The Mercurian magma ocean, first crust, and implications for planetary formation mechanisms

by

Stephanie M. Brown

Submitted to the Department of Earth, Atmospheric and Planetary Sciences in Partial Fulfillment of the Requirements for the Degree of Bachelor of Science in Earth, Atmospheric and Planetary Sciences at the Massachusetts Institute of Technology

May 08, 2009

Copyright 2009 Stephanie M. Brown. All rights reserved.

The author hereby grants to M.I.T. permission to reproduce and distribute publicly paper and electronic copies of this thesis and to grant others the right to do so.

Author redacted

Department of Earth, Atmospheric and Planetary Sciences May 17, 2009

Certified by redacted

[Linda T. Elkins-Tanton] Thesis Supervisor

Accepted by redacted

Samuel Bowring Chair, Committee on Undergraduate Program

The author hereby grants to MIT permission to reproduce and to distribute publicly paper and electronic copies of this thesis document in whole or in part in any medium now known or hereafter created.
Abstract

The size of the Mercurian core and the low ferrous iron bearing silicate content of its crust offer constraints on formation models for the planet. Here we consider a bulk composition that allows endogenous formation of the planet’s large core, and by processing the mantle through a magma ocean, would produce a low-iron crust. More Earth-like bulk compositions require silicate removal, perhaps by a giant impact, to create the planet’s large core fraction. The earliest crusts expected in a giant impact scenario are discussed in comparison to the endogenous model. We find that the endogenous model can produce a large core with either a plagioclase flotation crust or a low-iron magmatic crust. For the giant impact model, in the absence of a plagioclase flotation crust, the impact may be constrained to occur within about 300,000 years of the planet’s initial fractionating magma ocean, at which time the giant impact can remove most of the silicate iron oxide budget of the planet before gravitational overturn carries it into the deep planetary interior. Thus a specific bulk composition is required to make Mercury endogenously, but specific timing of events is required to make it exogenously through giant impact. Measurements taken by the MESSENGER mission, when compared to predictions given here, may help resolve Mercury’s formation process.
Acknowledgements

I would like to thank my advisor, Linda Elkins-Tanton, for the invaluable guidance and support with this project and with my academic and professional development.

A suggestion from Paul Hess was the initial inspiration for this research. The National Science Foundation Astronomy program and the MIT Undergraduate Research Opportunities Program funded this work.
Table of Contents

Abstract........................................................................................... 02
Acknowledgements........................................................................ 03
Table of Contents........................................................................ 04

1.0 Introduction................................................................................... 05
  1.1 Surface and interior composition constrains.............................. 05
  1.2 Mercury formation mechanisms.............................................. 06
  1.3 Structure of our models......................................................... 08

2.0 Models............................................................................................ 10
  2.1 Compositions............................................................................. 11
  2.2 Magma ocean solidification processes..................................... 13
  2.3 Modeling Solidification.............................................................. 14

3.0 Results.......................................................................................... 16
  3.1 Flotation crusts........................................................................ 16
  3.2 Volcanic crusts........................................................................ 17

4.0 Discussion........................................................................................ 19
  4.1 Endogenous models.................................................................. 20
    4.1.1 Opaques on the surface of Mercury.................................... 22
    4.2.2 FeO in the mantle and in lavas and the size of the core...... 23
  4.2 Exogenous model..................................................................... 24

5.0 Conclusion...................................................................................... 26

6.0 References...................................................................................... 29

7.0 Figures........................................................................................... 34

8.0 Tables............................................................................................ 41
1.0 Introduction

We model magma ocean solidification of several hypothetical bulk mantle compositions of Mercury and are able to constrain the compositions of their corresponding first crust, which may remain today. It may be possible to compare our predicted crusts with measured compositions from the MESSENGER mission and thus further constrain Mercury formation mechanisms.

1.1 Surface and interior composition constrains

Data collected from Earth-based observation and by the three flybys of MARINER 10 between 1974 and 1975 and compared with the Moon show that Mercury's surface is low in iron bearing silicates, containing between 0 and 3 – 4 wt% FeO in the silicates (Denevi and Robinson, 2008; Solomon, 2003). Mid-infrared spectra have detected the presence of calcic plagioclase feldspar and low-FeO pyroxene (Robinson and Lucey, 1997; Robinson and Taylor, 2001; Solomon, 2003). Recent measurements by the flybys of Mercury MESSENGER are consistent with these ground-based observations and limit total iron content to be less than 6 wt% (Solomon et al., 2008), with iron oxide in silicates at less than 2 – 3 wt% (McClintock et al., 2008; Robinson et al., 2008). Mercury's surface also has an unknown global darkening agent thought to be produced from the presence of opaque minerals or ferrous iron-bearing silicates (Devevi and Robinson, 2008; Robinson et al., 2008, McClintock et al., 2008). Considering the constraints on global iron oxide in the silicates, the global darkening agent is preferred to be an opaque oxide, commonly thought to be ilmenite based on lunar analogues (Devevi and Robinson, 2008; Robinson et al., 2008, McClintock et al., 2008), although it may be a more magnesium rich opaque (Riner et al., 2009).
While Mercury’s surface silicates are iron-poor, the bulk planet contains a high metallic iron fraction. Mercury’s high density (Anderson 1975; Anderson et al., 1987; Newcomb, 1895; Rabe, 1950) indicates a massive core, thought to be ~75% of the planetary radius and 60% of the mass of the planet (Urey, 1952; Anderson and Kovach, 1967; Reynolds and Summers, 1969; Siegfried and Solomon, 1974; Solomon, 2003). Uncertainties in the core radius and mass have been explored in response to the recent finding that the outer core is molten (Margot et al., 2007; Solomon 2007). Light alloying elements, such as sulfur, may explain this molten state and their presence may indicate a larger core than previously thought (Harder and Schubert, 2001; Hauck et al., 2007; Riner et al., 2008). Considering a range of 0 – 10% light alloying element in the core, the density ranges from $5.2 \pm 0.7 \text{ gm/cm}^3$ with plausible values between 5.0 and 5.4 gm/cm$^3$. The resulting core radius ranges from 1675 km – 2075 km with a corresponding 53 – 78% of the mass of the planet (Riner et al., 2008).

