DERIVATION OF PRIMARY MAGMAS AND MELTING OF CRUSTAL MATERIALS ON VENUS: SOME PRELIMINARY PETROGENETIC CONSIDERATIONS*

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Abstract. Crustal formation and evolution processes are of critical importance in the geochemical and thermal evolution of planets. As an aid to understanding these processes on Venus, we develop a general paradigm for: (1) the derivation of primary magmas, and (2) the range of possible conditions for remelting of crustal materials and the evolution of the products of remelting. We use as a basis for this paradigm the present knowledge of the bulk and surface composition, thermal structure, and surface geological and geochemical processes. For the range of conditions of derivation of primary magmas and crustal remelting, a wide range of magma types is possible, and no magma type can be arbitrarily excluded from consideration on Venus. We conclude that magmatic and volcanic activity on Venus, in its broadest sense, could be very similar to that on the Earth, although eruption styles are expected to vary due to environmental conditions (Head and Wilson, 1986). Major differences in magmatic and volcanic activity are likely to occur in two environments on Venus: (1) those analogous to terrestrial island arcs, where due to the absence of water, melts should be SiO_2-undersaturated, and the more fluid melt products may produce widespread deposits of SiO_2-poor ferrobasalts rather than more viscous SiO_2-rich magmas and composite volcanoes, and (2) those in plains regions influenced by mantle plumes and hot spots, where highly picritic melts may periodically flood vast regions of the surface.

1. Introduction

On the basis of recent spacecraft investigations, the nature of the surface of Venus (Pettengill *et al.*, 1980; Masursky *et al.*, 1980; Campbell *et al.*, 1984; Barsukov *et al.*, 1986; Basilevsky *et al.*, 1986), its composition (Surkov *et al.*, 1983, 1984. 1987) and atmosphere (Donahue and Pollock, 1983), and atmosphere-surface interactions (Barsukov *et al.*, 1982; Nozette and Lewis, 1982) have become more well known. The emerging picture of geologic and geophysical processes operating to shape and modify the surface and interior provides a basis for the discussion of petrogenesis and petrogenetic processes on Venus, following some early considerations (e.g., Warner, 1983).

Data from the Pioneer-Venus spacecraft (Pettengill *et al.*, 1980; Masursky *et al.*, 1980), from Earth-based observatories (Campbell *et al.*, 1983, 1984, 1989; Goldstein *et al.*, 1978), and from Soviet Venera landers (Surkov *et al.*, 1983, 1984, 1987) and orbiters (Barsukov *et al.*, 1986; Basilevsky *et al.*, 1986; Sukhanov *et al.*,

1989) have provided a more detailed view of Venus. These data have revealed evidence for hot spots (Pronin, 1986; Morgan and Phillips, 1982; Stofan *et al.*, 1988), crustal spreading (Head and Crumpler, 1987; Sukhanov and Pronin, 1988; Crumpler and Head, 1990), plains volcanism and volcanic edifices (Campbell *et al.*, 1989; Aubele and Slyuta, 1990; Slyuta, 1990), and orogenic belts, convergence and crustal thickening (Crumpler *et al.*, 1986; Vorder Bruegge and Head, 1989; Head, 1990a). Attention has also been paid to processes of crustal formation and evolution (Head, 1990b; Kaula, 1990) and to estimates of its thickness (Zuber, 1987; Grimm and Solomon, 1988). On the basis of this emerging picture, we develop a paradigm for the range of conditions for the production, evolution, and destruction of crustal material, and the nature of related petrogenetic processes on Venus. We ask such questions as: What is the range of magma types possible on Venus?; Is there any reason to suspect that basic processes of magmatism might be radically influenced by the bulk composition or thermal structure of Venus?; What is the range of processes for the production, evolution, and destruction of crustal material, and what are the related petrogenetic processes?; What is the influence of volatiles on petrogenesis, both in terms of mantle volatiles, and those derived from surface/atmosphere interactions and recycling?; What does the range of basic petrogenetic processes imply about the long-term evolution of the crust of Venus? This paradigm is designed to be of help in focusing future detailed analyses of specific petrogenetic environments and in making predictions and tests for further exploration.

We begin by reviewing the available information on the composition of Venus in order to consider the range of source materials for melting, and we then discuss the actual melting processes in various geological environments and under a range of temperature/pressure conditions. We consider melting of mantle material, melting of differentiates, and magmatic differentiation processes and products. Finally, we examine the range of processes and magma types for Venus and make predictions on the types and styles of activity that should exist in different environments.

