Early magnetic field and magmatic activity on Mars from magma ocean cumulate overturn


Brown University, Department of Geological Sciences, 324 Brook St., Providence, RI 02912, United States

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Abstract

Significant and perhaps complete melting of the young terrestrial planets is expected from their heat of accretion and core formation. The process of subsequent magma ocean fractional solidification creates a cumulate mantle unstable to gravitational overturn. Overturn should be fast (≤1 to 10 Ma) and result in increasing mantle density with depth. This stable stratification inhibits later thermal convection, preserving geochemical heterogeneities. Overturn places cold cumulates against the core–mantle boundary, which creates sufficient heat flux to drive a core dynamo, producing a brief, strong magnetic field. During overturn, hot cumulates rise from depth and melt adiabatically, creating an early crust to record this field and leaving behind mantle reservoirs with isotopic fractionations dating from the early evolution of the planet.

Keywords: Mars; magma ocean; magnetic field; cumulates; mantle heterogeneity

1. Introduction

The conversion of kinetic energy to heat during accretion may produce about $4\times10^{30}$ J (e.g., [1]), while approximately $2\times10^{30}$ J are required to melt Mars, considering both temperature rise from that of the solar nebula to reach the melting temperature and heat of fusion. Accretion of a Mars-sized planet may occur within $\sim10^5$ years [2]. Such fast accretion would limit the loss of accretional heat and may imply significant heating by short-lived isotopes. Among the short-lived radionuclides, $^{26}$Al contributes the most heat. Between $10^5$ years and the effective death of the nuclide, $^{26}$Al initially in chondritic abundances contributes about $2\times10^{30}$ J of energy to a planet the size and composition of Mars, sufficient in itself to melt the planet entirely. The potential energy release of core formation in Mars is expected to raise the average temperature of the planetary interior by an additional 300°C [3]. The sum of energy from accretion, core formation, and short-lived radionuclides is therefore several times that required to melt the planet (e.g., [1–]
If these processes occur quickly enough to avoid losing heat to space, or in the presence of an insulating atmosphere [8,9], then large-scale melting of the planet is expected.

The time of core formation from measurement of $^{182}$W anomalies provides an additional constraint on the rate of planetary formation. Over the last decade estimates for the time of core formation of the terrestrial planets has shortened from within 10 to 13 My [10–12] to less than 10 My after solar system formation [13,14]. On this timescale heating from $^{26}$Al would be insignificant, but even so other sources of heat could largely melt the planet.

Righter et al. [15] determined that Martian Ni, Co, Mo, W, P and Re abundances estimated from melt inclusions in Martian meteorites are consistent with a magma ocean of only 700 to 800 km depth. Rubie et al. [16] estimated the depth of a terrestrial magma ocean to be 550 km based on Ni partitioning. If similar nickel content and partition coefficients pertain for Mars, then the Martian magma ocean may have been about 1500 km in depth. A differing oxygen fugacity in the Martian mantle, however, may affect the partitioning behavior of siderophile elements. While a fully molten mantle may be possible based on the energetics of planet formation, these geochemical arguments suggest that at least a significant fraction if not all of the Martian mantle was initially molten.

Fractional solidification of a magma ocean creates a mantle stratigraphy that is gravitationally unstable and will overturn through Rayleigh–Taylor instabilities. Fractional solidification and subsequent mantle overturn provides a self-consistent model that explains several aspects of Mars believed to have developed before 4.0 Ga. These aspects include differentiation of mantle source regions into isotopically distinct and ancient mantle reservoirs [17]; development of an early and brief but strong magnetic field; the formation of an early crust to record that field; the preservation of mantle reservoirs through a stably stratified mantle resistant to convective mixing; and heat flow low enough to allow the preservation of lateral variations in crustal thickness over the lifetime of the planet.

In the absence of a conductive lithospheric lid, heat loss due to solid-state thermal convection may have been sufficient to sustain an early core dynamo on Mars. Thus, the existence of an internally generated magnetic field early in Mars history could be taken to indicate the higher rates of heat transfer associated with plate tectonics [18]. However, volcanism represented by the suite of Martian meteorites has sampled isotopically distinct mantle reservoirs that formed early in the evolution of planet. The preservation of such early-formed isotopic reservoirs is unlikely in the presence of convective mixing due to the large-scale mantle circulation associated with plate tectonics.

