SPREADING CENTER PROCESSES UNDER VENUS CONDITIONS: IMPLICATIONS FOR CRUSTAL FORMATION, PETROLOGY, AND STRUCTURE; Paul C. Hess and James W. Head, Department of Geological Sciences, Brown University, Providence, RI 02912

Evidence has been previously presented for crustal spreading in the Aphrodite Terra region of Venus[1] and several analyses have discussed the nature of spreading processes under Venus environmental conditions.[2,3] Here we further assess the nature of crustal spreading centers under Venus environmental conditions.

STRUCTURE AND GEOMETRY OF SPREADING REGION: The structure of the crust-mantle interface at a Venusian spreading center can be modelled with some simple observations obtained from terrestrial analogues. Fast-spreading oceanic ridges on Earth contain shallow axial magma chambers that reside only a few kilometers beneath the ocean floor.[4] The axial magma chamber beneath the East Pacific Rise, for example, extends to within 1.4 km of the ocean floor. The liquid portion of the chamber is quite small, and consists of a thin lens, perhaps no more than a few hundred meters thick and approximately 4 km wide. This pool of magma is surrounded by a wider reservoir of hot and probably partially molten rock which is at least 6 km wide and deep[5]. This igneous body isostatically supports a ridge about 200-400 m high and 8 km wide. While the reservoir of hot rock is continuous along the ridge, the axial pool of magma is discontinuous. The gaps in the axial magma chamber are typically associated with low points on the ridge and in the disappearance of axial summit graben.[6] No evidence of high level magma chambers has been found on slowly spreading ridges. A microseismic study on the axis of the Mid-Atlantic Ridge south of the Kane fracture zone, for example, determined that the brittle plate thickness extended to about 8 km depth. These data are consistent with various models of the oceanic crust.[7] For very small spreading rates of 1 cm/yr or less, lateral heat conduction is sufficient to efficiently dissipate the latent heats released by the crystallization of basaltic liquid. Under these conditions, liquids are frozen into sheeted dike complexes and high level magma chambers can have only an ephemeral existence. In general, cooling is more rapid near the ridge axis at low spreading rather than fast spreading rates. It is noteworthy, nevertheless, that the size of the subaxial igneous body and particularly, the actual liquid portion of this body are very small for even fast spreading ridges. It is easy to understand how difficult it is for permanent shallow magma bodies to exist under slowly spreading ridges.

Slowly spreading ridges on Venus may not suffer the same fate as slowly spreading terrestrial ridges. Cooling near the ridge axis on Venus will be less severe for a number of reasons.[2] The most obvious difference is that the surface on Venus is more than 450° C higher than on Earth. Perhaps even more significant is the absence of hydrothermal cooling, an important factor determining the thermal structure of terrestrial spreading centers. According to Phipps Morgan et al.[8] hydrothermal cooling transfers heat about 10 times more efficiently than that due to conduction alone. The absence of hydrothermal cooling on Venus combined with the elevated surface temperatures should make the axial regions much hotter than on Earth.[2] Finally, the elevated interior temperatures on Venus, and the resulting high levels of associated melting, transfers more latent heat to the axial regions. We have estimated that two to three times the thickness of basaltic crust is produced resulting in a proportionally equivalent transfer of latent heat.[2,3] It may be concluded, therefore, that large, high level magma chambers may be more likely on Venus than on Earth. This does not mean, however, that the chambers are largely molten. A wide high level magma chamber composed of basaltic liquid covered by a lid of dense basaltic crust would be mechanically unstable.[7] The roof to the chamber is subject to collapse and vertical downdropping. This would be accompanied by escape of liquid to the surface and the rapid cooling of magma. A chamber filled with hot but viscous cumulate mush would not be as susceptible to these stoping processes.