2. The Composition of Venus and the Nature of Source Materials for Melting

The composition of primary melts depends on a number of critical parameters, few of which are known for Venus with any certainty. The most important variables include the bulk composition and mineralogy of the source region, the depth and temperature of melting, the composition and abundance of volatile components such as CO_2 and H_2O, and the oxidation state. Presently, the composition of Venus is constrained only by the bulk density and the determination of elemental content of rocks by the Venera space probes (Surkov *et al.*, 1983, 1984, 1987; Nikolayeva, 1990) (Table I). The abundances of U, Th, and K obtained by gamma-ray spectroscopy are more similar to those of terrestrial rocks than either lunar rocks or meteorites. The chemical composition of the rocks from the Venera 13 landing site resembles terrestrial alkali basalts whereas those from the Venera 14

TABLE I

Oxide (wt%)	Venera 13[1]	Venera 14[1]	Vega 2[1]	MORB[2]	Alkali basalt[3]
SiO$_2$	45.1±3.0	48.7±3.6	45.6±3.2	49.7	45.1
TiO$_2$	1.6±0.45	1.25±0.41	0.2±0.1	0.7	2.0
Al$_2$O$_3$	15.8±3.0	17.9±2.6	16±1.8	16.5	13.4
FeO	9.3±2.2	8.8±1.8	7.7±3.7	8.3	12.7
MnO	0.2±0.1	0.16±0.08	0.14±0.12	–	–
MgO	11.4±6.2	8.1±3.3	11.5±3.7	9.0	11.5
CaO	7.1±0.6	10.3±1.2	7.5±0.7	13.8	10.4
Na$_2$O	ND	ND	ND	1.9	2.9
K$_2$O	4.0±0.63	0.2±0.07	0.1±0.08	0.1	0.7
SO$_3$	1.62±1.0	0.88±0.77	4.7±1.5	–	–
Cl	0.3	0.4	<0.3	–	–
Totals	96.4	96.7	93.7	100	98.7

[1]X-Ray Fluorescence Spectrometer Data for the Surface of Venus (Surkov *et al.*, 1983).
[2]Glass composition of typical primitive mid-ocean ridge basalt (MORB) (Hess, 1989).
[3]Representative composition of alkali basalt (Hess, 1989).
ND=not determined (Surkov *et al.*, 1983).

landing site bear a strong similarity to ocean floor tholeiite basalts (Surkov *et al.*, 1983) (Table I). Goettel *et al.* (1981) showed that the minor density differences between the Earth and Venus can be explained by assuming a deeper basalt-eclogite transition and a somewhat lower FeO content. Notwithstanding these fine points, it is likely that the Earth and Venus have similar major element compositions (see discussion in Phillips and Malin, 1983). The main differences, at least from a petrological perspective, might reside in the volatile inventory of the mantles of the two planets (McGovern and Solomon, 1989). These differences could have originated during the accretion of the planets and/or during later thermal evolution, degassing, or recycling of metamorphosed highland materials.

On the basis of these overall considerations, the source materials for primary magmas on Venus are assumed to be: (1) peridotite mantle, (2) peridotite mantle that has been modified to varying degrees by recycling, (3) tholeiite basalt, and (4) tholeiite basalt variously metamorphosed by near-surface processes. These sources are consistent with our present understanding of Venus. We now consider the origin of primary and derivative magmas on Venus subject to the following assumptions: (1) Parameterized convection models of Venus with an Earth-like mantle indicate that Venus' mantle is about 100–150 °C hotter than the Earth's mantle (Stevenson *et al.*, 1983; Phillips and Malin, 1983); (2) Because of the elevated surface and interior temperatures and the higher degree of melting, it is estimated that basaltic crust produced in a spreading center environment would be about 10–20 kilometers thick (Sotin *et al.*, 1989; Hess *et al.*, 1989); (3) The mechanical lithosphere is much thinner than on Earth, perhaps only 30–40 km in thickness (Solomon and Head, 1989; Hess and Head, 1990); and (4) Underthrust or subducted crust reintroduces CO$_2$ but not H$_2$O into the mantle.

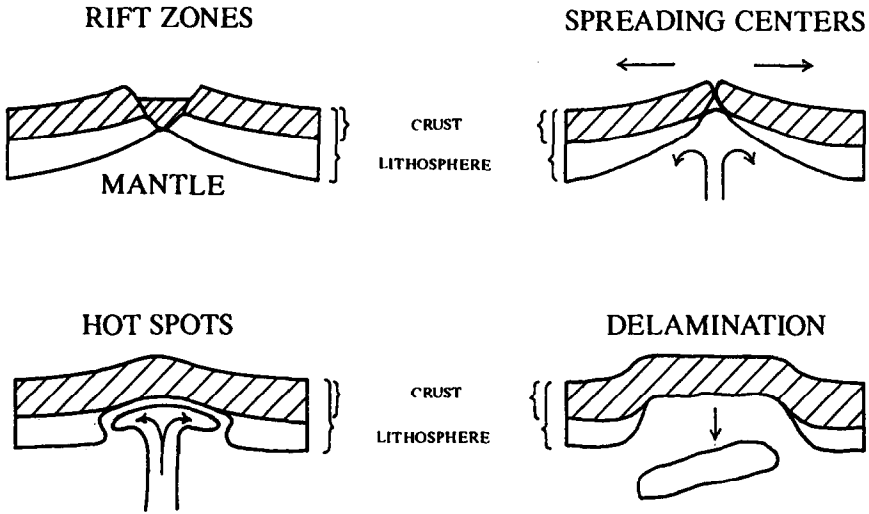RIFT ZONES   SPREADING CENTERS

Fig. 1.   Potential environments of primary melting of mantle material (rift zones, spreading centers, hot spots, delamination) on Earth and Venus.