On the Moon anorthosite is recognized as strong evidence for a magma ocean, but among the collection of SNC meteorites no anorthositic crustal material similar to that present on the Moon has been identified. Fractional solidification of a deep magma ocean in a larger planet like Mars would sequester aluminum at depth in dense majorite and garnet phases, significantly reducing the amount of aluminum available for late plagioclase crystallization (see Fig. 8 in [17]). The purpose of this paper is to evaluate the consequences that magma ocean fractional solidification and mantle overturn may have for the evolution of Mars and perhaps the other terrestrial planets.

2. Magma ocean solidification

The fundamental processes within a crystallizing magma ocean are well summarized in Solomatov [19]. Vigorous convection in the low-viscosity liquid magma ocean is assumed to maintain a homogeneous liquid composition and an adiabatic variation of temperature with depth. Because adiabats are steeper than solidi, a solid region develops at the bottom of the magma ocean and thickens as the ocean cools (Fig. 1). During rapid solidification, cumulates retain their solidus temperatures and compositions.

During solidification, a region of crystals plus liquid, where the temperature lies between the solidus and liquidus, will be present above the fully solid region. The deeper portion of this partly solid zone will form a stress-supporting network at some critical crystal fraction (between 10% and 60% depending on crystal shape [20,21]). The viscosity of this “networked” zone becomes close to that of a solid, deforming by thermally activated creep.

The mineralogy of the resulting solidified mantle (Fig. 1) follows the bulk mantle composition of Bertka and Fei [22] and the phase relations and bulk
solidus and liquidus of Longhi et al. [23] (Fig. 1, left and center). In this example, the magma ocean is assumed to be 2000 km deep above a core with radius 1396 km, corresponding to a pressure of 24 GPa [24]. As the magma ocean solidifies, phases are fractionally crystallized in small increments from the remaining liquid of the magma ocean. Majorite and $\gamma$-olivine crystallize from the bottom of the ocean at 24 GPa to 14 GPa. The magma ocean continues to crystallize as garnet, olivine and pyroxene until the low-pressure end of garnet stability is reached. The final portion of the magma ocean crystallizes as olivine and pyroxene until an Mg# of ~20 is reached in the solids. Low-pressure and evolved phases are disregarded.

The modeled solidus temperatures of the evolving magma ocean liquids are based on results from the MELTS program [25], which predicts a solidus $\sim 150^\circ$C scolder at 95% total crystallization than that of the initial bulk liquid composition. The final cumulate stratigraphy, lying at its solidus, ranges in temperature from 2100 $^\circ$C at the core–mantle boundary to 950 $^\circ$C near the surface.

The distribution of incompatible elements during solidification, in particular heat producing isotopes of U, Th, and K, depends highly on the presence of even small amounts of melt that may be retained within solidifying solids. This will in turn depend on the relative rates of melt migration and compaction of the mostly solid mantle, an aspect of the model under current consideration. These models investigate the effects of 1% to 10% of interstitial liquid retained in magma ocean cumulates. Radiogenic incompatible elements are partitioned with the solid phases using distribution coefficients [26, 27].

Crystallization is expected to occur in cold, mostly liquid downwellings that impinge on the top of previously solidified cumulates. At sufficiently high pressure olivine and pyroxene are more buoyant than coexisting liquids [28]. In our models the depth of neutral buoyancy for olivine and pyroxene occurs at 7.5 GPa; however, garnet is negatively buoyant at all pressures. Dense garnet crystals may then settle while olivine and pyroxene remain suspended in the convecting liquid. Segregation of the
dense garnet depends on the relative rates of settling and reentrainment at the boundary with underlying solids [29,30]. If garnet does settle and pyroxene and olivine remain entrained in the region between 14 and 7.5 GPa, cooling and crystal formation will continue until the reached it critical crystal fraction, after which pore fluid convection may continue until reequilibrated the solid phases. New crystals will settle to the top of the region of critical crystal fraction at 7.5 GPa. A likely outcome of this process is effective batch crystallization of pyroxene and olivine in the pressure interval from the top of the garnet layer to 7.5 GPa.

