**PREDICTING MERCURY'S ANCIENT CRUSTAL COMPOSITION.**
S. Brown and L. T. Elkins-Tanton, MIT, Dept. of Earth, Atmospheric, and Planetary Sciences, Cambridge MA, 02139, ltelkins@mit.edu.

**Introduction:** Mercury's unusually large iron core comprises up ~75% of the radius and 60% of the mass of the planet [1]. This high fraction of iron in the planet when compared to the Earth has driven several hypotheses for its origin.

Weidenschilling (1978) proposed that differences in the response of iron and silicate particles to aerodynamic drag during accretion resulted in a planet poorer in silicates than are the other terrestrial planets. Two other hypotheses propose that Mercury was modified after planetary accretion by removing silicates from the differentiated, cooled mantle. One hypothesis proposes that the hot young Sun vaporizes 70-80% of the silicate mantle, and then removes them with a strong solar wind [3,4]. The other hypothesis relies upon a giant impact to remove the silicates from a Mercury 2.25 times the mass it is today [5].

We propose that Mercury's large core and iron-poor surface are consistent with planetary formation from a bulk composition richer in metallic iron than that inferred for the Earth or Mars (see also [6,7]). The hypothesis is parallel to the sorting hypothesis: we propose that a primary bulk composition of Mercury produced its large core without later silicate removal. Rather than sorting alone, we suggest that high temperatures and highly reducing conditions nearest to the early Sun helped produce this unusually high metallic iron fraction.

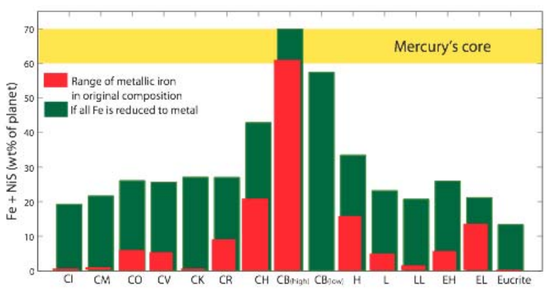


**Figure 1**: Metallic iron fractions measured (red) and maximized through reduction of iron oxide to iron metal (green) for chondrite classes; only the CB has sufficient iron to create Mercury's core without mixing with additional material.

The formation hypothesis presented here is motivated in part by the discovery of chondrites with an unusually high metallic iron composition, the Bencubbinites (CB chondrites). These chondrites contain at least 60 wt% metallic iron [5]. Should Mercury have been built from a CB bulk composition, its metallic iron fraction can be entirely explained without further reduction. It has been proposed that the accretion of metal-rich CB chondritic planetesimals is a plausible alternative to previous hypotheses for Mercury's formation [6,7].

We present the results of self-consistent models of magma ocean solidification to produce three distinct and measurable predictions for Mercury's ancient crustal composition. The results of these models may be compared to the existing observations and await further data from the MESSENGER mission.

Currently there are limited constraints on the surface composition of Mercury. Spectra at visible and near-infrared wavelengths constrain FeO to be less than 3–4 wt%. Mid-infrared spectra indicate the presence of calcic plagioclase feldspar with sodium and low-FeO pyroxene and also indicate a spatially heterogeneous surface [1].

**Model:** We assume that the energy accumulated during the accretionary process is converted into enough heat to melt the planet, producing a magma ocean on Mercury with initial depth 600 km (from r=1,840 km to 2,440 km).

Initial Compositions Assumed in Calculations for Mercury

| | | CB Average (wt%) | H&Z mantle (wt %) |
|---|---|---|---|
| Silicate Mantle | $Na_2O$ | 0.1 | 0.3 |
| | MgO | 25.0 | 38.4 |
| | $Al2O_3$ | 3.1 | 3.9 |
| | $Si0_2$ | 67.9 | 46.3 |
| | $K_2O$ | 0.01 | 0.01 |
| | CaO | 2.6 | 3.2 |
| | $TiO_2$ | 0.2 | 0.2 |
| | $Cr_2O_3$ | 1.1 | 0.5 |
| | MnO | 0.1 | 0.1 |
| | FeO | 0 to 5 | 7.7 |
| Core | Fe | 91.3 | |
| | NiS | 8.7 | |

**Table 1**: Bulk silicate mantle compositions: An average CB chondrite [5,9] with all or most iron reduced and removed to the core, and a hypothetical Earth mantle composition [10].

Magma ocean solidification proceeds through two major phases. First, the solidification of the magma ocean follows the crystallization of silicate minerals from the core-mantle boundary to the surface based on models presented in Elkins-Tanton *et al.* [11-13]. This step produces cumulates with density that increases with radius through iron enrichment, and are therefore gravitationally unstable to overturn, as discussed in Elkins-Tanton *et al.* [11-13]. Second, the gravitationally-unstable solidified silicate mantle cumulates overturn to a stable configuration. The overturn process creates a mantle that is gravitationally stable and therefore resistant to the onset of thermal convection. Hot cumulates that formed deep in the magma ocean rise to shallower depths during overturn and may melt adiabatically, producing the earliest basaltic crust.

