EARLIEST PLANETARY CRUSTS: CONSTRAINTS ON THE FORMATION OF MERCURY AND IMPLICATIONS FOR BODIES OF DIFFERENT SIZES. S. Brown and L. T. Elkins-Tanton, MIT, Dept. of Earth, Atmospheric, and Planetary Sciences, Cambridge MA, 02139, brownsm@mit.edu, ltelkins@mit.edu.

Introduction: Three mechanisms may form the earliest crusts of terrestrial planets: production of a volcanic crust through mantle melting; flotation of buoyant minerals in a magma ocean; and retention of undifferentiated material. The mechanism that operates depends largely upon the size of the body in question.

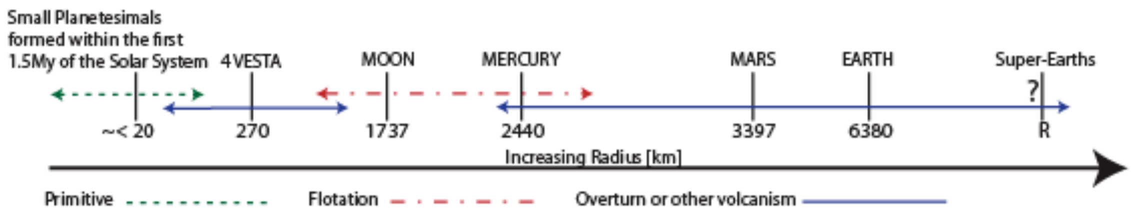

Figure 1: Origin of Earliest Crust by Radius of Terrestrial Body

A volcanic crust may be produced as a result of magma ocean processes. Solidification of a magma ocean naturally produces a silicate mantle with denser layers near the surface, therefore unstable to gravitational overturn. The mantle is predicted to rapidly flow in the solid state to gravitational stability. Hot buoyant silicate mantle rising from depth will melt through depressurization and erupt lavas onto the planetary surface, as long as there is no flotation crust [1, 2]. These lavas solidify to form the planet's earliest basaltic crust. The overturn model produces predictions for the compositions of cumulates that rise sufficiently to melt. 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.

Flotation is only possible on bodies with gravity significant enough to drive crystal separation in magma, but not so large that the low-pressure stability of buoyant minerals is confined to only the final percentages of mantle solidification.

Very small bodies and planetesimals present early in the Solar System will melt from the inside out and the crust would be the undifferentiated bulk composition. Internal dynamics may cause volcanism and so lavas may partially cover the primitive crust.

Mercury's large core and high density obscure the size and nature of the original accreting planet – did the planet form with enough metallic iron to create its large core (endogenous) or was silicate mantle later removed in a giant impact (exogenous)? Was the crust produced from overturn or from flotation of buoyant minerals? Its size is sufficiently small that flotation is possible, but large enough that magma ocean cumulate overturn would occur and would create an earliest volcanic crust in the absence of flotation. Here we consider both models and examine their predictions for the planet's crustal composition, and we compare them to measurements of Mercury's crust that show it has low iron oxide content [3, 4]

Endogenous formation: We consider three bulk mantle compositions for our models: a high metallic iron fraction (Bencubbinite chondrite or CB), a Mercury model composition [5], and an Earth-like composition [6].

All three models may produce a low-iron ancient crust through flotation of buoyant phases, with the constraint that the CB-like composition has to contain > 2.5 to 4 wt% FeO for magma ocean liquids to be sufficiently dense to allow flotation (Table 1). If there is less than about 4 wt% of FeO in the bulk magma ocean, flotation will not occur, and lavas produced by adiabatic melting during solid-state overturn are able erupt to form the earliest crust (Table 2). All these models produce a very low-iron crust and are therefore consistent with existing measurements.

Harder and Schubert [7] have shown that the modeled size of Mercury's core depends directly upon assumptions of its composition. 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 [4]. However, if the mantle is heterogeneous, as our models indicate, surface lavas may not be indicative of the compositional range of Mercury's mantle and as such, 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. In Mercury interior models such a mantle containing higher iron fraction at depth would predict a smaller core radius. Indeed, if *MESSENGER* data indicates a smaller core, this may

Table 1. Earliest crustal compositions resulting from magma ocean models.