Whether crystals remain suspended or settle out is fundamental to understanding the chemical evolution of a magma ocean. Mass conservation requires that the average vertical velocity of convective motions across any horizontal plane in the magma ocean vanish, so both upward and downward fluid motions must occur. The settling of dense crystals results in persistent sinking relative to convecting liquid, so crystals must eventually settle out even if the settling velocities are small compared to convective motions [31,32]. The ability of crystals to accumulate at the bottom depends on a competition between settling and entrainment [19,29,32]. Adjacent to a horizontal boundary, vertical velocities are small. However, convective motions can move particles horizontally along the boundary toward regions of convergence beneath buoyant upwellings, where viscous stresses can lift the particles.

A magma ocean cools by losing heat at the planetary surface. Cold downwellings are thus expected to provide a significant fraction of the convective heat flux and, as indicated above, crystallization is expected to occur primarily in downwelling plumes. Experiments indicate that the entrainment rate due to cold downwelling plumes is not sufficient to prevent the settling of particles [29]. The heat of fusion from crystallization in downwelling plumes could be approximated as a heat source at depth generating hot plumes capable of entraining particles. However, if crystallization in downwellings begins at shallow depth, where cold downwellings first reach the liquidus, treating all of the heat of fusion as a heat source at the bottom of the magma ocean would overestimate the heat flux available to be carried by hot plumes.

Solomatov et al. [32] conducted laboratory experiments on a fluid layer heated from below and derived the following equation, in which particles with diameters smaller than \( D \) can be entrained in convective flow.

\[
D \approx \frac{10}{\Delta \rho g} \left( \frac{\eta \alpha g F}{c_p} \right)^{1/2}
\]

Here \( \Delta \rho \) is the density difference between the particles and the liquid, \( g \) is the acceleration of gravity, \( \eta \) is the fluid viscosity, \( \alpha \) is the coefficient of thermal expansion, \( c_p \) is the specific heat, and \( F \) is the heat flux transported by convective motions. Thus, higher viscosity and higher heat flux result in the entrainment of larger particles.

With a heat of fusion of 418.7 kJ/kg and the assumption that two-thirds of the magma ocean crystallizes on a time scale of \( \sim 10^4 \) years [8,30], then heat flux \( F \) could be as large as \( 10^7 \) W/m\(^2\). However, the heat flux in Eq. (1) should only be that carried by hot plumes, perhaps a small fraction of the total convective heat flux. Thus, the size of crystals that can settle will be larger than that estimated if all of \( F \) were carried in hot plumes.

With heat flux \( F \) of \( 10^3 \) W/m\(^2\) Eq. (1) indicates that particles larger than about 40 \( \mu \)m will settle from the convective flow. This estimate assumes a density difference \( \Delta \rho \) of 200 kg/m\(^3\), gravitational acceleration \( g \) of 3.7 m/s\(^2\) (although \( g \) will be lower at smaller radii in the planet), magma viscosity \( \eta \) of 0.1 Pa s, thermal expansivity \( \alpha \) of \( 3 \times 10^{-5} / \)C, and heat capacity \( c_p \) of 1.2 kJ/kg. The value of 0.1 Pa s for a liquid of picritic composition [33] is consistent with values from Solomatov [19], but higher than values from Rubie et al. [16]; lower viscosities would allow smaller mineral grains to settle. If the effect of crystallinity is considered, however, effective viscosity in regions where hot plumes form may be much larger.