3. Melting of Mantle Material Peridotites

Melting of mantle material may occur below a variety of surface environments (Figure 1), as is seen on the Earth, including rift zones, oceanic divergent plate boundaries, intraplate areas where subcrustal mantle melting produces localized accumulations of melt products ('hot spots'), and in zones of convergence and crustal thickening where mantle isotherms have been perturbed by processes such as delamination.

The compositions of liquids produced by the melting of peridotite are relatively well constrained. A useful exercise is to characterize the MgO contents of primary liquids in equilibrium with peridotite at various temperatures and pressures (Figure 2). The liquids of melts with MgO up to 20% were obtained from experimental partial fusion of peridotite (Falloon et al., 1988; Falloon and Green, 1987, 1988), and the liquidus and solidus are for a fertile peridotite with about 39% MgO (Takahashi, 1986; Herzberg et al., in press). The 25% and 30% MgO contours are derived from crystallization experiments on komatiites (Bickle et al., 1977; Arndt, 1977; Agee and Walker, 1988) and do not necessarily equate to equivalent liquidi obtained by melting peridotite. The differences between the true liquidi and those described in Figure 2 are unlikely to be major, however, at least with respect to values of MgO, and the P and T coordinates.

Although a simple phase diagram is not universally applicable, it can be used to derive some general guidelines. Note that the MgO contents of near-solidus liquids increases dramatically with pressure. This observation is certainly valid, and the MgO contents are probably accurate to within 10% of the values shown in
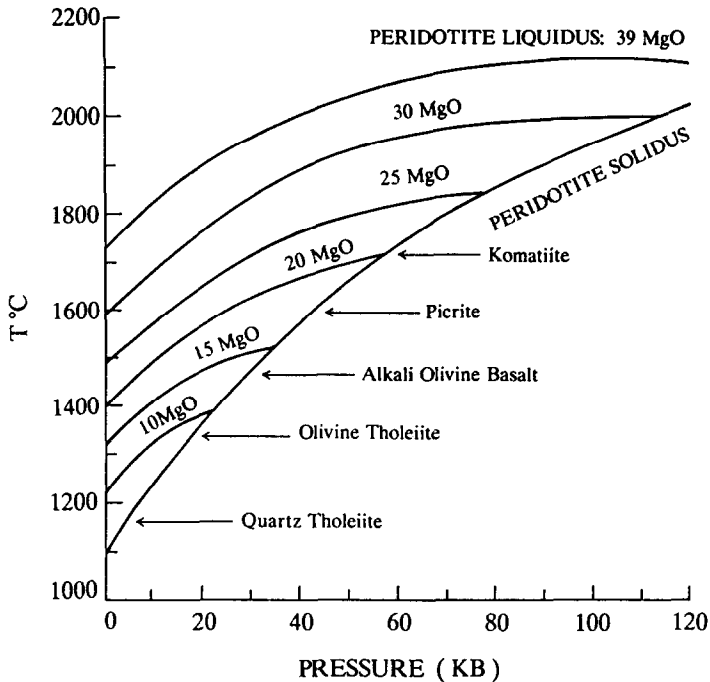
Fig. 2. Liquidus and solidus for a peridotite with approximately 39% MgO (modified after Takahashi, 1986). Contours give MgO contents of liquids obtained by partial melting of peridotite.

Figure 2. Additional information available from high pressure melting experiments constrains the composition of near solidus melts; these are liquids in equilibrium with peridotite to within about 50°C of the solidus. Quartz tholeiites, typically with more than 50% SiO_2, coexist with peridotite at low pressures up to about 5-6 kb. With increasing pressure, the liquidus volume of olivine contracts, and liquids first become olivine- and then nepheline-normative. Olivine tholeiites, typical of the terrestrial ocean floor (MgO ⩽ 10% MgO) are stable from about 7–15 kb, and gradually become more SiO_2-undersaturated and assume the characteristics of alkali olivine basalts with MgO ~10–13% between 15–30 kb. With further increases in pressure from 50 to 100 kb, the melts are enriched in normative olivine, evolving first to alkali picrites, then to basaltic komatiites (MgO ~18–25%) and finally to peridotitic komatiites (30% MgO). Note, however, that these pressure ranges are only approximate and should not be applied as universal constraints to basalt petrogenesis.

The near-solidus melts which represent relatively small degrees of melting are enriched in highly incompatible elements. The melts produced at temperatures and pressures significantly above the solidus become enriched in compatible elements and have their incompatible element budget diluted. Thus, melts enriched with incompatible elements imply an origin by small degrees of melting or the

existence of anomalous mantle locally enriched with incompatible elements. In-
creasing amounts of melting also gradually transforms nepheline-normative melts
to olivine-normative ones. Nepheline-normative alkali olivine basalts, which are
near-solidus melts at 15–30 kb, gradually evolve to olivine tholeiites or even
picrites at larger degrees of melting. Strongly SiO_2-undersaturated melts are pro-
duced by relatively small degrees of melting of volatile-free peridotite.