### 1.2 Mercury formation mechanisms

During the late 1970’s and the 1980’s, after MARINER 10, several hypotheses for the origin of Mercury were motivated by Mercury’s high core fraction when compared to the other terrestrial planets. These different hypotheses are based upon either planetary accretion from compositions inherently high in metallic iron and low in silicates (endogenous) or silicate removal from an initially larger Mercury (exogenous). We consider three hypotheses as the most plausible and distinct in recent literature (Solomon, 2003) and discuss them here. (See Taylor and Scott, 2003 for details on the hypotheses.)

In 1978, Weidenschilling proposed that aerodynamic drag in the early solar system had a different effect on silicates than on the iron that endogenously formed Mercury.
The hypothetical vulnerability of silicates to aerodynamic gas drag, when compared to iron, produced a planet enriched in iron. Weidenschilling’s aerodynamic sorting hypothesis, based upon a nebular condensation model, culminates with a crust produced by a crystal-liquid fractionation of a silicate magma ocean (Weidenschilling, 1978; Solomon, 2003).

The remaining two hypotheses modify Mercury after planetary accretion by removing silicates from the differentiated, cooled mantle. The first relies on a giant impact to remove the silicates from a Mercury 2.25 times the mass it is today (Benz et al., 1988, 2007). This giant impact hypothesis could transmit enough energy to re-melt the mantle of the planet (Benz et al., 1988). The giant impact is the favored hypothesis among many researchers (Boss, 1998; Taylor and Scott, 2003) for the formation of Mercury in part due to its similarities with the current theory of the formation of the Moon.

For a giant impact to produce Mercury, a variety of specific conditions must be met. The proto-Mercury target must contain 32 wt% metallic iron and be impacted head-on at 20 km/s or obliquely at 35 km/s. The ejecta from this impact must all be sub-centimeter-sized to prevent reaccretion onto Mercury (Benz et al., 1988, 2007). The giant impact hypothesis relies upon accretion across a wide radius of the protoplanetary disk in order to account for the high relative velocity of the impactor. However, the other terrestrial planets all have higher FeO contents in their crusts than does Mercury; if the impactor came from a wide feeding range, crustal composition would be expected to be more similar (Taylor and Scott, 2003).
In the final model discussed here, 70 to 80% of the silicate mantle is vaporized by radiation from the hot young Sun and is then removed by a strong solar wind (Cameron, 1985a; Fegley and Cameron, 1987). Temperatures at the distance of Mercury would have to reach between 2,500K and 3,500K to vaporize the mantle. Cameron proposes a three stage model that forms the cores of the terrestrial planets before the Sun begins to form, placing Mercury’s core between the necessary high temperatures around 2,500K and 3,500K during high mass accretion of the Sun, enough to volatize the mantle (Cameron, 1985a, 1985b, 1988; Cameron et al., 1987; Cameron and Decampli, 1982; Decampli and Cameron, 1979). This Solar System formation model is not consistent with more modern models reviewed and discussed by Boss (1998, 2003) and Chambers (2003, 2004).

Temperature in the Solar System evolves as the system forms from the nebula; it was hottest during the highest mass accretion rates and has been cooling since. Temperature also decreased with radius from the Sun. Planetesimals began aggregation around 2–3 million years after the formation of CAIs (McKeegan and Davis, 2003) and Woolum and Cassen (1999) show that by 1 Ma, the nebula has cooled to between 200K – 800K at 1 AU. Safronov (1972) places disk temperatures between 115K – 186K in the zone of Mercury based on solar radiation calculations. Cameron himself realized the unlikely conditions necessary to form a Mercury depleted in silicates with this method and posits that the giant impact hypothesis as more probable (Cameron, 1988).

1.3 Structure of our models

With a view toward the likely formation processes of the other terrestrial planets, we propose that Mercury’s large core and iron-poor surface may be consistent with planetary
formation from a bulk composition richer in metallic iron than that inferred for the Earth or Mars, rather than by solar ablation (Cameron, 1985a; Fegley and Cameron, 1987) or giant impact (Benz et al., 1988, 2007). The hypothesis presented here is parallel to Weidenschilling’s (1978) sorting theory in that we propose a primary bulk composition of Mercury produced its large core, but rather than sorting alone, we suggest that high temperatures and highly reducing conditions in the early Solar System helped produce a Mercurian bulk composition that would eventually accrete to form a large iron core with low-iron silicate mantle.

At distances close to the Sun, material had the opportunity to become reduced because temperatures were around 2,500K to 3,500K as Cameron (1985a, 1985b) notes. However, instead of assuming that Mercury was present during these temperatures and was ablated due to these high temperatures, we consider the situation in which extremely reducing conditions in the earliest stages of planetary accretion nearest to the Sun may have produced the unusual metallic iron fraction by reducing iron otherwise bound into silicates.

Mercury likely formed from materials condensed at a range of major axes from the Sun, but we suggest that high temperatures around Mercury’s orbit may have reduced their iron, regardless of their origin. Equilibrium condensation (discussed by Latimer (1950), Urey (1952), Lewis (1972)) predicts volatility and FeO content to increase with increasing heliocentric distances. Reducing conditions are expected at high temperatures, when metallic iron and sulfur will condense. Mercury was later accreted, at lower temperatures, from this reduced material. The creation of Mercury’s high metal
fraction by accretion of metal-rich chondrites material has previously been proposed and briefly investigated by Taylor and Scott (2003).

