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Duration and extent of lunar volcanism: Comparison of 3D convection models to mare basalt ages

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ABSTRACT

It is widely accepted that lunar volcanism started before the emplacement of the mare fills ($\approx 3.1-3.9$ Ga b.p.) and lasted for probably more than 3.0 Ga. While the early volcanic activity is relatively easy to understand from a thermal point of view, the late stages of volcanism are harder to explain, because a relatively small body like the Earth's Moon is expected to cool rapidly and any molten layer in the interior should solidify rather quickly. We present several thermal evolution models, in which we varied the boundary conditions at the model surface in order to evaluate the influence on the extent and lifetime of a molten layer in the lunar interior. To investigate the influence of a top insulating layer we used a fully three-dimensional spherical shell convection code for the modelling of the lunar thermal history. In all our models, a partial melt zone formed nearly immediately after the simulation started (early in lunar history), consistent with the identification of lunar cryptomare and early mare basalt volcanism on the Moon. Due to the characteristic thickening of the Moon's lithosphere the melt zone solidified from above. This suggests that the source regions of volcanic rock material proceeded to increasing depth with time. The rapid growth of a massive lithosphere kept the Moon's interior warm and prevented the melt zone from fast freezing. The lifetimes of the melt zones derived from our models are consistent with basalt ages obtained from crater chronology. We conclude that an insulating megaregolith layer is sufficient to prevent the interior from fast cooling, allowing for the thermal regime necessary for the production and eruption of young lava flows in Oceanus Procellarum.

1. Introduction

Today, the giant impact hypothesis for the formation of the Moon is widely accepted (e.g., Cameron and Canup, 1998). It is assumed that the proto-Earth, which was already differentiated into an iron core and overlying silicate mantle, experienced a collision with another proto-planet, which had roughly the size of Mars (Cameron and Benz, 1991). This led to the formation of a moon, which had a relatively deep molten outer shell (magma ocean) due to heavy bombardment with impactors in the late stage of accretion (Benz et al., 1986; Canup and Esposito, 1996; Cameron, 1997; Stevenson, 1987; Weisberg and Hager, 1998; Kokubo et al., 2000).

Differentiation and density segregation of crystals from the melt led to the formation of an old low-density, plagioclase-rich (anorthositic) primary crust (Shervais and Taylor, 1986). Volcanic rocks on the Moon seem to come from great depth and are generally younger than the primary crust. The most prominent evidence for lunar volcanism is the maria (Head, 1976; Shervais and Taylor, 1986) that formed when basaltic lava flooded some of the large basins (e.g., Imbrium, Crisium, Orientale). The petrology of the Apollo mare basalt samples suggests a formation of the basalts in a depth between 150 and 500 km by partial melting (Heiken et al., 1991). Taylor (1982) suggested a re-melting process of the deeper layers of the crystallized magma ocean as a source of the mare basalts.

Radiometric ages for the basaltic rocks collected at the Apollo landing sites show that the main eruption phase occurred between 4.0 and 3.2 Ga ago (Stöffler et al., 2006). On the other hand, some basalt samples collected during the Apollo 14 mission were dated to be 4.2 Ga old (Head et al., 1978), similar to the numerous dated highland rocks. The implication is that the onset of mare basalts eruptions already occurred early in lunar history at times prior to the end of the heavy lunar bombardment (Taylor et al., 1983). However, evidence for volcanism acting within the first 300 Ma after the formation of the Moon has largely been destroyed or covered by later impact events (Head et al., 1993). The total duration of lunar volcanism is assumed to be about 3.0 Ga long, reaching well into the Eratosthenian Period (3.1 Ga

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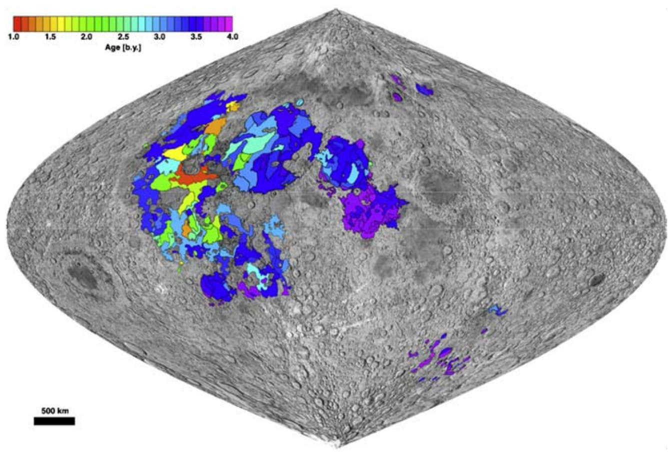

Fig. 1. Distribution of mare basalt model ages (from Hiesinger et al., 2003).

until 1.1 Ga before present) (Head and Hiesinger, 1999; Hiesinger et al., 2000, 2003) or even into the Copernican Period (Schultz and Spudis, 1999; Hiesinger et al., 2003) (Fig. 1).