In an idealized model, garnet is assumed to settle, forming a \( \sim 150 \)-km-thick layer beginning at 14 GPa (Fig. 1, right side; Fig. 2, left side). Following garnet segregation pyroxene and olivine continue to crystallize, but in the pressure interval between the top of the garnet layer and the end of their positive buoyancy at \( \sim 7.5 \) GPa the phases will not settle to the bottom boundary. Magma ocean cooling...
and crystal formation will continue until the region has reached its critical crystal fraction, after which pore fluid convection may continue to reequilibrate the solid phases. New crystals will settle to the new bottom boundary layer at 7.5 GPa. A likely outcome of this process is effective batch crystallization of pyroxene and olivine in the pressure interval from the top of the garnet layer to 7.5 GPa. Batch crystallization over this pressure interval will change the cumulate density shown in Fig. 2 (left) to an average of the narrow range of values shown, changing the resulting density distribution before overturn only slightly.

3. Magma ocean cumulate overturn

The cumulate stratigraphy from fractional solidification results in the unstable density stratification shown in Fig. 2 (left). A gravitationally unstable stratigraphy will overturn via Rayleigh–Taylor instabilities to a stable profile with intrinsically densest materials at the bottom, on timescales that depend on the rate of thermally activated creep. The instability to overturn is largely independent of the order and proportion of fractionating mantle minerals: mafic mantle minerals all result in progressive iron enrichment in the liquid and the formation of an unstable stratification. The density, temperature, and trace element profiles of the cumulates after overturn are also shown in Fig. 2.

Overturning depends on a competition between the rate of solidification and the rate of overturn of solidified cumulates [17]. The highest rate of overturn occurs near the end of solidification. A simple estimate of the timescale for complete overturn can be obtained from the Rayleigh–Taylor timescale of an unstably stratified viscous layer of thickness $d$, uniform viscosity $\eta$, and constant compositional density gradient $\gamma$ with stress-free top and bottom boundaries [34]

$$t_{RT} \approx \frac{4\pi^2 \eta}{\gamma gd^2}. \tag{2}$$
This simple example shows that the overturn time decreases rapidly as solidification proceeds, because it varies inversely as the square of the solidified layer thickness $d$, and because the density gradient $\gamma$ (see Fig. 2, left) and average gravity $g$ continually increase. The viscosity of near-solidus cumulates is taken to be comparable to values from the Earth’s upper mantle [35]. A nominal value of $10^{19}$ Pa s value is used here; but asthenospheric values on Earth may be as low as $10^{18}$ Pa s [35], and variations of creep rate with water content and grain size widen the possible range of viscosities of a newly solidified magma ocean. Solid viscosity is proportional to the grain size cubed [36]. Because grain size is likely to increase with time after initial crystallization [37] until stresses again increase during convective motions, viscosity may increase with time due to grain growth until convection begins. If the initial magma ocean liquid was hydrous, even the nominally anhydrous minerals in the magma ocean cumulates will contain some water content, further reducing their viscosity. The effects of lower viscosity in a newly crystalline magma ocean are to speed overturn.

At the beginning of solidification, overturn of a layer a few kilometers thick with $\gamma = 7 \times 10^{-5}$ kg/m$^3$/m requires on the order of $10^9$ years. When cumulates are more than 1800 km thick, $\gamma$ may be as high as $3 \times 10^{-4}$, and overturn time is on the order of only $10^5$ years. If it occurs on time scales of $10^5$ years, heat will conduct distances on the order of only a few tens of kilometers during overturn. Cumulates are thus assumed to retain their solidus temperatures during overturn, adjusted for adiabatic rise or fall.

The time required to crystallize a magma ocean depends strongly upon the presence of a stable solid lid or an atmosphere, factors which control the rate of heat loss at the planetary surface [8]. The top layer of the magma ocean, the solid stagnant lid, would act as both an insulator, slowing the crystallization of the magma ocean, and as an inhibitor of complete cumulate overturn, holding the shallowest cumulates in their positions through high viscosity due to temperature. The growth and persistence of such a lid is questionable, however. Abe [8] estimated that a water-rich atmosphere may provide sufficient insulation to maintain the magma ocean in a completely liquid state at its surface. In the absence of such insulation, the shallowest cumulates will quench to solid against the cold of near-space or a thin atmosphere. Once solidified, the material in the lid will be denser than the liquids beneath. If the lid overlies a convecting liquid region, it will be prone to foundering through its own weight, convective stresses on its bottom boundary, and disruption by impacts. If the lid forms over a magma ocean with critical crystal fraction extending to the base of the lid, such a lid may survive magma ocean crystallization and overturn, and perhaps to the present day. This may be the first opportunity for earliest crust-like formation on a planet with a magma ocean.

