Kinetics of Melting and Dissolution in Lunar Materials

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An understanding of the petrogenesis of lunar magmas, particularly mare basalts and the parent magmas to the Mg-rich suite, reamains an unfulfulled goal. This fact is not surprising given the complexity of the problem. On the Moon, the source region for lunar magmas is not primitive mantle but rather a series of cumulate rocks that vary widely both in mineralogy and major and minor element contents. The stratigraphy of the cumulate mantle is not likely to be very regular given that the cumulate pile is formed initially in an unstable configuration and subsequent thermal and compositional convection will create heterogeneities on a number of length scales. These lithologic heterogeneities, the large range of pressures and temperatures over which melts are generated on the Moon, and the close juxtaposition of cumulate rock with widely varying solidii introduce significant complications to the nature of the melting relations that control melt generation. These factors, coupled with the likelihood that polybaric fractional melting of varying efficiencies ultimately control the composition of planetary magmas, are ample reasons why the lunar magmas remain the enigma they are. To make progress, phase equilibria studies must be coupled with a detailed understanding of the time scales and the dynamics of crystal and melt reequilibration processes.

I. Petrogenesis of Mare Picrite Glasses

New experimental techniques have been applied to determine the phase equilibria and the dissolution kinetics of ilmenite in mare basalt liquids. In previous procedures we have measured the solubility of ilmenite by simply adding normative $FeTiO_3$ to mare liquids to induce saturation with varying success because such runs suffer from quenching problems; even some TiO_2-rich undersaturated melts produce a feathery melange of quench ilmenite crystals. These effects are particularly deleterious at high temperatures and pressures appropriate for mare liquidus conditions i.e. temperatures approaching and exceeding 1400°C. Under these conditions, ilmenite saturated melts are very fluid, exceptionally dense and contain high contents of normative ilmenite. As a consequence, wide-beam electron probe techniques must be applied to determine and integrate the composition of mare liquids and their quench ilmenites. But this analytical technique is problematical for melts saturated with ilmenite because it's difficult to distinguish and avoid liquidus ilmenite grains from their quenched brethren. In order to mitigate these problems, we returned to the experimental techniques developed by Hess and Finnila (1997) wherein ilmenite saturation was imposed by dissolving ilmenite into a given ilmenite-undersaturated mare picrite melt.

A dissolution couple is formed by packing powders of a picrite mare glass over a layer of synthetic ilmenite powder in a cylindrical graphite capsule. As ilmenite dissolves into the melt the liquid becomes density stratified and stable with respect to compositional convection. The concentration profiles are modelled with effective diffusion coefficients applied to a system with a moving boundary, thus we can obtain both solubility and diffusivity data. The critical experimental advantage of this technique is that all melts are generally undersaturated or just saturated with ilmenite (exceptions are noted below) so that ilmenites recovered after the quench are quench phases. After run durations of more than two hours, the concentration profiles are flat and record the solubility of ilmenite in the chosen mare basalt composition at the given P and T.

But additional solubility data are obtained as the ilmenite-basalt couple evolves into the final equilibrium described above. Melt-ilmenite compositions at their interface evolve with time but remain in local equilibrium (Liang, 2001; in prep). Because the composition of the boundary layer liquid is limited by the respective diffusion coefficients of the major diffusing components and not by the kinetics of the dissolution of ilmenite, these liquids are ilmenite-saturated but have compositions that depart significantly from the far-field compositions obtained at equilibrium. Nevertheless, with the assumptions of local equilibria the compositions represent loci of points on an ilmenite-saturated surface.

As an illustrative example, the composition of a picrite mare basalt melt in local equilibrium with ilmenite at 1350°C, 1.0 GPa contains about 32% TiO_2 and 20% SiO_2 (Fig. 1). With time, the composition of the interfacial liquids as well as the far-field liquid reaches the constant composition of 26% TiO_2 and 26% SiO_2. The contrast in composition between the local mare basalt liquid composition in the short duration experiments and the final equilibrium state are controlled by the variable diffusivities of the liquid components, particularly SiO_2, TiO_2 and to a lessor degree FeO.

As already emphasized, the compositions of ilmenite-saturated mare basalt melts under liquidus conditions in the lunar interior are very enriched in normative ilmenite. For example, the ilmenite-saturated Apollo 15 red glass composition contains about 26% TiO_2 at 1350, 1 GPa; increasing temperature, increases the solubility whereas increasing pressure acts to decrease the solubility; an increase in pressure of 1.5 GPa approximately offsets an increase in temperature of 100°C. Since high Ti mare glasses contain less than 17% TiO_2 (Shearer and Papike, 1993) and the slope of the liquidus at high pressures is steeper than the slope of the TiO_2 isopleth for the solubility of ilmenite, it follows that mare basalts and their glasses are more than 50% undersaturated with respect to ilmenite under near liquidus conditions in the lunar interior. The question is why?

One possible explanation lies in the extraordinary densities and fluidities of such magmas. Densities of ilmenite-saturated magmas corrected for high pressures exceed 3.5 gm/cm^3 and viscosities are less than a few poise. Depending on the permeability and wetting characteristics of the cumulate mantle, ilmenite saturated magmas should migrate towards the lunar interior. Assuming that the lunar interior is pyroxene-rich (i.e. low CaO-pyroxene) as the cumulate overturn model predicts (Hess and Parmentier, 1995), the silicate mantle would have considerable capacity to extract normative ilmenite from the descending streams of picrite liquids. Experiments suggest that low CaO pyroxene in equilibrium with TiO_2-rich picrite melts contains 2-3 wt% TiO_2 (Unpublished; Delano, 1980). Given that the TiO_2 partition-coefficient for orthopyroxene/melt is D=0.1-0.2 (Xirouchakis et al., 2001) it follows that the cumulate orthopyroxene derived from the TiO_2-poor magma ocean would contain very little TiO_2. Similar arguments predict that relatively primitive orthopyroxenes would also contain little Al_2O_3, CaO and variously incompatible trace elements. These primitive orthopyroxenes are therefore good sinks for such normative components. The downgoing melts would undergo complex dissolution/recrystallization processes and compatible and incompatible elements would be variously fixed into the cumulate mantle under exchange equilibria controlled by chromatographic fractionation. It can readily be appreciated that the evolution of the chemical fronts for compatible and incompatible elements would evolve differently; the result is that the cumulate mantle would be metasomatized in such a way that compatible and incompatible elements would be fractionated/decoupled from each other. It is also likely that this mantle, although variously enriched in normative ilmenite, would have substantial regions/thicknesses where the cumulates are undersaturated with respect to ilmenite. The TiO_2 contents of partial melts from these metasomatized regions would then depend on the initial TiO_2 contents and the degree of melting. Similar conclusions apply to the moderately incompatible oxides CaO and Al_2O_3 as well as the highly incompatible oxides. The highly compatible elements, however, would vary less with the degree of metasomatism. The efficiency of these enrichment processes, however, depend on a number of variables and processes that must be investigated with experiments and theoretical analyses.

II. 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, 2002, 2000a)! It follows that the light REE enrichment of the Mg-rich parent magmas cannot be attributed to simple 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, 2001). 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 its 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.

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