The effect of CO_2 on the solidus of peridotite is most dramatic at pressures
above about 19 kb (Figure 3) where a carbonate phase is stable (dolomite or
magnesite) (Falloon and Green, 1989). Below this pressure, peridotite coexists
with a free CO_2 fluid. The solubility of CO_2 in silicate melt is very low at these
pressures and the solidus for lherzolite-CO_2 is only 20–50 °C below that for the
CO_2-free solidus. The compositions of the near-solidus melts are not known but
because of the low CO_2 solubility (~1% CO_2 at 20 kb; Pan and Holloway, 1990),
their compositions should not depart significantly from those obtained from CO_2-
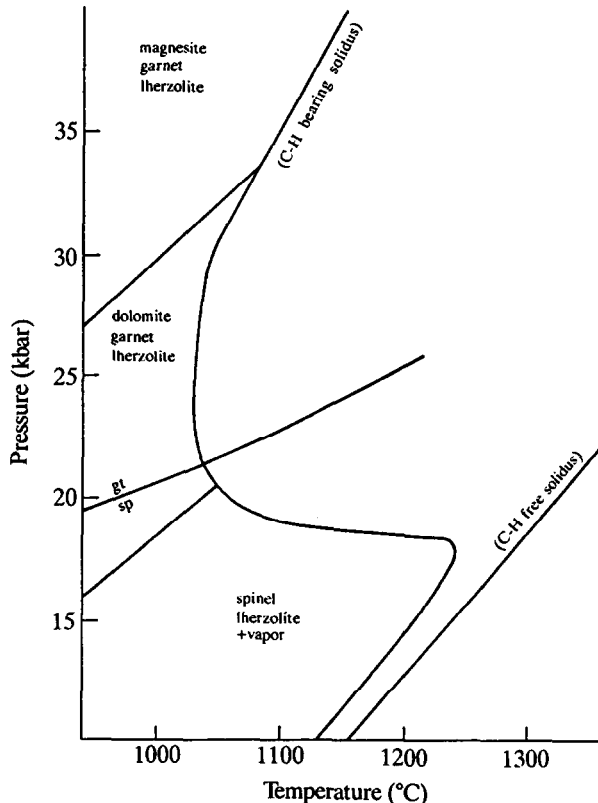free peridotite. The solidus above 19 kb is depressed by more than 400 °C from

Fig. 3. Phase relationships for Hawaiian pyrolite (95 wt.%) plus dolomite (5 wt.%) illustrating
CO_2 + peridotite relationships (modified from Falloon and Green, 1989, 1990).

the CO_2-free solidus, shifts that are comparable in magnitude to those obtained under H_2O-saturated conditions. There is no free CO_2 fluid phase in an olivine-bearing mantle at these pressures as all CO_2 is contained within dolomite or magnesite. The near-solidus melt, moreover, is not basaltic or even largely of silicate composition. It is a carbonatite melt rich in carbonate, CaO, MgO and possibly Na_2O and virtually depleted in SiO_2 and Al_2O_3. The limits of the carbonatite melt field have not been established but it is likely that there is a gradual increase in the SiO_2 content of liquids produced with increasing temperature above the solidus. Strongly SiO_2-undersaturated liquids ranging from basanites to even kimberlites are expected at high pressures and temperatures (Wyllie, 1989; Eggler, 1989). It is concluded that the role of CO_2 alone in igneous petrogenesis becomes significant only at pressures exceeding about 20 kb.

The melting relations of peridotite with mixed volatiles of H_2O and CO_2 are complex and depend strongly on the oxidation state. The phase relations most applicable to planetary mantles (at least for Earth and Venus) are obtained under relatively oxidizing conditions (oxygen fugacities at the QFM buffer or higher) and with H_2O and CO_2 contents relatively low (less than 0.3 and 2.5% respectively; Figure 4). The shape of the solidus is similar to that obtained under H_2O-free conditions, but melting temperatures are significantly depressed (Wallace and Green, 1988). Below 20 kb, the mantle consists of amphibole peridotite and a CO_2-rich vapor. The solidus is coincident with the breakdown of amphibole and
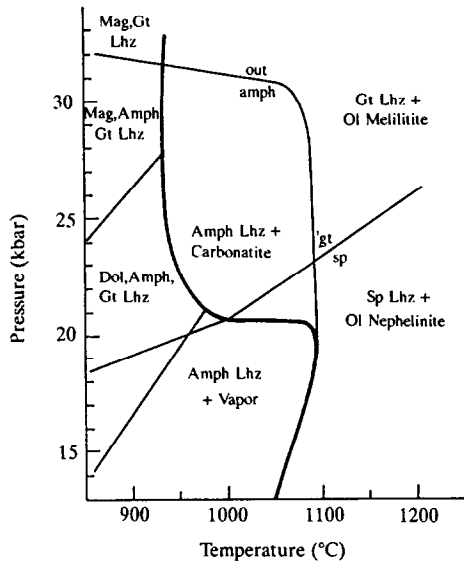

