Petrogenesis of Mare Basalts, Mg-Rich Suites and SNC Parent Magmas

Final Performance Report

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Petrogenesis of Mare Picrite Glasses

The successful models for the internal evolution of the Moon must consider the volume, distribution, timing, composition and, ultimately, the petrogenesis of mare basaltic volcanism. Indeed, given the paucity of geophysical data, the internal state of the Moon in the past can be gleaned only be unraveling the petrogenesis of the various igneous products on the Moon and, particularly, the mare basalts.

The picritic glasses are among the most primitive lunar basaltic liquids, and thus most useful in constraining the depth and composition of their source region [Delano, 1980] despite having undergone a certain degree of shallow level olivine crystallization.The bulk of the lunar volcanic glass suite can be modeled as the partial melting products of an olivine + orthopyroxene source region deep within the lunar mantle. TiO_2 contents vary from 0.2 wt % -17.0wt% [Shearer and Papike, 1993]. Values that extreme would seem to require a Ti-bearing phase such as ilmenite in the source of the high-Ti (but not in the VLT source) because a source region of primitive LMO olivine and orthopyroxene, even when melted in small degrees cannot account for the observed range of TiO_2 compositions. The picritic glasses are undersaturated with respect to ilmenite at all pressures investigated therefore ilmenite must have been consumed during melting, leaving an ilmenite free residue and an undersaturated melt [Delano, 1980, Longhi, 1992, Elkins et al, 2000 among others]. Multi-saturation pressures for the glasses potentially represent the last depths at which the liquids equilibrated with a harzburgite residue before ascending to the surface. These occur at great depths within the lunar mantle. Because the liquids have suffered some amount of crystal fractionation, this is at best a minimum depth. If the melts are mixtures, then it is only an average depth of melting. Multisaturation, nevertheless, is still a strong constraint on source mineralogy, revealing that the generation of the lunar basalts was dominated by melting of olivine and orthopyroxene.

We performed a series of piston cylinder experiments in which ilmenite was dissolved into a high Ti synthetic red glass to the point at which the glass reached ilmenite saturation (Beck et al., 2003). The experimental charges were constructed as dissolution couples: a layer of glass was placed on top of synthetic ilmenite powder in a graphite capsule. The charges were annealed at 1100°C and then brought to final run temperatures, held for 0.5 – 8 hours and quenched. The experiments were performed over a pressure range of 1 – 2.5 GPa, and temperatures varied from 1215°C –1475°C. Run products included glass, ilmenite, Cr-ulvospinel and olivine and were analyzed with a Cameca microprobe.

Ilmenite solubility in basalt (represented by the TiO_2 content of the saturated glass) is a function of temperature and pressure. Figure 1 shows the TiO_2 content in the liquid (and ilmenite solubility) increasing linearly with temperature and decreasing with increasing pressure. Temperature dependence is strong, for example, at 1.0 GPa the TiO_2 content increases from 14.1 wt% at 1250°C to 26.47 wt% at 1350°C, • 0.12 wt%/oC. This agrees with previously reported values from. The pressure effect on solubility is ª -6.7 wt%/GPa. Figure 2 shows Ilmenite saturation as a function of both temperature and pressure. Dashed lines are contours representing constant TiO_2 in a liquid in equilibrium with ilmenite.
The dT/dP slopes of the saturation surfaces are ª 67°C/GPa. If the high Ti lunar picritic glasses began as partial melts in equilibrium with olivine + orthopyroxene + ilmenite (± clinopyroxene ± garnet) then this saturation data provides insight into the composition,