Estimates of the global lava eruption rates at the surface are on the order of $10^{-2}\,km^3\,a^{-1}$ during 4.3–3.1 Ga b.p. (Basaltic Volcanism Study Project, 1981) decreasing to 10^{-5} and $10^{-4}\,km^3\,a^{-1}$ in the Eratosthenian (Basaltic Volcanism Study Project, 1981; Head and Wilson, 1992). Various authors speculated that the production rate of melt in the mantle was between 50 and 100 times higher than the surface eruption rates (Head and Wilson, 1992). It is not unlikely that magma remained in the mantle or stalled in the crust as dikes and did not reach the surface. Fig. 1 shows the distribution of model ages of lunar basalts, obtained by crater chronology by Hiesinger et al. (2003). This map indicates that the younger basalts are located mostly on the western part of the lunar near side, in an area that is characterized by large amounts of heat-producing elements, that is, the Procellanus KREEP Terrane (PKT) of Jolliff et al. (2000). This distribution could suggest that volcanism was initially widespread and later constrained to smaller regions with high concentrations of heat-producing elements or thinner crust.

Although results from geology, petrology and geochemistry revealed numerous evidence for lunar volcanism being active for a time period of nearly 3.0 Ga, there are nonetheless difficulties to explain a long-lasting molten zone within the lunar mantle (Konrad and Spohn, 1998). Compared to the other terrestrial planets the Moon is a relatively small body and is expected to cool rapidly, even if its formation was hot, as proposed by Stevenson (1987) and Weisberg and Hager (1998). Any molten layer should therefore solidify rather quickly. Hence, it depends on whether the Moon can stay warm long enough to explain volcanic eruptions as young as about 1–2 Ga. Spohn et al. (2001) showed that a molten layer in the upper lunar mantle can persist for some time, but the younger eruptions of basaltic lava were difficult to explain in these models. There are at least four factors that have influence on the cooling history of a planetary body:

(1) the amount of kinetic energy added to the body by meteorite impacts;
(2) the amount and distribution of heat-producing elements;
(3) the crustal thickness;
(4) the thickness of the megaregolith.

The latter two, the thickness of the crust and the megaregolith, are important because they serve as a thermal lid, which insulates the hot interior from cold space. The thicker the crust and the megaregolith, the longer the interior will remain hot. As the effects of crustal thickness have already been studied for Mars by Schumacher and Breuer (2006), in this paper we will investigate whether the effects of a thick megaregolith on the thermal evolution of the Moon are sufficient to keep the lunar interior molten for long enough time in order to explain the young mare basalts in Oceanus Procellarum. We present three-dimensional convection models, to investigate to what spatial and temporal extent a molten zone in the lunar interior can exist. We investigate the influence of the upper boundary, namely the lunar surface on the duration and extent of a partially molten zone in the lunar interior. Former models by Spohn et al. (2001) did not include the variation of the upper boundary temperature. While Schumacher and Breuer (2006) showed that the influence of the upper boundary temperature on the existence of the partial melt zone and crustal production rate can be significant, their models were one-dimensional parameterized convection models for Mars. Warren et al. (1991) modelled the global cooling of the Moon and large asteroids, which are covered with regolith. Their results indicate that the bodies cool significantly slower, if an insulation layer of fragmental impact debris is covering the surface. Here we present results from a three-dimensional model, which allows us to vary the surface boundary temperature in order to investigate its influence on the cooling history of the Moon. The intention is not to construct a model for the lunar crust or top insulating layer. We rather present a simple way for providing the temperature at the crust–mantle interface as an input parameter for the three-dimensional convection code. We will address the following questions: (1) Is it possible that the Moon cooled slowly enough to allow volcanic eruptions as recently as 1–2 Ga ago? (2) What is the influence of a kilometer-thick insulating layer on the cooling history of the Moon?

2. Theory and model

In this paper the lunar mantle is modelled as an internally and bottom heated, isochemical fluid in a spherical shell. The principle of this convection model is widely accepted and is used for various models of thermal evolution of terrestrial planets, e.g., the Earth (Zhang and Yuen, 1996), Mars (Harder, 2000; Breuer et al., 1998) or the Moon (Konrad and Spohn, 1997; Spohn et al., 2001). Important parameters to compute the thermal evolution are the initial temperature profile, the concentration of radioactive heat sources, material parameters and laws and boundary conditions. In this section we will introduce the used parameters and give a short overview about the numerical method and solved equations.

2.1. Outer boundary: the influence of an insulating layer

The numerical model that was used for this work focuses on the dynamical behaviour of the mantle, in particular convection. The non-convecting upper layers of the Moon are left out of the model, as is the core, in order to keep the numerical effort reasonable. Nevertheless, the topmost layers have an influence on the model, because they constitute the upper boundary condition of the modelled volume. In this section we discuss the vertical structure of these layers, their physical properties and the resulting thermal insulation they provide. Keihm and Langseth (1977a,b) studied the temperature profile within the uppermost layers, and the thermal insulation they provide for the interior. Schumacher and Breuer (2006) made the case for Mars, investigating the influence of a variable thermal conductivity of the upper layer (lithosphere) on the behaviour of convection in the underlying mantle. They found that the low thermal conductivity of a surface layer can increase the mean temperature of the convecting material below.