Here we compare two potential mantle compositions for Mercury (Table 1). The first is the silica-rich composition that results from removing all or most iron from the average CB chondrite composition [5,9]. The second is a modeled average Earth mantle [10].

To model solidification of the low-iron CB chondrite composition, we use predicted mineral assemblages from the magma ocean code FXMO [14]. Solidification 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+ilmenite and chromite. The Earth mantle composition contains far less silica and thus quartz is not stable in its magma ocean solidification.

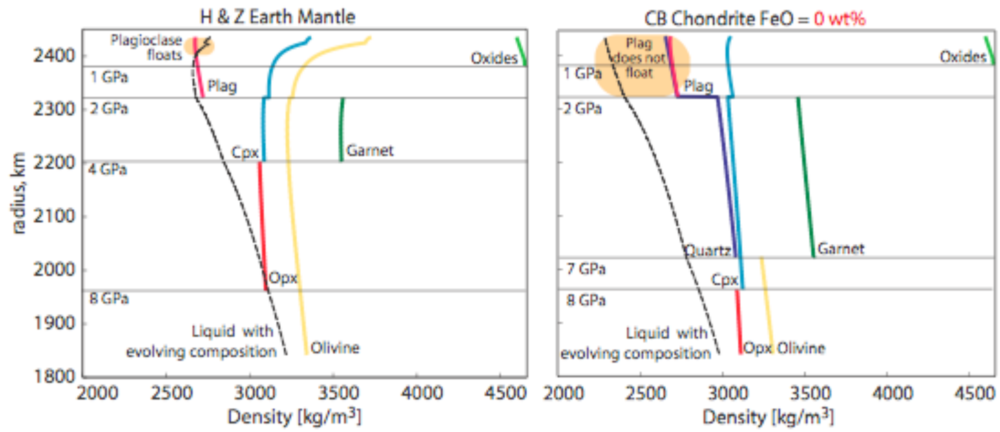


**Figure 2**: Phase densities during solidification of two model compositions for a Mercury magma ocean. On the left, progressive iron enrichment allows plagioclase to float, while on the right, no flotation occurs.

**Results:** There are two major opportunities for the formation of an early crust through magma ocean processes. In the first, as hypothesized for the Moon [15-17], plagioclase floats in a denser coexisting magma ocean liquid and forms a conductive lid. While plagioclase flotation has been suggested as a definitive marker of a magma ocean, flotation will only occur if the plagioclase is less dense than the magma and if plagioclase stability is reached before a significant crystal network forms in the remaining magma ocean liquids. Iron enrichment creating progressively denser evolving magma ocean liquids is the major contributor to plagioclase flotation.

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.

For the Earth-like mantle composition, plagioclase is stable and sufficiently buoyant to float in the magma ocean liquids toward the end of solidification (Fig. 2, left). In contrast, the iron-poor magma ocean liquids in the CB-chondrite models do not reach densities high enough to float plagioclase (Fig. 2, right). Melting during overturn in this case will form the earliest crust. Addition of ~3 wt% FeO to the CB bulk mantle composition creates late magma ocean liquids sufficiently dense to float both plagioclase and quartz.

**Summary and Conclusions:** 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. Unless iron exceeds ~3 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. With higher initial iron 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 or Mars will produce a plagioclase flotation crust. These three scenarios are distinct and create measurable predictions (Table 2).

| | CB 0% FeO | CB >3% FeO | Earthlike |
|---|---|---|---|
| Flotation crust | NONE | quartz>plag | plag |
| Basaltic crust: | 54-75% | NONE | NONE |
| $SiO_2$ | | | |
| MgO | 15-54% | | |
| $Al_2O_3$ | 1-7% | | |
| CaO | 1-7% | | |
| $TiO_2$ | minimal | | |
| $Cr_2O_3$ | minimal | | |
| Crustal thickness | | ~40-60 km | ~30 km |

**Table 2**: Predicted earliest crusts.

**References:** [1] Solomon (2003) *EPSL* 216:441-55. [2] Weidenschilling (1978) *Icarus* 35:99–111. [3] Cameron (1985) *Icarus* 64:285-94. [4] Fegley and Cameron (1987) *EPSL* 82:207-22. [5] Benz *et al* (1988) *Icarus* 74:516-28 [6] Weisberg *et al* (2001) *MAPS* 36:401-18. [7] Taylor and Scott (2001) Workshop on Mercury: Space Environment, Surface, and Interior. [8] Scott and Keil (1999) 34:A104. [9] Lauretta *et al* (2007) LPSC. [10] Hart and Zindler (1986) *Chem. Geo.* 57:247-67. [11] Elkins-Tanton *et al* (2003) *MAPS 3*8:1753-71. [12] Elkins-Tanton *et al (*2005) *EPSL* 236:1-2. [13] Elkins-Tanton *et al* (2005) *JGR* 110(E9):12. [14] Longhi (2007) personal communcation. [15] Wood *et al* (1970) Proc. Apollo 11 LPSC, 965-988. [16] Smith *et al.* (1970) Proc. Apollo 11 LPSC, 897-925. [17] Warren (1985) *Ann. Rev. Earth Planet. Sci. 13*, 201-240.