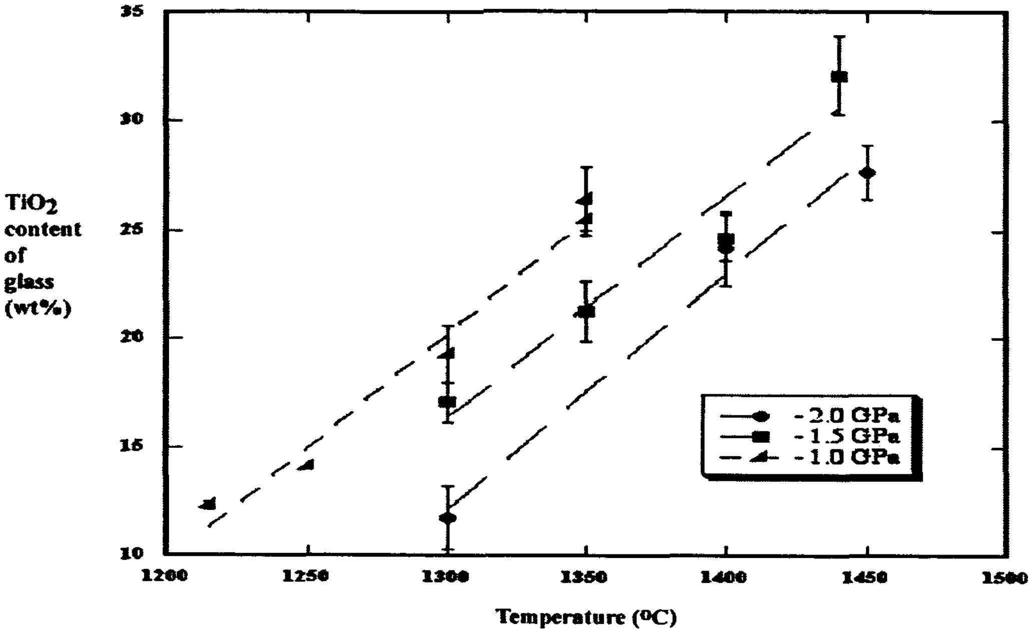

Figure 1. The TiO_2 content of ilmenite saturated glasses as a function of temperature. Dashed lines are linear regressions: blue is 1.0 GPa data, black is 1.5 GPa an red is 2.0 GPa. Each data point is an average value over an experimental chare. Error bars represent one standard deviation from the average.

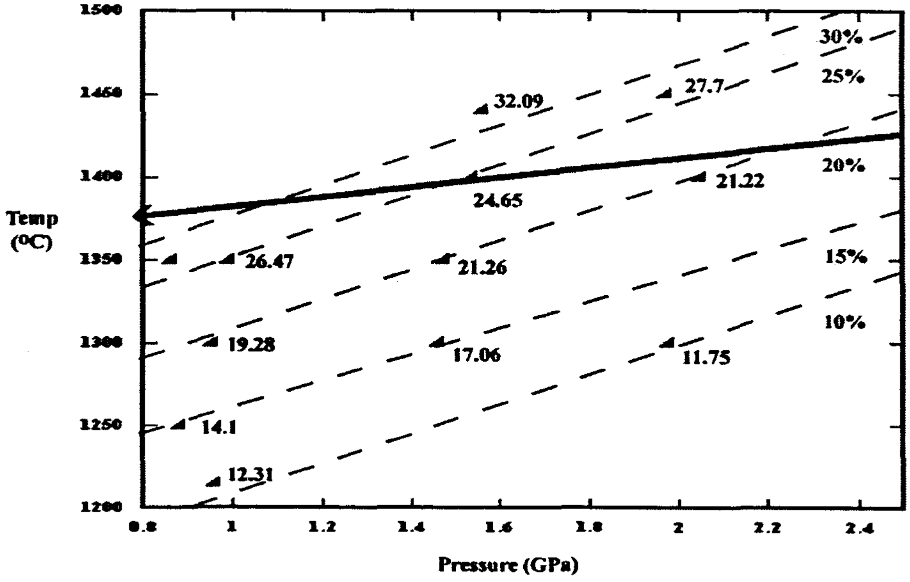

Figure 2. Ilmenite solubility in a high Ti read glass as a function of temperature and pressure. Dashed lines are contours of constant TiO_2 in saturated glass in 5% increments (from 10% - 30% TiO_2). Red triangles are experimental glass data (also shown on figure 1) labeled with their TiO_2 content. The solid arrow is an estimated lunar mantle adiabat. The adiabat dT/dP slope is significantly shallower than the ilmenite 'saturation surfaces' illustrating that a partial melt in equilibrium with ilmenite at depth will have a very high TiO_2 content but become undersaturated as it rises to the surface.

density and viscosity of the melts during the initial stages of melting, prior to consumption of ilmenite. We can also constrain the composition of the melts at multisaturation pressures and temperatures. and also the degree to which to the high Ti picritic glasses have been diluted with ilmenite-free liquids. For example, the Apollo 15 red glass multisaturation point is 1450°C and 2.45 GPa; at these conditions a liquid in equilibrium with ilmenite has more than 20% TiO_2. Figure 2 includes a lunar mantle adiabat estimated from using a potential temperature of 1350°C. If this dense liquid was able to rise adiabatically (perhaps in a diapir) the amount of TiO_2 in the ilmenite saturated liquid becomes progressively higher, leading to further ilmenite dissolution. When ilmenite is exhausted the liquid becomes undersaturated and diluted by further melting of olivine and orthopyroxene.