We are effectively reconsidering the equilibrium condensation hypothesis within the framework of radial mixing – which is the major prediction from dynamical models of planetary accretion. We are not, however, presenting any kind of specific new nebular model here. We present the results of models for the formation of Mercury that use known chondritic meteorite compositions to match the large core fraction of the planet and follow self-consistent models of magma ocean solidification to produce predictions for Mercury's crustal composition. Our hypothesis accounts for the large metallic iron fraction of Mercury and predicts a bulk mantle and crust composition low in FeO and volatiles and with refractory elemental abundances similar to chondrites (none enriched). The results of these models may be compared to the existing observations, and await further data from the MESSENGER mission.

2.0 Models

The observations of Mercury's surface composition and its core size offer an initial constraint on early planet- and crust-forming processes. We assume that a magma ocean existed after the accretion of the planet. We model the solidification of the magma ocean and investigate the consequences of differing initial bulk compositions. Significant changes in bulk compositions result in alterations in final mineralogy and structure of the planet as well as initial crustal compositions.
2.1 Compositions

The Mercury formation hypothesis presented here is motivated in part by the discovery of chondrites with an unusually high metallic iron composition: Bencubbinite (CB) chondrites. This rare primitive material contains at least 60 wt% metallic iron (Weisberg et al., 2000). Should Mercury have been built from a CB-like bulk composition, its metallic iron fraction can be entirely explained without further iron reduction. Taylor and Scott (2003) propose that the accretion of metal-rich CB chondritic planetesimals is a plausible alternative to previous hypotheses for the formation of Mercury. Other chondrite classes could not create Mercury even if all their constituent iron was metallic (Figure 1). Because Mercury orbits near the Sun, however, ejection of its bulk planet-building material (possibly the CB material composition) out to Earth’s orbit is far less likely than is our obtaining samples of bulk material consistent with building the Earth and Mars, and we would therefore have fewer examples of its material (Gladman et al., 1996; Love and Keil, 1995; Wetherill, 1984;).

FIGURE 1

CB chondrites are mostly chondritic breccias, commonly described as “shock heated.” The CB meteorite group contains two distinct subgroups, CBa and CBb, and petrologic evidence suggests that the CBb subgroup is more primitive. CBa chondrites may be the products of impacts onto a CBb compositional target (Weisberg and Ebel, 2005; Weisberg et al., 2006).

The origin of these meteorites is controversial. Many have suggested that there is evidence of formation by condensation in the nebula; however, others have used the same evidence to suggest that these meteorites formed from condensation of a metal rich vapor
cloud produced by an impact with a planetesimal (Campbell et al., 2002; Weisberg et al., 2004). Currently, the only suggestion for the mechanism for forming such a metal-rich gas is by a high velocity or a molten planetesimal impact involving a metal rich body, possibly with a CR or a CBb like composition (Burbine et al., 2002; Campbell et al., 2002; Lauretta et al., 2007; Taylor and Scott, 2003; Weisberg et al., 2000, 2004).

Current models of planetary accretion indicate that planets’ feeding zones extend over a range of distances from the Sun (O’Brien et al., 2006; Raymond et al., 2006). Therefore, formation from a single meteorite composition is highly unlikely. The CB chondrites offer an example composition with sufficient iron to build Mercury’s large core and a resulting mantle with interestingly high silica, but we do not argue that any planet is formed from any one primitive composition.

In this study we use three potential mantle compositions for Mercury. The first is the silica-rich composition that results from removing all iron (into a putative core) from the average CB chondrite composition (Lauretta et al., 2007; Weisberg et al., 2001). As the amount of FeO in this composition relies upon the extent of the reducing conditions, we consider a varying amount of FeO in our starting bulk magma ocean compositions, while holding the remaining oxides constant. The second is a modeled average Earth mantle from Hart and Zindler (1986) (Table 1). The Hart and Zindler (1986) composition is derived from measured compositions of Earth-mantle lherzolites (compositional endmembers of undepleted fertile mantle rocks). Morgan and Anders (1980) provides the third composition, produced from an equilibrium condensation model from solar elemental abundances with certain values filled in by using lunar abundances and inferences for Mercury. The resulting mantle mineral assemblages differ significantly
and produce different predictions for the mantle and early crustal compositions of the planet. These initial bulk compositions are given in Table 1.

TABLE 1

2.2 Magma ocean solidification processes

We assume that the energy of accretion and core formation is converted into enough heat to melt the planet, producing a whole-mantle magma ocean with an initial depth of 600 km, from a radius of 1,840 km at the core-mantle boundary to 2,440 km at the surface. Although magma oceans are not a certainty of planetary formation, they are a likelihood, and here we start with the simplest assumption, a whole-mantle magma ocean.

Magma ocean solidification proceeds through two major phases. First, because the slope of the adiabat in the well-mixed magma ocean is steeper than the slope of the solidus of the bulk magma ocean, as cooling proceeds the adiabat first intersects the solidus at depth. Thus the solidification of the magma ocean proceeds from the core-mantle boundary to the surface (Solomatov, 2000; Elkins-Tanton et al., 2003).

The majority of mantle cumulate minerals possess a crystal site that incorporates either magnesium or iron. Because magnesium is incorporated preferentially, evolving magma ocean liquids, and therefore later-solidifying minerals, are enriched in iron. This process can produce solid mantle cumulates with densities that increase with radius through iron enrichment, and this iron-enrichment process is critical for flotation of buoyant minerals in the magma ocean. Density gradients in the solid cumulate can also be produced by mineral phase changes.

