HEAT PIPE PLANETS. William B. Moore[1,2], Justin I. Simon[3], A. Alexander G. Webb[4], [1]Department of Atmospheric and Planetary Sciences, Hampton University, Hampton, Virginia 23668, USA (bill.moore@nianet.org). [2]National Institute of Aerospace, Hampton, Virginia 23666, USA. [3]Center for Isotope Cosmochemistry and Geochronology, ARES, NASA-JSC, Houston, TX 77058, USA (Justin.I.Simon@NASA.gov). [4]Department of Geology and Geophysics, Louisiana State University, Baton Rouge, Louisiana 70803, USA (aagwebb@gmail.com).

Introduction: When volcanism dominates heat transport, a terrestrial body enters a heat-pipe mode, in which hot magma moves through the lithosphere in narrow channels. Even at high heat flow, a heat-pipe planet develops a thick, cold, downwards-advecting lithosphere dominated by (ultra-)mafic flows and contractional deformation at the surface. Heat-pipes are an important feature of terrestrial planets at high heat flow, as illustrated by Io. Evidence for their operation early in Earth's history [1] suggests that all terrestrial bodies should experience an episode of heat-pipe cooling early in their histories.

Conceptual Model: A planet that is cooling via heat-pipes experiences persistent global volcanism that constantly resurfaces the planet. Older layers are progressively buried and advected downwards to form a thick, cold, single-plate lithosphere. Because each layer is deposited at the surface and then pushed down into smaller and smaller spherical areas, lithospheric contraction is a persistent process. At the base of the lithosphere, material is reabsorbed into the mantle or remelted, feeding continuing volcanism. Heat-pipe operation leads to: 1) Cold, thick and strong lithospheres, 2) Dominance of compressive stresses, 3) Continuous replacement of lithospheric material, 4) High melt-fraction (mafic to ultra-mafic) eruptions, and 5) A rapid transition to conductive lid or plate tectonic behavior (Fig. 1). Solar System-wide preservation of ancient, large-scale topography and density variations, predominance of mafic volcanic material, absence of significant extensional strain, and broad regions of uniform surface ages indicating a rapid decline in resurfacing are common features of the terrestrial planets that support periods of heat-pipe cooling.

Mercury: Crustal shortening recorded in the lobate scarps corresponds to a radius change of only a few kilometers [2,3], which imposes strong constraints on the cooling of the mantle and the freezing of the relatively large metallic core. Extensive resurfacing until 4.1 to 4.0 Ga [4] by volcanic eruptions emplaced smooth plains [5] with very little activity since. The eruption of volcanic flows covering broad areas without the development of large volcanic structures suggests that the lavas were low-viscosity, consistent with their mafic to ultra-mafic composition [6].

These observations are readily explained as the result of heat-pipes operating in the first half-billion years of Mercury's evolution. The ongoing resurfacing of heat-pipe volcanism continually replaces the lithosphere, erasing all memory of the earlier shape or size of the planet and thus limiting accumulation of surface strain over this period. The present lithosphere would therefore only accumulate strain due to much slower cooling since heat-pipe volcanism ceased and weak rigid-lid convection or conduction took over.

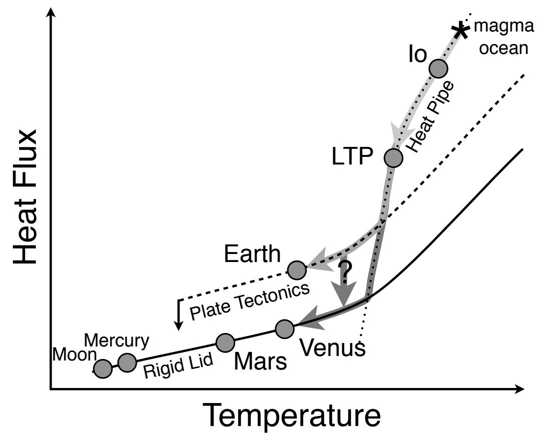

Figure 1. Illustration of terrestrial planet heat flow vs. internal temperature. The sense of evolution as heat sources and internal heat content decline is shown by arrows. The known terrestrial bodies are labeled, LTP stands for Large Terrestrial Planet (i.e. super-Earth), and the initial magma ocean stage is indicated. Heat loss is high and thermal evolution is rapid in the upper right of the diagram, where the magma ocean gives way to heat pipes, and heat flow decreases as heat pipes transition to either plate tectonics or rigid lid convection.

Venus: A thick lithosphere characterized by vertical tectonics [7,8] and a predominance of volcanic resurfacing [9] are the hallmarks of heat-pipe operation. Analysis of structures in Ovda regio [10] indicate that the origin of crustal plateaus are also consistent with vertical collapse of regions of thickened crust [11] likely combined with decay of thermal support [12], indicating shifting locations of the heat-pipes.

Mars: Endogenic models for the formation of the crustal dichotomy between the elevated southern hemisphere and the depressed northern hemisphere invoke either horizontal (plate-tectonic) or vertical (instability or plume) motions. Horizontal models [13] lack evidence of the necessary plate boundary structures that should coincide with the dichotomy boundary [14], while the internal instability models (e.g., [15]) are too slow to form the dichotomy given the ancient age of the northern basement. It has been suggested that the primary crustal formation process may have initiated a hemispheric-scale instability in a dense cumulate layer [16]. This model lacks a means to preserve a sharp contrast in crustal thickness because the instability would become weaker and relax over geologic time. Heat-pipe operation is capable of producing a thick and strong lithosphere very early in Mars' history, preserving the crustal dichotomy and its sharp boundary.