Petrogenesis of Mg-rich suite Magmas

The origin of the parent magmas for the Mg-rich suite continues to present a challenge- particularly it is difficult to model the comparatively primitive major element composition coupled with the extremely evolved, KREEP-like incompatible element content. An additional constraint that has not attracted sufficient attention is that the trace element enrichment, particularly in the light REE, is not reflected in the Nd isotope content. Many of these light REE enriched rocks are characterized by positive ε_{Nd} (Nyquist and Shih, 1992, Hess, 2000a)! It follows that the light REE enrichment of the Mg-rich parent magmas cannot be attributed to assimilation of KREEP-rich products of the magma ocean since the latter would be characterized by either negative or zero ε_{Nd} unless the ε_{Nd} of the Moon itself is positive (Hess and Parmentier, 1999). Simple assimilation of KREEP-rich liquids with the magnesian parent magmas cannot be a correct theory of petrogenesis. But some connection to KREEP and its apparent asymmetric distribution on the lunar crust should be considered as part of the story.

Gamma-ray data from the Apollo and Lunar Prospector missions suggest that Procellarum-Imbrium (PKT) terrane is the site of enrichments in Th and presumably other highly incompatible elements associated with the KREEP component of the magma ocean (Lawrence et al., 1998; Elphic et al., 2000; Warren and Rasmussen, 1987). Rather than being evenly distributed over the crust, the KREEP component is concentrated in the Procellarum-Imbrium region and the material seems to be radially distributed around this center to points as far as the South Pole Aitkin basin on the back side of the Moon (Jolliff et al., 2000; Haskin, 1998). If the KREEP component is the last dregs of the magma ocean and if the KREEP material underplates the lunar crust uniformly, then it is puzzling that the KREEP appears to be excavated only at the PKT site. This is particularly worrisome because the South Pole Aitkin basin is not only bigger but also deeper and possibly older than the basins in the Procellarum region. Why then wasn't the KREEP material excavated also in the SPA basin?

Haskin and colleagues (Haskin, 1998; Wieczorek and Phillips, 2000; Korotev, 2000) suggest that the high-Th geochemical province is the product of the final stages of crystallization of the magma ocean. Korotev (2000) argued that the crystallization of the last dregs of the magma ocean was not only prolonged but also experienced a renewed thermal activity, which resulted in the remelting of the magnesian cumulates below the crust. The latter appear in that stratigraphic position because of the overturn of the cumulate pile of the magma ocean (Ryder, 1991; Hess and Parmentier, 1995). Wieczorek and Phillips (2000) predict that the residual KREEP layer remains partially molten for a few billion years and that this layer provides sufficient thermal energy to partially melt the deep interior and produce mare basalts. The magnesian-suite troctolites and norites which are characterized by very refractory major element compositions but highly incompatible element contents (Papike et al., 1996) are believed to be products of this KREEP liquid which has dissolved substantial quantities of magnesian cumulates (Korotev, 2000; but see Hess, 1999; 2000a).

We have modelled the thermal evolution of thickened KREEP liquid layer that extends over a significant fraction of the lunar nearside with a prescribed thickness, heat production and beneath a insulating anorthosite crust (Hess and Parmentier, In Press). This layer represents a subcrustal heat source beneath the Procellarum-Imbrium terrane and may be responsible for the localized and anomalous enrichments of highly incompatible elements in this terrain. A KREEP liquid layer thickened from 5 to 10 km with more than 300 times chondritic of heat producing elements does not continue to cool but undergoes a reheating and grows in thickness by dissolving several times it mass of anorthosite and ultramafic cumulates. But such a liquid layer cannot give rise to the parent magmas to the magnesian suite, because it cannot account for the extraordinarily high incompatible element content, the primitive major element content, nor the positive ε_{Nd} of the magnesian suite as was discussed previously. This growing layer will also form an impenetrable barrier to the eruption of mare basalts and is generally inconsistent with a number of geophysical constraints of the Moon, specifically the existence of mascons subsequent to the basin filling by mare basalts. A thickened KREEP liquid layer can, however, explain the asymmetry in the distribution of KREEP-rich rock on the lunar crust. Such a KREEP layer must be efficiently cooled and be quickly solidified to avoid some of the difficulties described above. A thinned anorthosite crust would allow efficient heat loss and induce crystallization of such a layer early in lunar history.