In the second stage of magma ocean solidification, the gravitationally unstable solidified silicate mantle cumulates overturn to a stable configuration as discussed in
Elkins-Tanton et al. (2003, 2005a, 2005b) and Elkins-Tanton (2008). During this process, low-density cumulates that originated at depth rise buoyantly and regions will melt adiabatically as they near the surface, potentially producing an earliest igneous crust. Simultaneously, higher iron cumulates sink into the planetary interior. Overturn creates a mantle that is gravitationally stable and therefore resistant to the onset of thermal convection.

Mantle overturn is driven by the density instabilities that are produced during mantle solidification. The density instabilities are commonly produced by the incorporation of magnesium over iron into the structures of the minerals; however, the opaque layer opaques in this case do not contain silica) that forms close to the surface will contain dense titanium-bearing minerals that will also drive overturn, as has been suggested for the Moon (Hess and Parmentier, 1995). In the case of no iron oxide in the original bulk composition, overturn is driven by this titanium layer. However, the viscosity may be sufficiently high to prevent this overturn and a titanium layer may partially remain (Riner et al., 2009) a possible outcome not modeled here, but discussed in section 4.1.1.

FIGURE 2

2.3 Modeling solidification

To model solidification of the low-iron oxide CB chondrite silicate composition, we used existing magma ocean evolution code from Elkins-Tanton et al. (2003, 2005a, 2005b) and Elkins-Tanton (2008) and modified it for Mercury. This code requires an a priori set of assumptions about the mineral assemblages being solidified at every step, and these assemblages are specific for each of our bulk compositions. For Mercury, we used John Longhi’s magma ocean code FXMO, which is a Gibbs free energy
minimization program for magma ocean solidification. The mineral assemblages predicted from FXMO along with existing experimental data are then set as initial conditions in our magma ocean evolution code, which calculates the equilibrium compositions and densities of each fractionating increment of mantle cumulates, and tracks the composition of the evolving residual magma ocean liquids.

In the CB bulk composition, high silica produces mantle cumulate phases including quartz, as shown in Figure 2. The model for solidification of this CB bulk composition proceeds from the bottom up with the following phase assemblages: olivine + orthopyroxene, olivine + clinopyroxene, quartz + pyroxene + garnet, followed by the same with plagioclase replacing garnet, and finally, pyroxene + plagioclase + quartz + opaques (ilmenite and chromite). The Earth mantle composition contains far less silica and thus quartz is not stable in its magma ocean solidification evolution. The Morgan and Anders model also does not have enough silica to crystallize quartz and would have phase assemblage similar to Earth but with higher orthopyroxene fractions (Figure 2).

The solidus that we assume for all models is similar to that of a terrestrial peridotite over the pressure range appropriate for Mercury’s magma ocean. For this relation and other details of model calculations, see Elkins-Tanton (2008). In these models the solidus determines the temperatures of the newly solidified mantle cumulates. Their temperatures in turn contribute to calculations of cumulate density through the thermal expansivities of the minerals. Solidi for a range of compositions do not vary enough to significantly change the predictions of these models.
3.0 Results

3.1 Flotation crusts

There are two major opportunities for the formation of an early crust through magma ocean processes. In the first, as hypothesized for the Moon, plagioclase or other buoyant minerals float in a denser coexisting magma ocean liquid and form a conductive lid. Iron enrichment creating progressively denser evolving magma ocean liquids is the major contributor to plagioclase flotation (Smith et al., 1970; Warren, 1985; Wood et al., 1970). The second process that may produce an initial crust is cumulate overturn in the solid state following magma ocean solidification, in which buoyant cumulates rise from depth and melt adiabatically, producing secondary magmas that erupt onto the planet's surface.

The planetary body most associated with a magma ocean is the Moon, thought to have been produced from a body the size of Mars impacting the Earth (Hartmann and Davis, 1975; Cameron and Ward, 1976). During solidification of the lunar magma ocean, plagioclase became buoyant and floated to the surface to form the initial crust. This plagioclase flotation has been suggested as a definitive marker of a magma ocean, but flotation will only occur if the plagioclase is less dense than the magma and if plagioclase stability is reached before a significant crystal fraction forms (the network of crystals would then prevent flotation). The pressure range of Mercury is sufficiently larger than that of the Moon that it suppresses the crystallization of plagioclase until nearer the planetary surface, at \( \sim 1.5 \text{ GPa or 88 km depth} \).

For the Earth-like (Hart and Zindler, 1986) and the Mercurian Morgan and Anders mantle compositions, plagioclase is stable and sufficiently buoyant to float in the magma
ocean liquids toward the end of solidification (Figure 3c, d), assuming there is no crystal network to impede flotation.

In contrast, the iron oxide-poor magma ocean liquids in the CB chondrite composition models do not reach densities high enough to float plagioclase (Figure 3a). To test the sensitivity of plagioclase flotation on Mercury to iron oxide content in the initial magma ocean bulk composition, we incrementally add iron oxide into the CB Mercury bulk composition (Figures 3, 4). In a CB-chondrite-type composition buoyant minerals will not float in the magma ocean unless its initial FeO content is at least 2.5 wt% (Figure 3b, Figure 4). With these higher initial FeO contents, both plagioclase and quartz float in the Mercurian magma ocean, creating a novel iron-poor, silica-rich earliest crust.

FIGURE 3

FIGURE 4

3.2 Volcanic Crusts

In a second mechanism for creating an early crust, adiabatic melting in upwelling solids during compositionally-driven overturn produces a melt that erupts onto the planetary surface. Even in an initial magma ocean with insufficient iron oxide to float plagioclase, sufficient cumulate density gradients are produced to cause gravitationally driven solid-state overturn following solidification. The solidification model produces predictions for the original depth and compositions of cumulates that, during gravitationally driven overturn, rise sufficiently to melt (Figure 5). Their exact melt composition cannot be calculated precisely as specific experimental studies would be
required, but approximations can be made based on the source bulk composition using the free energy minimization program pMELTS (Ghiorso et al., 2002).