Fig. 4. Phase relationships of peridotite and CO_2/H_2O (amphibole-rich Hawaiian Pyrolite (0.3% H_2O) plus 5% dolomite or 1.4% magnesite). Carbonatite melt field is bounded by the carbonate solidus at 930°C (heavy line) and amphibole breakdown and silicate melt field above 1,080°C (25 kbar) (modified from Wallace and Green, 1988).

the production of a nepheline-normative basaltic melt. Above 21 kb, the mantle consists of an amphibole-dolomite bearing peridotite with no vapor phase – all CO_2 and H_2O is incorporated in carbonated and hydrated minerals. Carbonatite melt is developed at the solidus but gradually evolves to a nepheline-normative basalt (olivine melilitite) as amphibole melts incongruently and carbonate dissolves in increasing amounts of silicate melt. Above 31 kb, magnesite-bearing peridotite and a H_2O-rich fluid phase react at the solidus to produce carbonatite melt and an immiscible silicate melt with olivine melilitite to kimberlite affinities. Higher temperatures may eliminate the carbonatite and produce a CO_2-rich SiO_2-undersaturated silicate melt.

Other scenarios are possible, of course. A relatively oxidized mantle that contains more than about 0.4% H_2O has a CO_2-H_2O fluid even above 20 kb because the H_2O concentration is in excess of that which can be taken up by amphibole. The fluid-saturated solidus is depressed to lower temperatures than in the fluid-absent case (Falloon and Green, 1990). The fluid is H_2O-rich at P >20 kb and CO_2-rich at P < 20 kb. The near-solidus liquids, however, are similar in composition to those developed in the fluid-absent case.

The melting conditions are changed if the mantle is relatively reduced, for example, at oxygen fugacities corresponding to the iron-wustite buffer. The fluid phase may contain CH_4 and H_2O rather than CO_2 and H_2O and depending on the fluid CH_4/H_2O ratio, melting temperatures may be well in excess of those obtained under more oxidizing conditions (Wallace and Green, 1988). It is believed that SiO_2-undersaturated basaltic melts are produced at high pressures in a mantle bathed by CH_4-rich fluids.

It is clear from this review that the nature of fluids in igneous petrogenesis is not only of critical importance but also very complex. Because volcanic and plutonic processes are so sensitive to the composition and abundance of fluids, a more thorough analysis and understanding of Venus volcanic processes, surface-atmosphere interactions, and observations on volcanic landforms will be required to assess their significance further.

4. Melting of Basalts: Tholeiites and Modified Basalts

We now consider the remelting of products initially derived from melting of a peridotitic mantle to produce a basaltic crust. Environments in which this process might occur (Figure 5a) include: (1) the roots of basaltic lower crust below regions of extensive crustal thickening (e.g., possibly the high mountains and orogenic belts of Ishtar Terra; Vorder Bruegge and Head, 1989, Roberts and Head, 1990a,b; Tessera terrain, highly deformed plateaus thought to represent zones of thick crust, Bindschadler and Head, 1990), (2) regions where the crust/lithosphere is delaminated (Bird, 1979; Turcotte, 1989), or (3) regions where crust has been underthrust and perhaps subducted into the mantle (Hsui and Toksöz, 1979; Head, 1990a; Burt and Head, 1990).

(a)   ENVIRONMENTS OF CRUSTAL MELTING
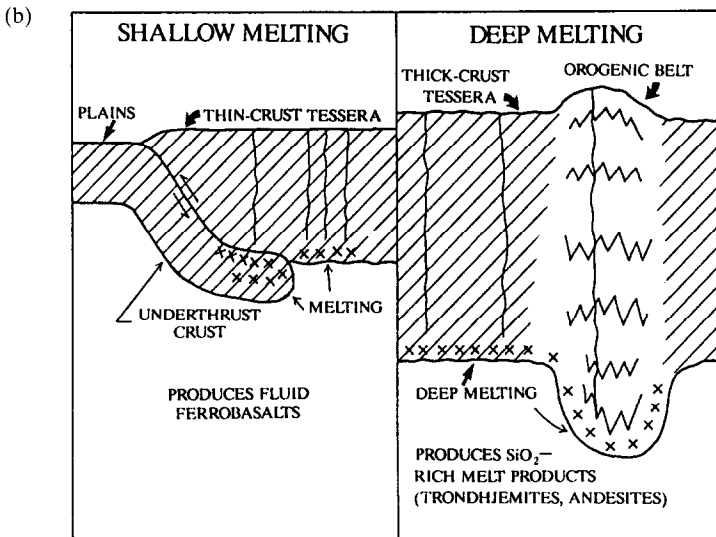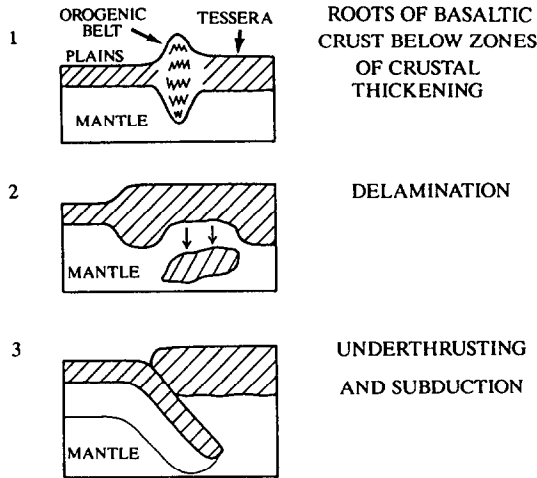   ON VENUS