In the absence of a solid conductive lid, the main insulating effect would be any atmosphere degassed during planetary formation and magma ocean crystallization. Including the effect of an insulating atmosphere, Abe [8] indicates that a 50 GPa terrestrial magma ocean is two-thirds crystallized by volume within 10,000 years (Tonks and Melosh [30] estimate 50% crystallization by 10,000 years), but crystallization then slows and complete crystallization does not occur until 30 million years. Estimates for Martian magma ocean crystallization also range to several tens of millions of years ([8,9,19,30,38], cf. [39]). The time required for crystallization thus overlaps the timescales of overturn (billions of years at the beginning of crystallization to about 1 million years when crystallization is complete). As long as the completely solid portion remains chemically isolated from the still partly liquid regions, however, the solid can be rearranged before solidification is complete and the outcome will be the same as that modeled here (Fig. 2, left side).

The timing of mantle geochemical reservoir formation inferred from SNC meteorites is in agreement with theoretical estimates for magma ocean crystallization. Estimates of the age of Martian core formation (silicate–metal segregation) based on the Hf-W isotope system vary from a maximum of 10 million years to 30 million after planetary accretion [12,13]. Mantle differentiation (melting and fractional solidification to generate isotopically distinct reservoirs) may have postdated or may have been contemporaneous with core formation. The time of mantle differentiation (liquid–solid silicate fractionation) based on most estimates from Sm–Nd and $^{142}$Nd systematics is less than 30 million years after planetary accretion [11,40]. These ages place constraints on the magma ocean.
crystallization time: the magma ocean must have crystallized by about 30 million years at the latest, forming isotopic reservoirs that have remained separate during the subsequent evolution of the planet.

If cumulates crystallized with a density gradient varying monotonically with radius, overturn would proceed in a simple way as inversion of the mantle to a monotonic density gradient decreasing with radius. The model at the top in Fig. 3 shows the final, nearly laterally homogeneous result of overturning a cumulate mantle with a monotonic density gradient. Solidification models indicate that cumulates, however, do not form a monotonic density gradient; the pre-overturn profile in Fig. 2 clearly shows that cumulates in portions of the lower and upper mantle have identical densities with different bulk compositions and correspondingly different mineralogies.

Numerical models indicate that non-monotonic density profiles like these overturn to form mantles that are laterally heterogeneous compositionally (for computational details, see Appendix A). The two bottom models in Fig. 3 show the result of overturning a cumulate mantle with an initially non-monotonic density gradient. As the density structure in these models approaches a final state of a purely radial density stratification, the densest materials from near the surface have sunk to the core–mantle boundary and buoyant mid-mantle material has risen to the

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**Fig. 3.** Numerical models of overturn. Patterns of overturn are shown with a monotonic density gradient (top), the pre-overturn density gradient from Fig. 2 with isoviscous rheology (center) and with temperature- and pressure-dependent rheology (bottom). The numerical model used is a two-dimensional axially symmetric version of CitCom. For each model the evolution of density stratification and the movement from initial depth are plotted. Initial depth is a proxy for cumulate composition; different values of initial depth imply different cumulate compositions. Times are shown as multiples of overturn times (Eq. (2)). Cumulates with non-monotonic density gradients overturn as monotonically stratified cumulates do but produce a compositionally laterally heterogeneous mantle. With pressure-dependent rheology downwelling material increases in viscosity as it falls, requiring a longer time to achieve a final stable state; however, large-scale overturn still occurs on a similar timescale to other models shown here. In terms of initial depth, overturn creates a long-wavelength, long-term heterogeneity that is most enhanced by the presence of temperature- and pressure-dependent viscosity. In this axially symmetric geometry overturn leads to a strong mode-one component to the distribution of temperature and composition in the mantle following overturn.
surface. However, upgoing and downgoing plumes from other initial depths have stalled at neutral buoyancy, creating a radius interval containing large-scale lateral heterogeneity in the initial depth variable. Initial depth of crystallization is a proxy for both degree of differentiation of the magma ocean and for temperature, with shallower initial depth corresponding to more iron- and incompatible element-rich compositions. Thus, the lateral heterogeneity in initial depth in Fig. 3 would correspond to lateral heterogeneity in temperature, mantle composition and concentrations of radiogenic and trace elements.