FIGURE 5

When the bulk magma ocean FeO content is below the critical 2.5 to 4 wt% value, the earliest planetary crust is likely formed by erupting lava that originates from melt during overturn. Two regions that form at depth during solidification are sufficiently buoyant and hot that they may melt adiabatically during overturn and contribute to this posited earliest igneous crust. The first region originates from 600 km to 330 km depth and temperatures of 1420°C to 1510°C. The second region originates from ~120 km at 1120°C to 60 km depth at 1,130°C (Figure 6).

FIGURE 6

These source regions would produce lavas with compositional ranges shown in Table 2. The deepest melting region produces lavas that are basaltic in composition while the shallowest melting region produces lavas that are rhyolitic in composition.

TABLE 2

The CB-like bulk composition with < 2.5 wt% FeO is the only composition that does not produce a flotation crust. In this case, the mineralogy of the surface can be predicted to some degree by calculating the mineral modes of the melt, solidified and equilibrated at the surface (Table 2). The deep melting region produces an albite – olivine – clinopyroxene rich crustal rock, perhaps a troctolite. The shallow melting region produces a lava composed mostly of quartz and some smaller amounts of clinopyroxene, anorthite, and spinel. Each region contributes to about half of the melt volume, producing a crust with thickness on the order of ~10 km.
In summary, by modeling solidification of a Mercury magma ocean, it is possible to predict the composition of the planet’s initial crust, which may remain today. Using a CB chondrite composition successfully reproduces Mercury’s core fraction and creates a mantle very low in iron oxide. Unless FeO exceeds 2.5 to 4 wt% in the bulk mantle, plagioclase will not float and form an earliest crust. The initial crust on such a planet will consist of high-silica and high-magnesia lava erupted from melting produced during solid-state mantle overturn. With higher initial iron oxide fractions, the CB chondrite composition will create an initial flotation crust of quartz and plagioclase, but dominated by quartz, in contrast to the Moon. Mantle compositions more similar to the Earth will likely produce a plagioclase flotation crust. These three scenarios create distinct and measurable predictions.

4.0 Discussion

The strongest existing constraints on Mercurian formation models are its very low iron bearing crustal silicates (Denevi and Robinson, 2008; Robinson and Lucey, 1997; Robinson and Taylor, 2001; Solomon, 2003) and the large metallic iron core (Urey, 1952; Anderson and Kovach, 1967; Reynolds and Summers, 1969; Siegfried and Solomon, 1974; Solomon, 2003). In our magma ocean models, there are two opportunities to produce an earliest crust, which may remain on the planet today and can be compared to observed constraints. The first is through flotation of buoyant minerals during magma ocean solidification, and the second is through eruptions of magma produced by adiabatic melting during solid-state cumulate mantle overturn after magma ocean solidification.
4.1 Endogenous models

On a planet the size of Mercury any of the bulk mantle compositions considered create a plagioclase or plagioclase + quartz flotation crust, provided the bulk composition has at least 2.5 wt% FeO. Thus a CB chondritic Mercury with >2.5 to 4 wt% FeO in its mantle, or a Mercury with a Hart and Zindler (1986) Earth-like mantle, or a Morgan and Anders (1980) model mantle might all present a low-iron ancient crust through flotation of buoyant phases (Table 2).

The lavas produced by adiabatic melting during solid-state overturn would not likely erupt through a plagioclase flotation crust in the absence of crustal thinning, for example, from giant impacts. Should a flotation crust not form, as in the case of a low-iron silicate mantle, then the lavas may erupt to form the earliest crust. A low-iron CB chondritic Mercurian mantle would likely produce a low-iron earliest igneous crust and would be entirely consistent with the existing data on Mercury’s surface.

Taylor and Scott (2003) produce crustal composition predictions from a series of possible Mercury bulk compositions, including those of Morgan and Anders (1980), the vaporization model by Fegley and Cameron (1987), Goettel (1988), and a CB-chondrite like composition. They calculate a 10% partial melt at 1 GPa of the bulk compositions using John Longhi’s phase-equilibria program MAGPOX.

The igneous crust produced in our models from the CB chondritic composition contains a much higher silica content than the Taylor and Scott (2003) lava predictions, and lower Al₂O₃ and CaO. These differences arise from the material that melts; our source melts are mostly olivine and pyroxene from the deepest layers that crystallize first from a magma ocean with only a small fraction of the later-solidifying aluminum- and
calcium-rich plagioclase and clinopyroxene. These fundamental differences reflect that our models incorporate a differentiated planet, while the Taylor and Scott (2003) model assumes a homogeneous bulk mantle composition.

Therefore, regardless of the amount of iron oxide in the mantle, fractionation processes and overturn can cause the crustal silicates to be lower in iron oxide, potentially consistent with observations. Thus formation of Mercury from a CB chondritic-like bulk composition would create a planet with a sufficiently large metallic iron core and low iron content crust to match observations of Mercury. Formation from other compositions can also make a low-iron crust, but would not create the core fraction observed.

The process of fractional solidification of the magma ocean creates progressively more iron-rich cumulates toward the planetary surface, which during overturn sink into the planetary interior and are replaced by iron-poor materials both near and at the planetary surface. Compositional stratification resists thermal convection as large temperature deviations are required to overcome compositional density with thermal expansivity. Thermal convection begins after a period of quiescence, and such a planet may possess a more heterogeneous mantle today than one that began without compositional stratification. This is a possible explanation for the heterogeneity of Mars' mantle (Elkins-Tanton et al., 2005a, 2005b). Zaranek and Parmentier (2004) have calculated the time to onset of convection after stable stratification is created. Beyond that, vigor of convection will determine the eventual homogeneity of the mantle. Convective mixing on Earth over the age of the solar system is insufficient to fully homogenize the mantle (Becker et al., 1999).
There has been discussion on the presence of a light and possibly volatile alloying element, such as sulfur, to preserve a molten core (Harder and Schubert, 2001; Hauck et al., 2007; Riner et al., 2008). Under highly reducing conditions iron is likely to form bonds with alternative anions, such as sulfur. Such conditions might encourage the formation of a planetary core with an unusual fraction of light alloys.