Fig. 5. Environments for remelting of a basaltic crust on Venus (a) and (b) the environments and products of shallow and deep remelting of crustal material showing shallow melting of an underthrust or subducting slab, and deep melting of a tessera with enhanced crustal thickness and a convergent boundary in the form of a thickened crustal root below an orogenic belt. Interpreted relative thickness of crust in various Venus terrains is shown.

THOLEIITES

Consider first the melting of tholeiite basalt. The simplest case is the melting of basalt under anhydrous conditions and abundant experimental information exists, although some serious gaps in the data set remain. Melting of tholeiite basalt above 15 to 25 kb (about 53–88 km) begins at temperatures in excess of 1200°C and in the eclogite facies (or in the garnet granulite facies, depending on the bulk composition) (Ringwood, 1975) (Figure 6). The melt that coexists with the eclogite assemblage (garnet and clinopyroxene) typically is quartz-normative and strongly enriched in SiO_2 (Johnston, 1986). Small degrees of melting ($< 20\%$) generate trondhjemites (SiO_2 >65%) whereas intermediate degrees of melting (20–50%) yield andesites and basaltic andesites. In contrast, the first melts obtained at lower pressures (10–15 kb; 35–53 km) appear to be relatively SiO_2-poor. For example, the liquids obtained by small degrees of melting at 8 kb (28 km) of high aluminaba-salts are ferrobasalts containing only 42% SiO_2 and more than 20% FeO (Baker and Eggler, 1987). While more experimentation is needed, it is clear that a complete spectrum of magma types can be generated by melting basalt under anhydrous conditions. Large amounts of silicic magmas are generated at high pressures and large amounts of relatively SiO_2-poor, and highly fluid, ferrobasalts are obtained under more modest pressures. Thus, shallow crustal melting occurring in environments such as underthrust basaltic crust and basal melting of tessera crustal blocks less than about 50 km thick (Figure 5b) should result in the produc-
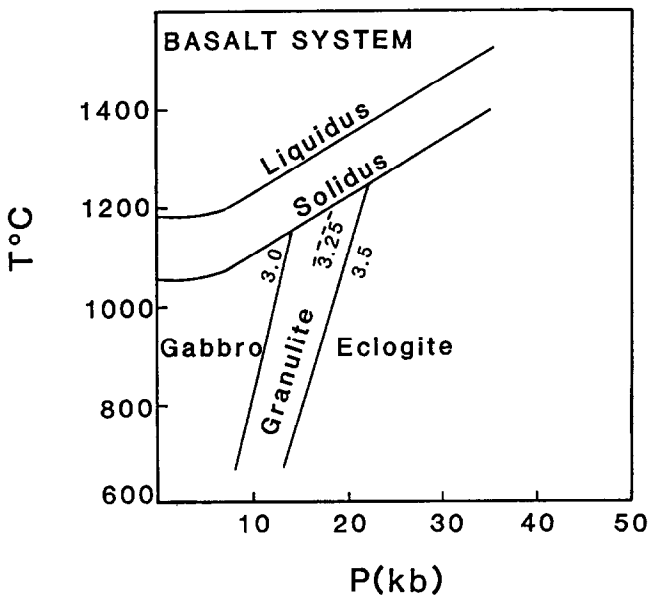

Fig. 6.   Pressure-temperature relationships in the basalt system on Venus. Stability fields and densities of gabbro, granulite, and eclogite are shown.

tion of highly fluid ferrobasalts. Deeper crustal melting, such as that which might be occurring at the base of zones of very thick tessera and the roots of mountain belts of Ishtar Terra, could produce more SiO_2-rich melt products such as trondhjemites, andesites, and basaltic andesites.

CARBONATED/HYDRATED BASALTS

Basalts exposed to the hot atmosphere of Venus may well have undergone carbonatization and perhaps some hydration (Nozette and Lewis, 1982; Warner, 1983; Fegley and Prinn, 1989) and thus a dry tholeiite source rock may not be truly representative of parts of the Venus crust that have been exposed to the atmosphere. Such metamorphosed basalts will be degassed as they are brought to melting temperatures. The question of the role of CO_2-H_2O fluids is not easily addressed, and a number of possibilities must be considered.