Cumulate Overturn on Mars

Our model for the formation and overturn of the mafic cumulates to the lunar magma ocean can also be applied to a Martian magma ocean with interesting geochemical and geophysical consequences (Hess and Parmentier, 2001). But first we must establish whether a magma ocean ever existed on Mars.

A variety of evidence argues for the existence of a magma ocean on Mars. **1)** The parent liquids to the SNC meteorites are significantly depleted in Al_2O_3 and CaO relative to terrestrial basalts [Longhi et al., 1992]. Mare basalts and picrite glasses, derived by melting of the cumulate Lunar mantle have similarly low Al_2O_3 and CaO contents. **2)** Whole rock Rb-Sr isochrons for shergotites preserve the record of a major, Martian-wide differentiation event that depleted the mantle in Rb and other incompatible elements [Jagoutz, 1991]. **3)** ^{182}W and ^{142}Nd anomalies and the apparent age of the Rb-Sr isochron place this differentiation event within 50 million years of the origin of the solar system [Lee and Haliday, 1997; Blichert-Toft et al., 1999]. **4)** The ^{143}Nd composition of the Martian mantle, assuming a single differentiation event at 4.5Ga, is more depleted than even the cumulate source regions of the mare basalts on the Moon [Borg et al., 1997].

5) The average thickness of the Martian crust is about 50 km [Zuber et al., 2000] and, at least in the southern hemisphere, probably dates a massive differentiation event at about 4.5Ga.

We therefore propose that the early differentiation of Mars, like that of the Moon, was dominated by the fractional crystallization of a planet-wide magma ocean which produced a complementary mafic cumulate mantle. The cumulate mantle was compositionally stratified and gravitationally unstable. Solid state differentiation then carried dense, iron-rich and relatively cool cumulates into the Martian interior ultimately resulting in a lower mantle that is denser and chemically more evolved than the upper mantle.

A gravitationally stable compositionally stratified Martian mantle explains a number of important characteristics of Mars. **1)** Mars, like the Moon, is isotopically much more heterogeneous than the Earth [Blichert-Toft et al., 1999; Borg et al., 1997]. While one might argue that the small size of the Moon or the absence of water to promote solid state creep inhibits thermal convection, such arguments do not apply to Mars. **2)** The presence of ^{142}Nd anomalies in some SNCs require that their source regions preserved their initial compositional heterogeneities [Borg et al., 1997]. **3)** The preservation of whole rock Rb-Sr isochrons and the ancient ^{182}W and ^{142}Nd anomalies argue against crustal recycling subsequent to stabilization of the magma ocean cumulates [Blichert-Toft et al., 1999]. **4)** The isotopic evidence of early differentiation in the Martian mantle, in contrast to the evidence from the terrestrial mantle, indicates that the Martian mantle has successfully resisted homogenization by thermal convection. We argue that thermal convection is inhibited by a stably compositionally zoned mantle.

Besides preserving the isotopic and compositional heterogeneity, overturning of an initially unstably stratified cumulate mantle may explain additional aspects of the evolution of Mars (Elkins et al., 2004). After the overturn, the mantle would be stably stratified. Since thermal convection would then be inhibited by the compositional stratification, the primordial and radiogenically generated heat in the Martian interior would be slow to dissipate. Similarly, the infertility of the cumulate mantle and the inhibition of thermal convection could limit magma production by adiabatic decompression, perhaps explaining in part the preservation of ancient crust.

The crustal dichotomy may be the product of the overturn of initially unstable magma ocean cumulates. Horizontal convergence would thicken the crust above long wavelength (spherical harmonic degree one) downwelling of dense late stage cumulates. Thinning due to extension or delamination of the lithosphere in the opposing (northern) hemisphere could induce pressure-release melting and resurfacing. A similar catastrophic resurfacing model has also been suggested for the Venus [Parmentier and Hess, 1992].

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