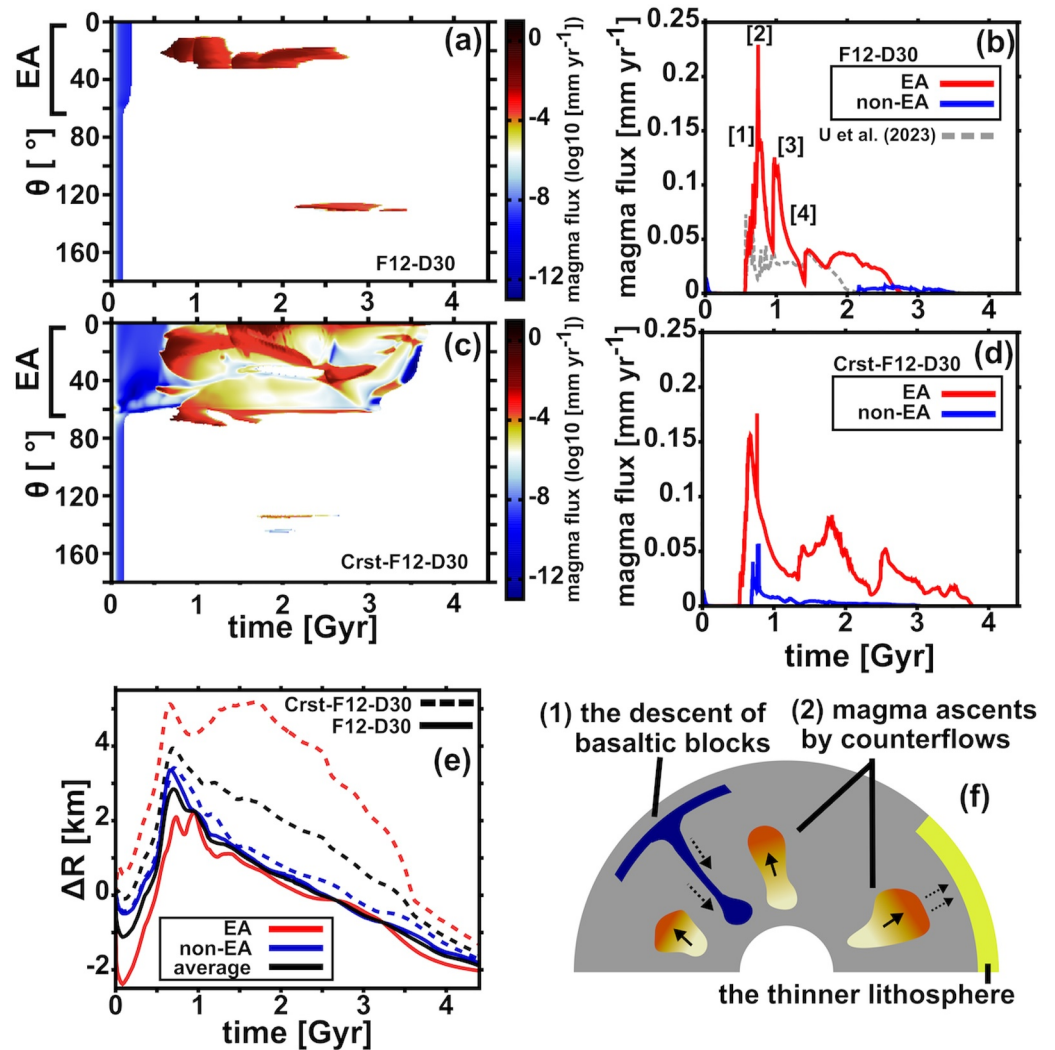


Figure 3. (a) The distribution of the upward flux of magma that passes upward through the depth level of 100 km and (b) the average flux of magma around the EA (the red line) and outside the EA (the blue line) at the depth level of 100 km in Case F12-D30 (the reference case). Also shown are (c and d) the magma flux calculated in Case crst-F12-D30 where the crust is less dense than basaltic magma; (e) the history of radius change ΔR calculated by Equation S24 in Text S1-3 in Supporting Information S1; (f) a schematic illustration of long-lasting volcanism with a couple of peaks in the PKT calculated in the model. In (b, d, and e), the blue lines and red lines stand for the average values in $[0 \leq \theta \leq 1/3 \pi]$ (beneath the EA) and in $[1/3 \pi < \theta \leq \pi]$ (outside the EA), respectively, while the black lines stand for the total average values in $[0 \leq \theta \leq \pi]$. In (b), the gray line shows the magma flux in our earlier model (U et al., 2023b) where the EA is not considered (other parameters are the same as those assumed in Case F12-D30); the numbers [1]–[4] correspond to those of Figure 2.