A consequence of a hotter mantle and elevated surface temperatures is that old lithosphere on Venus would be thinner than on Earth for equivalent material properties of mantle. Solomon and Head[9] argue that the flexural topographic profile of Freya Montes foredeep is consistent with a thermal gradient of 20-25 K/km and a thin mechanical lithosphere 15-20 km thick beneath typical plains regions. Other estimates based on simple conductive models of lithospheric heat transfer yield thicknesses of the thermal and elastic lithosphere of 40 km and 10 km respectively.[10] The thickness of the mechanical lithosphere is bounded by these values. From a petrological perspective, the most interesting measure of the lithosphere is the part that lies within the convective regions of the mantle. Parsons and McKenzie[11] suggest that the maximum thickness of the thermal boundary layer (bounded by the 1350°C $\pm$50 isotherm) is controlled by a convective instability which removes cold material from the base of the old plate and replaces this mass with hotter mantle from below. Thus, the appropriate measure of lithosphere thickness on Venus is that obtained for the thermal lithosphere or about 40 km. This is approximately 33% of the estimate for terrestrial oceanic lithosphere.

The zone of melting beneath a spreading center is bounded by the solidus of the mantle, the adiabatic temperature gradient and the cooling of the thermal boundary layer with the square root of age. For typical terrestrial conditions, melting begins at about 50-60 km below the ridge axis. The melt zone in two dimensions is therefore confined to a triangular area with a base of about 200-300 km and height of 50 km. In comparison, the melt zone under comparable conditions on Venus begins at a depth of about 80 km. If old lithosphere is approximately 40 km thick, it follows that the zone of melting is not simply confined to an area below the ridge axis but would exist under the lithosphere wherever passive or buoyant upwelling mantle crosses the solidus at depths greater than 40 km.

PETROLOGIC IMPLICATIONS: A consequence of the combined effects of a hotter mantle and a thin lithosphere would be a pervasive zone of deep-seated melting within a large radius of upwelling mantle. The formation of crust is not primarily at the ridge axis, as on terrestrial oceanic spreading centers, but would be widely distributed. Whereas the average composition of melt at the ridge axis includes contributions of melt parcels produced throughout the melt column, the composition of tholeiite basalt produced in the off-axis regions is much more primitive. Between depths of 80 and 40 km, only picritic melts of relatively low SiO_2 and high MgO contents are developed. In the axial regions, melt permeates the mantle by porous flow mechanisms until it is collected into macroscopic bodies. This process tends to homogenize and buffer the melt compositions to some grand average. In contrast, melts in the off-axis regions must be able to penetrate 40 km of lithosphere to reach the Venusian crust. This cannot be achieved by porous flow as the melts would quickly freeze within the lithosphere. Melts extruded to the surface from these regions must have access to conduits through the lithosphere. Only a small fraction of the melt from the sublithospheric mantle can successfully complete this transit. It is possible that volcanism occurs only after some critical state of stress is exceeded and a portion of the lithosphere ruptures catastrophically. This will act as a pressure release and serves to drain a large region of the mantle of its melt component. Volcanism, while episodic and widely spaced, is expected to occur locally as massive outpourings of flood basalt. The basalt will not have time to differentiate and will bear the primitive signature of its origin. The Venera 14 composition may be an example of this type of volcanic event.

A thin lithosphere also does not favor the production of silica undersaturated basalts of the alkali basalt-nephelinite association in the plains regions of Venus. Two conditions must exist to favor the formation of silica undersaturated basalts from typical peridotite. These are high ambient pressures and high partial pressures of CO_2. Anhydrous melts coexisting with the lherzolite assemblage olivine-orthopyroxene-clinopyroxene are quartz tholeiites for pressures up to 5 kb, olivine tholeiite at 8-12 kb, and alkali olivine basalt (silica undersaturated basalt) at pressures from 15 to 20 kb.[12] Adding CO_2 to lherzolite shifts the melt compositions to slightly less SiO_2-rich compositions at all pressures but the most dramatic effects occur only at pressures above 20 kb where highly SiO_2-undersaturated melts are generated.[13]

Mantle that is advected passively into the melting zone on Venus produces only tholeiite basalt. Mantle plumes that experience buoyant advection are hotter than average mantle and therefore experience melting at much greater depths. While the first melts produced at these great depths may indeed be SiO_2-undersaturated, they will quickly lose their SiO_2-undersaturated signature as pressure release melting produces increasing volumes of tholeiite basalts, and as the undersaturated melts continually re equilibrate with mantle at shallow depths. The only way that such undersaturated melts could reach the surface in pristine condition is if the melts were isolated into discrete masses and quickly erupted through fractures to the Venusian surface. While this cannot be ruled out, it is difficult to see how such fractures could develop in extremely hot, partially molten mantle of low viscosity.

