MELTING OF VENUSIAN MANTLE AT SPREADING CENTERS: P.C. Hess, J.W. Head, E.M. Parmentier , Dept. of Geological Sciences, Brown University, Providence, RI 02912

Parameterized convection models of Venus with an earth–like mantle (i.e., similar internal structure, initial radioactive heat production, and common elastic and thermal parameters) indicate that Venus' mantle is about 100°K hotter than the earth's mantle (1,2,3). The principal difference affecting the mantle temperatures are the surface temperatures of the two planets and the possibility that Venus is a one plate planet. Recently, evidence has been presented for the presence of features in Western Aphrodite Terra analogous to spreading centers on Earth (4). It is possible, therefore, that the thermal evolution of Venus, and its recent geological past, may have been influenced by thermal-tectonic forces comparable to those existing below present day terrestial ocean ridges. Initial analyses have shown that terrestrial spreading centers mapped to Venus' conditions result in the production of thicker crust, and that the gravity and topography data for Western Aphrodite are consistent with a spreading center model. (5).

Here we examine in more detail the petrological implications of the temperature and flow regime of the Venus mantle with earth-like sub-oceanic mantle characteristics but subject to higher average temperatures. A hotter mantle should be more extensively melted and will govern both the chemistry of the ridge basalts and the thickness of oceanic crust. Melting beneath ocean ridges on Earth probably is related to the adiabatic upwelling of mantle although this does not necessarily mean that the ridges lie above the upwelling limbs of convection cells (6). The geotherm controlled by the temperature of the upwelling mantle must be capable of producing oceanic crust 6 km thick (6). The amount of melt produced is estimated by superimposing the geotherm upon the melting phase relations of peridotite and by assuming a flow regime of the mantle beneath ocean ridges.

The melting phase relations are taken for a fertile peridotite after Takahashi (7). The solidus is approximated by:

1) $T°C = 1150°C + 12.5\ P(Kb)$

the diopside out curve by:

2) $T°C = 1200°C + 12.5\ P(Kb)$

and the orthopyroxene out curve by:

3) $T°C = 1300°C + 12.5\ P(Kb)$

Equation (2) corresponds to approximately 20% (by Wt) melt and (3) to 40% melt. T vs X (% melt) curves are assumed to vary linearly between (1) and (2), (2) and (3) and between (3) and the liquidus. The latent heat of fusion is not known but is estimated at $\Delta H = 150$ cal/gm to 120 cal/gm which is below that of forsterite (222 cal/gm) (8), above fayalite (105 cal/gm) (9) and close to diopside (154 cal/gm) (9). The lower value reflects the fact that the melting relations have a peritectic character. The heat capacity of peridotite is estimated as Cp = 0.3 cal/gm (9).

The mantle flow regime and the structure of the oceanic ridge is taken from (10), although in simplified form (Fig. 1). It is assumed that all melt within the melting zone migrates laterally and is applied to the growing oceanic crust at the ridge axis. The model probably results in an overestimation of crust thicknesses.

Fig. (2) gives the thickness of basalt crust obtained along adiabats intersecting the mantle solidus at 1300° to 1600°C, and using $\Delta H_{Fus} = 150$ cal/gm. Two curves are shown. The equilibrium melting model assumes that melt remains with the upwelling mantle until it is extracted to form crust. The fractional melting model assumes that the melt is efficiently removed as soon as it is formed. In this case, melting in the upwelling mantle stops as soon as diopside is consumed (curve (2)) unless temperatures are greater than about 1550°C at 1 atm.

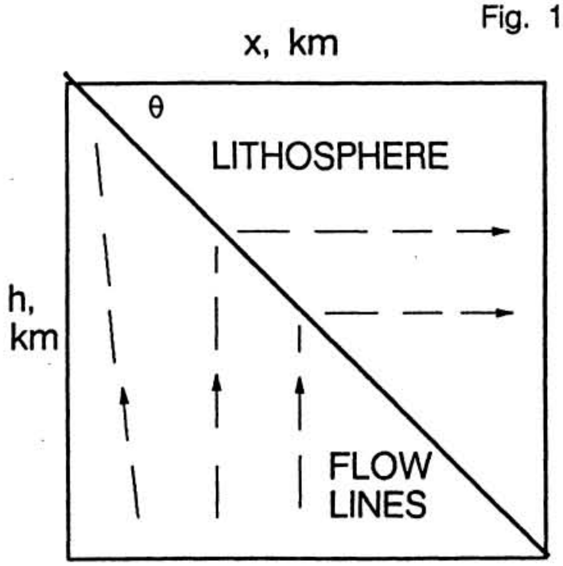
Oceanic crust of 6 km is thus produced by an adiabat that intersects the solidus at 1350°C and 16 kb, by either fractional or equilibrium melting. Lowering the heat of fusion to 120 cal/gm lowers the required adiabat by 20 – 30°C. However, if not all melt is flushed from the mantle, then higher initial temperatures are required. It is not likely, however, that more than 1400°C is required. Crustal thicknesses obtained for Iceland (~14 km) require intersections of about 1470°C and 1520°C for equilibrium and fractional melting respectively.