resurgence occurs only briefly in Case noBa-F12-D30 where the compositional high density in the EA is not considered (Figures S6b and S8 in Supporting Information S1). This is due to the smaller density difference between the mantle and magma $\rho_s - \rho_l$ in the EA compared to the reference case, resulting in a lower permeable flow (Equations 2 and 3). On the other hand, magma flux at the depth level of 100 km is not observed since 0.5 Gyr after the start of the calculation in Case noStra-F12-D30 where the HPEs and the composition ξ are vertically uniform in the initial mantle (Figures S6c and S9 in Supporting Information S1). The deep mantle is not enriched in HPEs in this case, and the amount of generated magma is too small to let plumes ascend to the uppermost mantle. Magmatism caused by partially molten plumes does not occur at the Rayleigh number $Ra \leq 2 \times 10^5$ (see Equation S3 and Text S2-4 in Supporting Information S1 which includes Movie S3 and Figures S10, S11 in Supporting Information S1) but continues for billions of years beneath the EA at $Ra \simeq 2 \times 10^6$ (Figures 3a and 3b). In contrast, the overall features of mantle evolution are not influenced by the

value of the sensitivity of viscosity to temperature E_T (see Equation S4 and Text S2-4 in Supporting Information S1 including Figure S12 in Supporting Information S1). These parameter dependencies of the numerical results are the same as those shown in U et al. (2023b).

Figure 3e shows that the Moon expands globally by more than 3 km for the first 0.7 Gyr of the calculated history and then contracts at a rate of around -1 km Gyr^{-1} in the past 1 Gyr in the reference case. The decomposition of radius change ΔR into the component caused by melting/solidification and by thermal expansion/contraction shows that the expansion is mainly caused by melting, while the contraction is mainly caused by thermal contraction (Figure S13 in Supporting Information S1). The overall feature of the radius change is not affected by the crustal density (see the black dotted line in Figure 3e). However, there is a drop in the radius beneath the EA ΔR_{EA} at around 3.7 Gyr due to the solidification of the mantle in Case crst-F12-D30 (see Figure S14 in Supporting Information S1). This is because a partially molten region that persists until that time along the crust-mantle boundary beneath the EA solidifies (Figure 3c; see also Figure S5 in Supporting Information S1).

4. Discussions

Figure 3f illustrates how an EA causes a long-lasting volcanism in our models. In the earlier stage of mantle evolution, magma is generated in the deep mantle by internal heating and ascends to the uppermost mantle as partially molten plumes. In the area beneath the EA, magma rises to a shallower depth than in the area outside the EA because the lithosphere there is thinner owing to the strong internal heating assumed in the initial condition. This magmatism results in the formation of basaltic blobs beneath the lithosphere. The basaltic blobs outside the EA eventually sink to the CMB, triggering further plume activity that causes a long-lasting volcanism.

4.1. Comparison With Earlier Models of Lunar Volcanic History

A comparison with earlier mantle convection models of lunar volcanic history, which mostly occurs in the PKT and its surrounding area, shows the crucial role of melt buoyancy in our models. Some models suggest that localized volcanism is caused by thermal plumes from an HPE-enriched IBC layer on the CMB (e.g., N. Zhang et al., 2013b; Zhong et al., 2000). In these models, however, the compositional density contrast between the basal IBC layer and the overlying olivine-rich mantle is substantially less than 160 kg m^{-3} (see Figure 6 in H. Li et al. (2019)), the value suggested from some recent mantle overturn models (e.g., Maurice et al., 2024; Yu et al., 2019; Zhao et al., 2019). At this density contrast, the basal layer remains convectively stable and does not rise as a plume by thermal buoyancy alone (de Vries et al., 2010; Le Bars & Davaille, 2004). In contrast, a large fraction, around 60%, of the IBC component in the deep mantle arises as partially molten plumes mostly driven by melt buoyancy in our models (Figure 2) although the density contrast is more than 200 kg m^{-3} in our reference case (Figure 1c).

Our models also suggest the reason why volcanism continues so long in the PKT. In some of earlier models, a locally HPE-enriched area at the top of the mantle assumed in the initial condition remains partially molten for more than 3 Gyr (Laneuville et al., 2013, 2018; Wiczeorek & Phillips, 2000). HPE-extraction from the enriched area by magmatism is, however, neglected in these models as we discussed before. In our models, magmatism continues long despite that HPE-transport by magma is taken into account, because the basaltic blobs formed by early magmatism descend to the CMB to trigger further ascent of partially molten plumes (Figures 2, 3a, and 3f). Besides, the melting history of the uppermost mantle depends on the crustal density. In Case crst-F12-D30 where the crustal density is lower than the density of the basaltic magma, HPEs are not extracted to the surface but remain at the base of the crust (see Figure S5 in Supporting Information S1). The HPEs at the crust-mantle boundary keep the lithosphere thin enough for revived plume activity to cause volcanism. The resurgence of magma flux is so significant beneath the entire EA for more than 3.5 Gyr due to the plume activity (Figures 3c and 3d). By comparing magma flux under varying initial conditions (Figures S6–S9 in Supporting Information S1), our simulations also show that both the initial mantle stratification and the EA beneath the crust play crucial roles in the resurgence of volcanism after 1 Gyr. The resurgence of the magma flux is observed in all of the cases with $Ra \approx 2 \times 10^6$ regardless of the values of E_T (Figures S10–S12 in Supporting Information S1 as well as Movie S3) because the Rayleigh number Ra is high enough for partially molten plumes to grow. Note that a detailed discussion of volcanic history is possible in our models because we calculate the magma flux directly, rather than estimating it from the distribution of magma in the mantle, as is done in earlier models (e.g., Konrad & Spohn, 1997; Solomon & Toksöz, 1973; U et al., 2022; Wood, 1972).