4. Results of magma ocean cumulate crystallization and overturn

All the models in this suite produced a final cumulate overturn stratigraphy where the coolest, shallowest cumulates sink to the core–mantle boundary. The initial temperature of the core depends upon the partitioning of potential energy of core formation between core and mantle materials. In the following we adopt an initial core temperature of 2100 °C. At this temperature the core is at thermal equilibrium with the earliest cumulates, but above its own liquidus temperature. Mantle near the surface solidifies at about 950 °C, so cumulate overturn produces a temperature contrast across the core–mantle boundary. The final cumulates to form also contain the bulk of the silicate mantle’s radiogenic and trace elements, and they carry these heat-producing elements to the core–mantle boundary when they fall.

Inverting a silicate cumulate stratigraphy does not produce sufficient potential energy release to raise the Martian mantle more than several degrees; this minor temperature increase alone does not result in melting. Moving hot cumulates to lower pressures, however, can allow them to exceed their solidi and produce significant volumes of decompression melt (Fig. 2).

A hemispheric-scale crustal dichotomy is one of the predominant features of Mars that developed early in its evolution. A strong spherical harmonic degree-one component of overturn may provide an endogenic origin for the crustal dichotomy. Models described here do not fully explore this possibility; three-dimensional models are required to avoid the constraints imposed by the axially symmetric geometry. However, all these axisymmetric models (Fig. 3, bottom) exhibit a strong mode-one component of overturn, particularly early in overturn when adiabatic melting would be occurring, and further studies are exploring whether this is sufficient to create the dichotomy through overturn decompression melting.

During overturn, the earliest crust could form from adiabatic melting of hot cumulates that rise above their solidii (Fig. 2). This model predicts two compositions of melt produced during overturn (Fig. 2), differing in volume by a factor of 2, and these melts may have had a degree-one distribution on the surface (Fig. 3). As shown in Fig. 2, the source compositions of these two melts would be significantly different. Both melts are basaltic, but relative to a melt from a source of chondritic bulk composition the deeper melt would have higher alumina and magnesia content, while the shallower, more voluminous melt would be depleted in alumina but rich in iron.

The total melt volume produced during overturn would be sufficient to spread over the surface of Mars to a depth of ~100 km. While larger than geophysical estimates of the current Martian crust of about 50 km [41], this model estimate must be considered an extreme upper bound. Segregation of melt from cumulates is likely to be incomplete, and a cold, solid lid may halt ascent of buoyant materials during overturn and limit their adiabatic melting. The modeled melt volumes are therefore considered roughly consistent with Martian crustal estimates.

Following overturn, the core loses heat that is conducted into the cooler, stably stratified mantle above it. Core heat flux is predicted by solving the transient heat conduction equation in a spherical geometry using the temperature distribution that results from overturn and several different distributions of radioactive heat-producing elements. The stable density profile should prevent initiation of thermal convection driven by heat flux at the core–mantle boundary on these time scales. The different radioactive element distributions reflect uncertainties about the percentage of liquid retained at each stage of crystallization as well as the location of the final, highly fractionated liquids that contain the bulk of heat-producing elements. One possible distribution of heat-producing elements is produced by the fractional crystallization of the chondritic magma ocean with 1% of melt retained in each increment of solid-
ification, resulting in an exponential trace element profile. The final 1% of liquid contains a trace element concentration of about 65 times chondritic. As an alternate simplification, all the radiogenic elements may be concentrated in a final evolved liquid fraction. In these cases all radiogenic elements are uniformly distributed in the final 50 km of the magma ocean, which falls during overturn to form a layer of radiogenic element-enriched material 250 km thick at the core–mantle boundary. Two cases are considered, one with an initially chondritic magma ocean and a second with an initial magma ocean with five times chondritic trace element concentration. These various heating distributions, along with one having no radiogenic heating, are shown as insets in Fig. 4.