The effects of melting of basalt in a pure CO_2 atmosphere are not well constrained. Some features, however, can be predicted from experiments performed on relatively simple systems. The addition of CO_2 will decrease melting temperatures and produce melts that are more silica undersaturated and more alkalic than in the totally anhydrous case (e.g., Eggler, 1978). The effects on the melting temperatures will be most significant at depths where basalt/eclogite is carbonated under near-solidus conditions. This fact is demonstrated very dramatically by the sharp depression of the solidus of carbonated peridotite (Figure 3) (Falloon and Green, 1989). The carbonation reactions in peridotite have olivine as a reactant, however, and are not appropriate as potential carbonation reactions in eclogite which typically is olivine free (unless the basalt is komatiitic). In contrast to peridotite, it appears that basalt/eclogite will remain carbonate free near solidus temperatures to depths of at least 100 km. These depths, which are only approximate estimates, are obtained by comparing the P-T coordinates of carbonation/decarbonation reactions involving pyroxenes (Wyllie, 1989; Eggler et al., 1976) and major eclogite phases to the P-T coordinates of basalt solidi. The conclusion is supported by the observation that the "eclogite" assemblage diopside-pyrope-enstatite remains free of carbonate to 30 kb and 1200–1450°C in the presence of CO_2 vapor (Eggler, 1978). It is concluded then, that the beginning of melting of basaltic crust of Venus would occur in equilibrium with a CO_2 fluid in the upper 100 km. The phase equilibria are unconstrained to greater depths, however.

The solubility of CO_2 in tholeiite basalt is limited to a few weight percent at pressures up to 30 kb (Pan and Holloway, 1990; Mysen et al., 1975). Melting temperatures, therefore, are only modestly depressed, certainly less than 100°C at 20–30 kb. Near-solidus liquids should be similar to those obtained in CO_2-free basalt, except with a tendency to be poorer in SiO_2.

Adding even small amounts of H_2O to the CO_2 fluid causes major changes in melting temperatures and the phase relations because H_2O is selectively enriched in the liquid at pressures below 20 kb (about 70 km depth) (Wyllie, 1980). Melting of tholeiite basalt with H_2O causes the liquidus to be depressed by several hundreds

of degrees and creates near-solidus SiO_2-enriched melts. For example, Spulber and Rutherford (1983) and Helz (1976) obtained tonalitic melts by melting tholeiite basalt at 1–5 kb (4–18 km) water pressure, and temperatures as low as 700°C (see also Beard and Lofgren, 1989). These experiments were done in a pure H_2O atmosphere. Nevertheless, Baker and Eggler (1987) produced dacitic melts by melting tholeiite in a CO_2 atmosphere containing less than 25% H_2O at similar pressures. At higher pressures, CO_2 is more efficiently concentrated in the melt than is H_2O and SiO_2-rich melts should no longer be favored in CO_2-rich environments.

Hydrous basalts will differentiate to SiO_2-rich derivative liquids under most conceivable conditions. If the initial H_2O is relatively high, such derivative liquids will eventually approach saturation with H_2O, and if brought rapidly to the surface in quantities sufficiently great to overcome surface pressures (Head and Wilson, 1986), may cause violent eruptions and ash deposits. Alternatively H_2O-undersaturated lavas, or lavas which have risen slowly enough to gently lose fluids through a foaming stage, may produce thick, short lava flows and obsidian domes.

5. Magmatic Differentiation

So far we have been concerned only with the generation of primary magmas. Processes of magmatic differentiation may produce derivative magmas far removed in composition from their primitive precursors. For example, the steep geothermal gradient that exists in the upper few hundred kilometers on Venus (Figure 7) creates conditions which favor the assimilation of hot basaltic crust by rising primary magmas (Figure 8). Similarly, the large crustal thicknesses predicted to exist beneath regions of high topography (Vorder Bruegge and Head, 1989; Head, 1990b) could lead to ponding of mantle-derived basaltic magmas at the highland crust-mantle interface until appropriate pathways or favorable density contrasts are obtained. Such magmas could undergo extensive differentiation and depending on the depth and volatile content, are capable of yielding: 1) carbonate-rich and/or SiO_2 – undersaturated magmas (conditions at depths greater that 25 kb, or about 88 km, in a CO_2 environment), 2) SiO_2-enriched, tonalitic magmas (depths greater than 15–20 kb, or about 53–70 km, but dry or H_2O-rich), and/or 3) ferrobasalt magmas (shallow, dry or CO_2-rich) (Figure 8).