4.1.1 Opaques on the surface of Mercury

Our models indicate that an oldest volcanic crust on Mercury may form with few or no opaque minerals such as ilmenite and chromite. Instead, these opaque phases solidify in a layer toward the end of the magma ocean solidification, analogously to models of the lunar magma ocean, and reside in the uppermost mantle at the end of solidification.

While we assume in our models that the late state cumulates overturn as described in section 2.2, these cumulates will not crystallize until temperatures have fallen to about 1,150°C (Hess and Parmentier, 1995; Van Orman and Grove, 2000). By the time the dense material has finished solidifying and any gravitational instabilities begun to form, the temperature would have fallen farther. These low temperatures near the Mercurian surface produce high viscosity in the Fe-Ti oxide layer. The viscosity may be sufficient, in fact, to prevent their overturn (Elkins-Tanton et al., 2002) and leave them near the planetary surface. Riner et al. (2009) discuss the possibility that the spectra of the low reflectance material seen on Mercury’s surface reflect this un-overturned oxide layer.

Titanium and iron could be delivered to the surface by later magmatic activity, either from standard planetary convection or from convection induced by impacts. Evidence for volcanism on Mercury is supported by data obtained by the first MESSENGER flyby (Head et al., 2008). Impact events may also excavate compositions
bearing iron oxide, and so the presence of oxides on the surface of Mercury (Robinson and Lucey, 1997; Robinson and Taylor, 2001) may be either from excavation or from melting and rising material from depth. In these magma ocean cumulate models, the source area for ilmenite and chromite lies between 60 km and 120 km depth (Figure 5). Models considered for the Moon estimate impact-triggered melts in craters to originate from 7 km to 45 km deep if the craters range from 50 km to 300 km in radius as depth goes as \(0.15 \times \text{radius}\) (Housen et al., 1983; O'Keefe and Ahrens, 1999). To deposit oxides on the surface of Mercury, impacts would have to melt material from the second layer of the planet (Figure 6).

4.1.2 FeO in the mantle and in lavas and the size of the core

Harder and Schubert (2001) have shown that the modeled size of Mercury’s core depends directly upon assumptions of its composition. If the entire planet consisted only of FeS the density of the planet would deviate by only 0.5% from the measured mean density. Models of the internal structure of Mercury are constrained only by the radius and mean density, but improvements on characterization of the moment of inertia and the gravitational field of the planet will help determine the size of the core if we assume densities for the crust, mantle, and core (Margot et al., 2007; Solomon, 2007).

Previous researchers have estimated that the amount of FeO in the lavas on the surface of Mercury directly reflects the amount of FeO within the mantle of the planet, because FeO has a partial melt partition coefficient ~1 (Robinson and Taylor, 2001). However, if the mantle is heterogeneous, as our models indicate, surface lavas may not demonstrate the entire composition space of Mercury’s mantle and as such, the unsampled regions in the mantle may contain more iron and be more dense than thought.
Specifically, Mercury may carry a relatively high FeO fraction at depth following magma ocean overturn, while crustal lavas contain little iron, reflecting their low-iron source regions. This may be evident on the Moon, as various bulk compositions (Dreibus and Wanke, 1985; Longhi, 2003) range between 7.5% - 14% FeO, while measured lava basalts contain 15.5% - 22% FeO (Papike et al., 1998).

In Mercury interior models such a mantle containing higher iron fraction at depth might predict a smaller core radius. If MESSENGER data indicates a smaller core, this may support the magma ocean model for Mercury’s mantle. It may be difficult to distinguish the effect of the mantle density as the uncertainties in the core composition may dominate the size of the core (Hauck et al., 2007; Riner et al., 2008).

4.2 Exogenous model

To make Mercury with the observed core and crust characteristics from a more Earthlike bulk composition, one of the exogenous models must be considered. Here we examine the Benz et al. (1988, 2007) giant impact model, which we consider the most likely of the exogenous models. If the proto-Mercury had a core fraction similar to the Earth’s, then the proto-Mercury was a planet about the size of Mars, with a radius of ~3,370 km.

It may be that accretionary impacts at Mercury’s radius are so energetic that the silicate mantle of the impactor is almost entirely blown away, along with some of the protoplanet. In this case, shallow cumulates are lost preferentially from the mantle, leaving deeper cumulates (Kokubo, personal communication). Following solid-state overturn of the initial magma ocean the deep planetary mantle will contain the majority of the initial iron budget. If the giant impact occurs at this point, the deep mantle may be
the part of the planet retained for the new, less massive mantle, and it will be rich in iron (results from models of Mars presented in Elkins-Tanton et al., 2005a). The energy of impact will likely create a second magma ocean, and flotation of buoyant low-iron phases such as plagioclase may be possible. An igneous crust created in this secondary magma ocean, however, will be too iron-rich to match observations of Mercury.

Alternatively, the giant impact may occur at or toward the end of magma ocean solidification. In this second scenario iron has been enriched in evolving magma ocean liquids and resides largely near the surface of the planet, as seen in Figure 7, and overturn has not begun to carry the iron to depth. If the impact occurs at that moment a majority of the iron may be lost from the planet and the remaining mantle will consist of ~50% gamma olivine and ~50% majorite (Elkins-Tanton et al., 2005a), and any secondary igneous crust will therefore be low in iron. This may describe how a Mercury formed by giant impact produced a crust low in iron-bearing silicates, previously unexplained (Taylor and Scott, 2003).