Alkali basalts on Earth survive the transit to the surface because the mechanical lithosphere extends to depths of 100 km or more. SiO_2-undersaturated melts produced below this depth could have access to the surface through fractures and conduits in the relatively rigid lithosphere. The mechanical lithosphere on Venus is only 40 km thick and is too far from the source where alkali basalt are generated. Thus, alkali basalts on Venus must somehow reach the safe haven of the lithosphere by traversing tens of kilometers of tholeiite-bearing mantle. Alkali basalts should, therefore, occur only near anomalously thickened crust and lithosphere.

IMPLICATIONS OF SLOW SPREADING RATES: Half-spreading rates in Aphrodite Terra may be very small,[1,2] perhaps around 0.5 cm/yr[2]. For typical terrestrial half-spreading rates (V>1 cm/yr), the rate of upwelling of mantle is rapid and the geotherm is determined by the advection of heat and the latent heat associated with phase changes. In the absence of movement or for small parcels of mantle that are moving very slowly, heat is lost primarily by conduction and temperatures will drop below those obtained along the adiabatic temperature gradient. The amounts of melt produced under these conditions are much less than obtained through adiabatic upwelling. The question as to which type of heat transport dominates in the mantle is determined by the thermal Peclet number, $Pe = Vl/K$, where V is the velocity of upwelling mantle, l is a length scale and K is the thermal diffusivity. If $Pe >> 1$ advective heat dominates and if $Pe << 1$ heat is lost primarily by conduction. For a spreading rate of $V = 5 \times 10^{-2}$ m/yr, a thermal diffusivity of 30 m^2yr and $Pe = 1$, l is 6 km. This means that mantle within 6 km of the surface or a colder plate boundary will cool primarily by conductive heat loss. This thickness is less than the estimated thickness of basaltic crust on Venus so that melting beneath even slow spreading ridges is governed by advection. The melting zone, however, is somewhat narrower than that estimated earlier, and the total amount of melt produced in slowly spreading ridges should be less than that of rapidly spreading ridges under otherwise identical conditions. Somewhat higher potential temperatures are necessary to generate a given thickness of basaltic crust below a Venusian spreading center than on Earth.

References: 1) J. Head and L. Crumpler (1987) Science, 238, 1380; 2) C. Sotin *et al.* (1989) EPSL, 95, 321; 3) P. Hess *et al.* (1989) LPSC XX, 406; 4) Burnett *et al.* (1989) Nature, 339, 206; R. Detrick *et al.*(1987) Nature, 326, 35; McClain *et al.* (1985) JGR, 90, 8627; 5) K. Macdonald (1989) Nature, 339, 178; 6) K. Macdonald and J. Fox (1988) EPSL, 88, 119; 7) N. Sleep (1975) JGR, 80, 4037; N. Sleep and Rosenthal (1979) JGR, 84, 3831; N. Sleep and B. Windley (1982) Jour. Geol., 90, 363; 8) J. Phipps Morgan *et al.* (1987) JGR, 92, 12,823; 9) S. Solomon and J. Head (1989) LPSC XX, 1032; 10) S. Solomon and J. Head (1982) JGR, 87, 9236;11) B. Parsons and D. McKenzie (1978) JGR, 83, 4485; 12) E. Takahashi and I. Kushiro (1983) Am. Min., 68, 859; 13) Falloon and Green, (1988); P. Wyllie (1980) JGR, 85, 6902.