Core heat flux is therefore driven by the initial temperature difference between the core (~2100 °C) and the cold fallen cumulates (950 °C), but in time is increasingly limited both by heat loss from the core and by radiogenic heating in the lowermost cumulates. In the first 100 My after overturn, in the absence of thermal convection, the core cools between 1000 and 1500 °C, depending upon radiogenic element content. The initially high heat flux from the core drops off exponentially for all models and falls below 0.02 J/m² s after 40 to ~150 million years. Stevenson [42] estimates that a superadiabatic core cooling rate on the order of 80 K/Gy (~0.022 J/m² s) is required to initiate a Martian core dynamo. Our calculations indicate that the superadiabatic heat flux
required may be as high as 0.10 J/m² s, but is likely between 0.05 and 0.02 J/m² s, using an equation from [42]. The Martian core therefore has sufficient superadiabatic heat flux in all of these models to maintain a core dynamo for at least 15 and perhaps as long as 150 million years after magma ocean overturn.

These results do not depend strongly upon the modeled temperature profiles or heat-producing element distributions in this study: if the heat flux across the core–mantle boundary required to initiate a core dynamo is as low as 0.02 J/m² s, any initial temperature difference across the core–mantle boundary larger than 100 °C would be sufficient to initiate a brief dynamo.

The density profile after overturn is highly stable, as shown in Fig. 2 and in [17]. This stratification will strongly inhibit the onset and mixing depth of thermal convection [43]. In the solid state, a temperature at the bottom of the mantle significantly higher than the melting temperature would be required to overcome compositional stability and cause solid-state thermal convection. This strong compositional stability preserves the isotopically distinct mantle source regions required by the SNC meteorites.

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Appendix A

The dynamics of overturn are explored numerically using a 2D axially symmetric spherical version of the thermochemical convection code CITCOM. Density differences driving convective overturn are a function of both composition and temperature. A particle method is used to advect the non-diffusing composition (modified by S. Zhong to consider an initially continuous composition field). The resulting buoyancy drives convective motions in a very viscous fluid with temperature- and pressure-dependent viscosity. For the linear density models, the viscosity and initial temperature are uniform. For the non-monotonic density profiles, the initial distribution of composition (representing density at 1 atm and at a potential temperature of 1 °C) used is the profile calculated in the fractional crystallization model shown in Fig. 3. The initial temperature is the predicted solidus. For these cases, a thermal Rayleigh number

$$Ra_T = \frac{\rho_0 g \Delta TR^3}{\eta \kappa} = 3 \times 10^9$$

(3)

and compositional Rayleigh number

$$Ra_C = \frac{\Delta \rho g R^3}{\eta \kappa} = 1 \times 10^{10}$$

(4)

are considered with a no-slip condition on the outer boundary, a free-slip on the inner boundary, and no-flux conditions in composition. Temperature has a no-flux condition on the outer boundary and is fixed at a non-dimensional value of 1 on the inner boundary. The models shown in Fig. 3 have 128 × 64 uniform grid and an initial particle distribution of 32 particles per element. Resolution tests indicated that finer grids yielded no major changes in time scale or the state of composition after final overturn.

For cases with the temperature- and pressure-dependent viscosity, an Arrhenius viscosity law with an activation energy of 100 kJ/mol and activation volume of $3.67 \times 10^{-6}$ m³/mol is considered. This activation energy, smaller than that expected for solid-state diffusion creep (~350 kJ/mol), is used to insure convergence of the flow solver. The activation energy (within the range predicted for diffusion creep) was chosen to allow a fairly uniform viscosity for the initial distribution of material along its solidus.

Times are given relative to the initialization of overturn are non-dimensionalized by the Rayleigh Taylor overturn time (2).

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