4.2. Comparison With the Observed Features of the Moon

The magma flux beneath the EA shown in our reference case is consistent with the observed history of volcanism in the PKT. Magma generated at depth by strong internal heating rises (Figure 2 for 0.56 and 0.76 Gyr), causing the peak of volcanism at 3.5–4 Gyr ago (e.g., Hiesinger et al., 2000; Whitten & Head, 2015). The resurgence of volcanism triggered by the descending basaltic blobs is consistent with the observed long-lasting volcanic history in the PKT (e.g., Cho et al., 2012; Tian et al., 2023). The early expansion and later contraction of the Moon in the reference case (Figure 3e) are also consistent with those suggested from gravity gradiometry data and tectonic features (e.g., Andrews-Hanna et al., 2013; Matsuyama et al., 2021; van der Bogert et al., 2018). The plumes triggered by descending basaltic blobs are depleted in HPEs compared to the plumes that ascend in the early history (see the frames of 0.76 and 1.28 Gyr in Figures 2c and 2d; see also the animation). The plume activity in the later history can account for the HPE-depleted young basalts observed in the PKT (Che et al., 2021; Q.-L. Li et al., 2021; Su et al., 2022). It is, however, difficult to account for the temporal trend of increasing Ti content in the HPE-depleted mare basalts in the PKT after around 2.3 Gyr ago (Kato et al., 2017; Sato et al., 2017; D. Zhang, Su, et al., 2022). This trend may be due to the remelting of basaltic blobs in the uppermost mantle by the plumes that rise later (Figure 2). To understand this trend clearly, it is necessary to model magma evolution carefully, considering fractional crystallization (e.g., Luo et al., 2023).

Further calculations in a 3-D spherical mantle are important to more quantitatively predict the history of magma flux, radial expansion/contraction, and thermal history of the Moon and to compare the predictions with such observations as the temperature distribution in the Moon (e.g., Khan et al., 2006, 2014). As already found by Guerrero et al. (2018) for the Moon, 2-D polar rectangular models of mantle convection tend to predict higher average mantle temperatures compared to 3-D spherical models, especially when the core size is small. In the reference case, the mid-mantle temperature at 4.4 Gyr is about 100 K higher than current lunar estimates (Figure S1d in Supporting Information S1), which may result from the geometry of the convecting vessel.

5. Conclusions

To understand the long-lasting localized volcanism for billions of years in the PKT (the Procellarum KREEP terrane), we numerically simulated a 2-D polar rectangular model of magmatism and mantle convection where a localized heat-producing elements (HPEs) enrichment beneath the crust, called enriched area (EA) is considered to model the PKT in the initial mantle stratification (Figure 1).

Our simulations show that the localized volcanism, which continued for billions of years, is induced by ascent of partially molten plumes driven by two mechanisms. The early volcanism occurs as partially molten plumes ascend from the deep mantle owing to internal heating there (Figures 2 and 3b). Magma ascends to the uppermost mantle beneath the EA because the lithosphere there is thinned by the strong internal heating in the EA. The later activity of volcanism, in contrast, is caused by the counterflows of sinking basaltic blobs that are formed beneath the lithospheric lid outside the EA by the earlier magmatism; a deeper part of the compositionally dense basaltic blobs sinks into the deep mantle, triggering further magma ascent beneath the EA from 1.2 to 2.8 Gyr (Figures 3b and 3f). We found that the mantle evolves as illustrated in Figure 3f when the thickness of the EA is $D < 120$ km and the ratio of the total amount of HPEs in the crust and the EA to the mantle is $8 < F < 24$ (see Figures S3, S4, and Table S3 in Supporting Information S1). Our models suggest that material transport and melt buoyancy by magmatism play critical roles in the lunar mantle evolution. In particular, the resurgence of volcanism in the later stage is important to account for the long-lasting volcanism in the PKT.

Data Availability Statement

The original data in all cases plotted in Table S3 in Supporting Information S1 and the numerical code used in this work are found at U et al. (2023a, 2024), respectively.

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