FIGURE 7

The timeframe required for this scenario is relatively well defined; impact must occur after a majority of the magma ocean is solidified, perhaps within 50,000 years of the last giant accretionary impact, and before overturn begins. Onset of overturn of an unstable stratified layer of viscous fluid of thickness $d$ and viscosity $\eta$ with stress-free top and bottom boundaries is:

$$t_{overturn} = \frac{4\pi^2 \eta}{\gamma \rho d^2}$$
where $\gamma$ is the compositional density gradient and $g$ is gravity (Hess and Parmentier, 1995). On Mercury, where $g$ is about 3.7 m/s$^2$, the overturn time of a 600 km thick layer with $\gamma = 3 \times 10^{-4}$ kg/m$^3$/m and $\eta = 10^{20}$ Pa-s is about 300,000 years. This is a generous interval during which proto-Mercury may have experienced another giant impact following an initial magma ocean in this accretion scenario.

We therefore suggest this is a consistent and reasonable process by which to make a Mercury with a large core and a low-iron mantle and low-iron earliest crust if impacts at that radius preferentially remove shallow cumulates: The planet is built from an Earthlike bulk composition and experienced a fractionating magma ocean during accretion. A giant impact occurs before the gravitationally unstable cumulate mantle overturns, and the young Mercury loses most of its silicate iron oxide budget, leaving a large core and an iron-poor mantle and crust.

5.0 Conclusions

Here we have considered a range of initial Mercurian mantle compositions, from an Earth-like composition to a CB chondritic, high silica composition. Models of conditions in the early solar system may help judge the likelihood of these compositions. During planetary formation the Sun's emissions in the visible were far weaker than they are today; however, high-energy emissions were much stronger (Ribas et al., 2005). This strong solar wind and X-ray flux may have produced a more metallic iron-rich composition nearer the Sun. Conversely, dynamical models of planetary formation indicate that radial mixing of planetary embryos during oligarchic accretion is ubiquitous, though less extreme nearest the Sun (O'Brien et al., 2006). Taken together, a more
reduced, CB-chondritic-like composition at Mercury's radius is a reasonable hypothesis, though more Earth-like bulk compositions are also reasonable possible given the likelihood of radial mixing.

The results presented here show that a CB-chondritic type bulk composition is consistent with creating Mercury endogenously, with a metallic iron core matching observations. A low-iron ancient crust can be made by plagioclase flotation, and in the case of a CB-chondrite type composition, a plagioclase plus quartz flotation crust. These crusts are only possible, however, if the bulk mantle composition has more than 2.5 to 4 wt% FeO to produce a dense magma ocean liquid. Lower iron compositions may produce a low-iron igneous crust through adiabatic melting during solid-state cumulate overturn after magma ocean solidification.

Considering a giant model that preferentially removes shallow cumulates, it may be possible to explain the low iron-bearing crustal silicates found on Mercury. In this case, the exogenous giant impact model for mantle removal from a young Mercury remains a viable hypothesis for creating a final Mercury with a thin mantle and a large metallic iron core. In this scenario the proto-Mercury is assumed to have a more Mars-like or Earth-like mantle composition, in keeping with its initial core fraction. The giant impact may occur before solid-state magma ocean overturn in the proto-Mercury, when iron has been progressively enriched nearer the surface of the planet and not yet sunk into the planetary interior. In this case the giant impact might remove the majority of the iron content of the mantle and might well produce a low-iron ancient crust, consistent with observations.

If the giant impact occurred after cumulate overturn on proto-Mercury, the remaining mantle materials would be enriched with the iron from iron-rich cumulates sunk into the
planetary interior. In this case a low-iron crust can only be made by flotation of buoyant materials, requiring an initial composition low in iron.

Evidence for a plagioclase flotation crust on Mercury found by MESSENGER would be consistent both with an endogenous model with $> 2.5$ to $4$ wt% FeO in the bulk mantle, or with an exogenous giant impact model. If, conversely, the low-iron crust on Mercury appears to date to $> 4.5$ Ga and resulted from an erupted magma, then the endogenous formation, CB-chondrite type bulk composition is most consistent.
6.0 References


7.0 Figures

Figure 1. Natural bulk metallic iron content for selected meteorite classes, and their metallic iron content if all iron oxide was also reduced to metallic iron. The metallic iron is presented in combination with NiS as we assume in a simplistic way that the bulk of the nickel and sulfur enter the core. In these compositions, NiS ranges from 0 – 2.28 mass percent of the planet. Only the CB chondrites have sufficient metallic iron to create Mercury’s core through endogenous processes. Data: Lodders and Fegley (1998): CI, CO, CK, CR, EH, and EL; Jarosewich (1990): CM, CV, L, LL, and eucrites; Kenkyujo et al. (1995): H; CB compositions averaged from Lauretta et al. (2007) and Weisberg et al. (2001).
Figure 2. Cross sections of the model Mercuries, with Earth as comparison. The magma ocean solidifies from the core to the surface with the mineral assemblages shown.
Figure 3. Density of minerals crystallizing from the magma ocean as a function of depth. The dotted line shows the density of the evolving coexisting magma ocean liquid. In some cases plagioclase or plagioclase and quartz become less dense than the liquid and can float. Flotation is only suppressed in compositions with less than ~3 wt% iron.
Figure 4. Density of evolving magma ocean liquids as a function of pressure and iron content. Initial iron content is marked on each curve; liquid density changes primarily because iron is enriched in liquids as low-pressure minerals are fractionated from it (see Figure 3). The plagioclase and the quartz lines are the densities of these minerals as a function of pressure in the planet. At weight percents higher than 2.5 wt% FeO there is enough iron to float small amount of plagioclase and quartz. Very near the planetary surface buoyant minerals may be prevented from floating by high crystal fractions. Initial iron contents of ~4 wt% or higher may be required to allow buoyant minerals to float.
Figure 5. Mercurian cumulate mantle density profile of <2.5 wt% FeO CB-like bulk composition immediately following magma ocean solidification, and following gravitational overturn to stability. The mineral labels correspond to the dashed, pre-overturn density profile. The clinopyroxene – garnet – coesite layer and the surface layer are more dense than the material they overlie, and are thus gravitationally unstable. In the solid state, the dense material will sink and the lighter material will rise to stability, forming a mantel density profile shown with the bold line.
Figure 6. Bulk mantle composition for CB chondrite starting composition after overturn to gravitational stability; the shaded regions may experience adiabatic decompression melting as they rise during overturn. Mineral assemblages are also shown.
Figure 7. Proto-Mercury mantle, approximately Mars-sized, prior to overturn. The shaded region indicates the region removed during a giant impact. This region contains most of the iron oxide; after impact, a mantle low in iron oxide would remain.
## 8.0 Tables