Bulk Mercurian mantle composition	CB chondritic avg. <2.5wt% FeO	CB chondritic avg. >2.5wt% FeO	Hart and Zindler (1986) Earth	Morgan & Anders (1980)	Morgan & Anders (1980)	CB chondritic average
Resulting crustal composition	Magma from cumulate overturn	Flotation crust	Magma from cumulate overturn	Magma from cumulate overturn	10% melt of bulk composition at 1 GPa	10% melt of bulk composition at 1 GPa
Reference	This study	This study	This study	This study	[8] a	[8] a
Flotation crust mineralogy	NONE	quartz > plagioclase	100%	100%	NONE	NONE
Igneous Crust:						
SiO_2	67.2				49.2	44.7
MgO	15.98				10.9	19
Al_2O_3	4.35				14.4	13.1
CaO	4.33				18.5	18.8
TiO_2	3.2				1.7	0.9
Cr_2O_3					1.6	0.3
Na_2O	4.56				0.3	
FeO	0.395				3.7	3.2
Thickness	< 10 km	~40–60 km	~30 km	~ 30 km		

[a] No flotation was considered in these models. The differences reflect that our models incorporate a differentiated planet, while the Taylor and Scott [8] model assumes a homogeneous bulk mantle composition.

Table 2. Mineralogy and thickness of crust of model CB < 2.5 wt% FeO

Quartz	Spinel	Albite	Anorthite	Fayalite	Forsterite	Clinopyroxene	Ilminite-Hemitite-Chromite	Crustal Thickness
30.13	*10.08*	*25.96*	*5.04*	*0.14*	*10.93*	*15.07*	*2.66*	*~10km*

support the magma ocean model for Mercury's mantle.

Exogenous formation: To make Mercury with the observed core and crust characteristics from a more Earthlike bulk composition, one of the exogenous models must be considered. If Mercury undergoes a magma ocean stage the timing of a possible giant impact is constrained, because iron is enriched near the planetary surface during magma ocean solidification, and then transported to depth by magma ocean cumulate overturn. For the resulting planetary crust after giant impact to be low in iron oxide the impact must occur such that it removes the majority of the planet's mantle iron. For the giant impact model 9], in the absence of a plagioclase flotation crust, the impact is 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 before gravitational overturn carries it into the deep planetary interior. Thus the planetary crust produced by mantle melting will be low in iron.

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.

Conclusions: Endogenous formation of Mercury with a material high in metallic iron is consistent with formation of the observed low-iron crust. For a giant impact to have created Mercury's large core fraction, its timing must occur very shortly after formation, so that iron is preferentially removed from the planet's mantle before overturn carries it to depth. Measurements by *MESSENGER* may be able to differentiate between a plagioclase (and quartz?) flotation crust and a low-iron volcanic crust, and therefore further constrain formation models.

Further, the initial size of a body determines the process or processes that can create its earliest crust. Only a narrow radius range of body can form flotation crusts (Fig. 1) or retain primitive crusts; most form volcanic crusts. The small gravity fields of small bodies do not provide a sufficient driving force, and on larger bodies, no buoyant minerals crystallize until the mantle is nearly solid.

References: [1] Elkins-Tanton et al. (2003) *MAPS, 38*, 1753-1771. [2] Elkins-Tanton et al. (2005) *JGR, 110*. [3] Robinson and Lucey (1997) *Science, 297,* 197-200. [4] Robinson and Taylor (2001) *Meteoritics & Planet. Sci., 36,* 841-847. [5] Morgan and Anders (1980) *Nat. Academy of Sci., Proceedings*, *77*, 6973-6977. [6] Hart and Zindler (1986) *Chem. Geo., 57*, 247-267. [7] Harder and Schubert (2001) *Icarus, 151*, 118-122. [8] Taylor and Scott (2003) *Treat. Geochem. 447-485.* [9] Benz et al. (1988) *Icarus, 74*, 516-528.