Another fundamental, outstanding observation is the extreme range of trace element abundances and isotopic compositions among the main group of martian meteorites. Tight age constraints come from the short-lived ^{146}Sm-^{142}Nd decay system [$t_{1/2}$ ~68 Ma] that suggests global differentiation of Mars a few 10's to a 100 Ma after Solar System formation (e.g., [17,18]). Martian meteorites show large variation in neodymium isotopes. These extreme variations are ~4 times greater than those of comparable mantle sources on Earth and the Moon and exist because the separation of mantle into distinct reservoirs on Mars was extreme and happened >4 Ga ago. One predication of the heat-pipe mode hypothesis would be early and significant extraction of incompatible material, perhaps to form a primordial crust [19].

The Moon: Unlike the other terrestrial planets, the Moon's surface is dominated by plagioclase-rich anorthositic rocks, with the basaltic maria covering about 1/6th of the surface, mostly on the near side. The prevalence of anorthosite in the lunar highlands is generally attributed to the flotation of less dense plagioclase in the late stages of the solidification of the lunar magma ocean [20,21]. It is unclear, however, that these models are capable of producing the extremely high plagioclase contents (near 100%) observed in both lunar samples and remote sensing data, since plagioclase formation does not begin until over 80% of the initial melt has solidified, and a stable solid matrix of denser minerals is capable of forming at about 50% melt. Formation simply as a floating cumulate is made even more problematic by the near uniformity of the alkali composition of the plagioclase, even as the mafic phases record significant variations in Mg content.

These problems can be resolved if the initial plagioclase-rich flotation crust is further refined through an episode of heat-pipe volcanism. Mafic magmas from the deep interior pond at the base of this crust and produce melts that are higher in plagioclase and incorporate a range of mafic components. Continued operation of this mechanism can result in the nearly pure plagioclase melts that are observed, while the residual mafic minerals of the primordial crust make up the Mg-suite rocks. Requirement for remelting earlier forming fractions of the lid (i.e., anorthosite crust) appears consistent with recent evidence for younger formation of the ferroan-anorthosite suite [22].

The Moon stands out as having a shape that is dramatically out of hydrostatic equilibrium. Its shape is not a fossil of a synchronous rotator at any distance from Earth, but instead must record some other orbital and/or rotational state. The only means by which this record can be preserved over geologic time is substantial lithospheric strength, but all present explanations for the observed shape rely on processes that occur very early in the Moon's evolution when it is much hotter. What is required is a way to rapidly produce a strong lithosphere even when the Moon is young and hot, which is precisely the expected behavior of a body experiencing heat-pipe cooling. The lithosphere is created rapidly and continuously, causing the shape to be recorded around the time the heat-pipe mechanism shuts off leaving behind a strong, distorted lithosphere.

Summary: The geological and geochemical evidence from the terrestrial planets in our Solar System is consistent with heat-pipe operation providing the main source of crustal formation and endogenic resurfacing. Since the equilibrium heat flux of a planet scales as mass/area (for radiogenic heating), terrestrial planets more massive than the Earth should experience longer heat-pipe episodes prior to the initiation of plate tectonics. Due to compressibility of terrestrial materials, a planet twice as massive as Earth should take more than twice as long to cool because the area increases less rapidly than (mass)$^{2/3}$. For the massive "super-Earths" up to five Earth masses, the lifetime of the heat-pipe phase may exceed the lifetime of typical parent stars and thus any subsequent plate-tectonic phase may never be observed.

References: [1] Moore, W. and A. Webb (2013) *Nature* 501:501–505, 2013. [2] Watters, T. et al. (2009) *EPSL* 285, 283–296. [3] Achille, G. et al. (2012) *Icarus* 221, 456–460. [4] Marchi, S. et al. (2013) *Nature* 499, 59–61. [5] Head, J. et al. (2011) *Science* 333, 1853–1856. [6] Charlier, B. et al. (2013) *EPSL* 363, 50-60. [7] Kucinskas, A. and D. Turcotte, (1994) *Icarus* 112, 104-116. [8] Moore, W. and G. Schubert (1997) *Icarus* 128, 415-428. [9] Herrick, R. and M. Rumpf (2011) *JGR* 116, E02004. [10] Romeo, I. and R. Capote (2011) *P&SS* 59, 1428-1445. [11] Nunes, D. et al. (2004) *JGR* 109, E01006. [12] Nunes, D. and R. Phillips, (2007) *JGR* 112, E10002. [13] Sleep, N. (1994) *JGR* 99, 5639-5648. [14] McGill, G. (2000) *JGR* 105, 6945-6960. [15] Roberts, J. and S. Zhong (2006) *JGR* 111, E6. [16] Elkins-Tanton, L. T. et al (2005) *JGR* 110, E12. [17] Debaille, V. et al. (2007) *Nature* 450, 525-528. [18] Caro, G. et al. (2008) Nature 452, 336-339. [19] Humayun, M., et al. (2013) *Nature* 503, 513-516. [20] Warren, P.H., (1985) *Ann.Rev. EPS* 13, 201-240. [21] Elkins-Tanton, L.T., et al. (2011) *EPSL*, 304, 326-336. [22] Borg, L.E., et al. (2011) *Nature* 477, 70-72.