A Venus ocean ridge is modelled by increasing the adiabat to 1450°C. Under these conditions, melting begins at about 24kb and crustal thicknesses are expected to lie between 11.5 to 14.5 km for fractional and equilibrium melting respectively (for ΔH = 150 cal/gm).

An Icelandic hotspot on Venus with an adiabat about 100°C greater than "normal" oceanic mantle should create between 16 to 24 km crust for the two melting models. Melting on Venus, therefore, involves deeper intersections of the adiabat with the mantle solidus, larger average degrees of melting, and melts with greater MgO content than Earth analogs. Using the results of (11), the pooled compositions of terrestrial basalts obtained by the intersections near 15Kbar, produce tholeiites with about 10% MgO for equilibrium melting. The mean density of the crust is estimated at 2.9 gm/cm^3. In contrast, the basalts on Venus obtained at intersections near 25 kb are tholeiites with about 12 to 13% MgO. Mean crustal density (calculated volatile-free and unmetamorphosed) are about 2% greater. In general then, the "oceanic" crust on Venus is believed to be 2 to 2.5 as thick as that on Earth, and considerably more olivine normative.

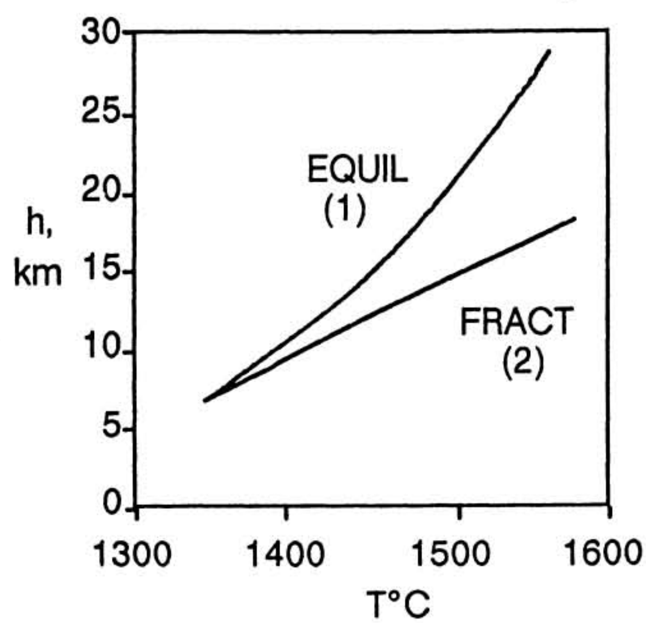
References: (1) Stevenson, D.J. et al. (1983) Icarus, 54, 466-489; (2) Phillips, R.J. and Malin, M.C. (1983) in Venus, D.M. Hunter et al., Eds, 159-214; (3) Turcotte, D.L. et al. (1979) Proc. Lunar Planet Sci Conf 10th, 2375-2392; (4) Head, J.W. and Crumpler, L.S. (1987) Science, 238, 1380-1387; (5) Sotin, C. et al. (in press) EPSL; (6) McKenzie, D. (1984) J. Pet, 25, 713-765; (7) Takahashi, E. (1986), JGR, 91, 9367-9382; (8) Stebbins, J.F. et al. (1984) Cont Miner Petrol, 86, 131-148; (9) Richet, P and Bottinga, Y. (1986), Rev Geophysics, 24, 1-25; (10) Ahern, J.L. and Turcotte, D.L. (1979) EPSL, 45, 115-122; (11) Klein, E.M. and Langmuir, C.H. (1987) JGR, 92, 8089-8115.

Fig. 1



Strucuture of sub-ridge mantle

Fig. 2



Thickness of basaltic crust (h) produced by equilibrium (1) and fractional (2) melting of peridotite by adiabatic upwelling of mantle at T°C.