**Table 1.** Bulk Mercurian mantle compositions used in models

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>47.1</td>
<td>46.3</td>
<td>67.9</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>6.4</td>
<td>3.96</td>
<td>3.1</td>
</tr>
<tr>
<td>MgO</td>
<td>33.7</td>
<td>38.4</td>
<td>25.0</td>
</tr>
<tr>
<td>CaO</td>
<td>5.2</td>
<td>3.2</td>
<td>2.6</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.3</td>
<td>0.2</td>
<td>0.2</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>3.3</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>FeO</td>
<td>3.7</td>
<td>7.7</td>
<td>0 - 3</td>
</tr>
<tr>
<td>Mg Number</td>
<td>94.2</td>
<td>89.9</td>
<td>100 - 93.7</td>
</tr>
</tbody>
</table>

*aCB chondritic average composition averaged from Lauretta et al. (2007) and Weisberg et al. (2001).*
Table 2. Earliest crustal compositions resulting from magma ocean models.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>CB chondritic avg. &lt;2.5wt% FeO</td>
<td>Magma from deep melting region</td>
<td>This study</td>
<td>Flotation crust</td>
<td>Average magma from cumulate overturn</td>
<td>Average magma from cumulate overturn</td>
<td>10% melt of bulk composition</td>
<td>10% melt of bulk composition</td>
</tr>
<tr>
<td>CB chondritic avg. &gt;2.5wt% FeO</td>
<td>Magma from shallow melting region</td>
<td>This study</td>
<td>Flotation crust</td>
<td>Average magma from cumulate overturn</td>
<td>Average magma from cumulate overturn</td>
<td>This study</td>
<td>Taylor and Scott (2003)</td>
</tr>
<tr>
<td>CB chondritic avg. &gt;2.5wt% FeO</td>
<td>Average magma from cumulate overturn</td>
<td>This study</td>
<td>Flotation crust</td>
<td>Average magma from cumulate overturn</td>
<td>Average magma from cumulate overturn</td>
<td>This study</td>
<td>Taylor and Scott (2003)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Flotation crust mineralogy</th>
<th>Igneous Crust:</th>
<th>SiO₂</th>
<th>MgO</th>
<th>Al₂O₃</th>
<th>CaO</th>
<th>TiO₂</th>
<th>Cr₂O₃</th>
<th>Na₂O</th>
<th>FeO</th>
<th>Resulting crustal thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>NONE</td>
<td>51.7</td>
<td>28.06</td>
<td>2.8</td>
<td>2.86</td>
<td>5.7</td>
<td>5.4</td>
<td>8.35</td>
<td>0.54</td>
<td>3.7</td>
<td>&lt; 10 km - 40 km - 60 km - 30 km</td>
</tr>
<tr>
<td>NONE</td>
<td>82.7</td>
<td>3.9</td>
<td>5.9</td>
<td>5.8</td>
<td>0.7</td>
<td>0.8</td>
<td>0.8</td>
<td>0.25</td>
<td>2.5</td>
<td>0.395</td>
</tr>
<tr>
<td>NONE</td>
<td>67.2</td>
<td>15.98</td>
<td>4.35</td>
<td>4.33</td>
<td>3.2</td>
<td>4.56</td>
<td>4.56</td>
<td>0.395</td>
<td>0.395</td>
<td>3.2</td>
</tr>
</tbody>
</table>

Note: Flotation crust quartz mineralogy is NONE.

a No flotation was considered in these models; see text for discussion.
Table 3. Mineralogy and thickness of crust of model CB < 2.5 wt% FeO

<table>
<thead>
<tr>
<th>Melt from shallow source</th>
<th>Melt %</th>
<th>Qtz</th>
<th>Spinel</th>
<th>Albite</th>
<th>Anorthite</th>
<th>Fayalite</th>
<th>Forsterite</th>
<th>Clinopyroxene</th>
<th>Ilmenite-Hematite-Chromite</th>
<th>Crustal Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>50%</td>
<td>60.26</td>
<td>11.47</td>
<td>5.78</td>
<td>10.07</td>
<td>0.06</td>
<td>1.78</td>
<td>9.83</td>
<td>0.74</td>
<td></td>
</tr>
<tr>
<td>Melt from deep source</td>
<td>50%</td>
<td>8.68</td>
<td>46.13</td>
<td>0.22</td>
<td>20.07</td>
<td>20.32</td>
<td>4.58</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Average magma</td>
<td>30.13</td>
<td>10.08</td>
<td>25.96</td>
<td>5.04</td>
<td>0.14</td>
<td>10.93</td>
<td>15.07</td>
<td>2.66</td>
<td>~10km</td>
<td></td>
</tr>
</tbody>
</table>