Modeling of the ascent and eruption of magma on Venus (Head and Wilson, 1986) has shown that the Venus surface environment (T = 650–750 K; P = 40–100 bars) has a potentially important influence on magmatic differentiation processes. These surface conditions mean that the temperature is higher at a given depth on Venus than on Earth, and that much smaller ratios of subsurface to surface pressure occur on Venus than on Earth. Some of the consequences of these factors are: (1) that there will be less cooling of magmas in the final stages of ascent (potentially influencing such factors as the ratio of intrusion to extrusion), and (2) that gas exsolution and consequent vesiculation and pyroclastic eruptions are inhibited (potentially influencing the density structure of the shallow crust and the
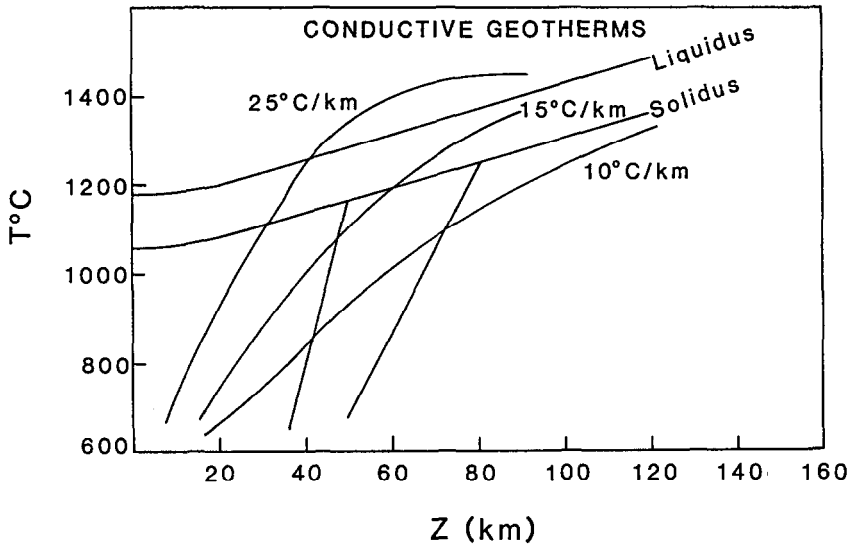
Fig. 7. Temperature and depth relationships for basalts on Venus. Three conductive geotherms are shown. The terrestrial geotherm is approximately 1 0 °C/km and the typical Venus geotherm is thought to lie between 15°C/km and 25°C/km (Solomon and Head, 1990). The pressure-temperature relations of the basalt system (Fig. 6) are also shown, with areas below the solidus corresponding to gabbro (left), granulite (middle), and eclogite (right).
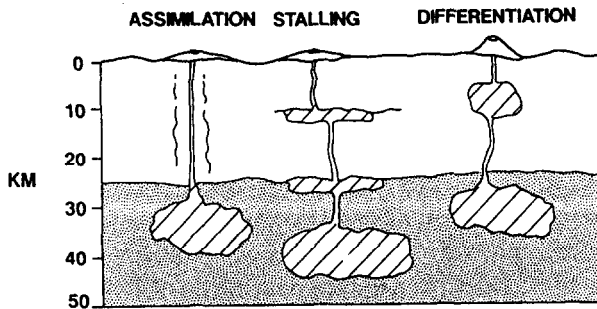

Fig. 8. Shallow magmatic differentiation environments and processes: derivative magmas and the assimilation of hot highland crust by rising primary magmas; stalling and ponding at a compositional or thermal boundary; and differentiation at a neutral buoyancy zone.

location of neutral buoyancy zones - Wilson and Head, 1990). Further analysis of the theory of ascent of magmas, the petrogenetic models discussed here, and the crustal and lithospheric structure will provide a better understanding of the extent and potential environments of magmatic differentiation on Venus.

6. Discussion and Comparison to Terrestrial Magmatic Environments

On the basis of these general considerations, it is clear that no magma type can be arbitrarily excluded from consideration for Venus and it is entirely plausible that petrogenetic processes and magmatic and volcanic activity on Venus could be

very similar to that of Earth in many ways, but having styles of eruptions influenced by environmental conditions (Head and Wilson, 1986). We now examine a range of terrestrial magmatic environments and compare these to expected conditions on Venus.

SPREADING CENTERS

Potential spreading centers on Venus, where relatively degassed mantle would be brought to the melting zone as it is on Earth (Figure 1), should be very similar to crustal spreading environments on Earth. If divergent plate boundary zones and crustal spreading exist on Venus (Head and Crumpler, 1987; Sotin et al., 1989; Crumpler and Head, 1990), then MORB-like volcanics should dominate these zones on Venus as they do on Earth, based solely on the probable geometry of melting. The elevated upper mantle temperatures will likely result in crustal thicknesses on Venus enhanced over the terrestrial case by a factor of 2–3 (Sotin et al., 1989; Hess and Head, 1989, 1990), but the present-day Venus basalts should be more MgO-rich and denser than terrestrial equivalents.

ZONES OF CRUSTAL THICKENING/CONVERGENT MARGINS

The major difference between Earth and Venus may exist at areas comparable to convergent margins, such as island arc systems. On Earth the subducted crust is hydrated (e.g., Hess, 1989), whereas on Venus, it may be anhydrous, or carbonated. The implications of the presence of basalts modified by atmosphere-surface interactions for remelting and the derivative products are linked to the style of crustal formation and evolution (Head, 1990b). For example, if crustal spreading is operating on Venus, then the vast majority of the crustal column is probably formed by intrusion as on Earth, and very little of the total crustal volume interacts with the atmosphere. Thus, when this crust is remelted at zones of crustal thickening or underthrusting, the influence of the volatiles may be minimal. If the crust is formed and evolves by vertical recycling, the effect would also depend on the ratio of intrusion to extrusion. If vertical recycling is largely intrusive, then the atmospheric interaction products are probably of minimal importance. If on the other hand the crust is produced by flow units piled one on top of another until melting of the basalt pile occurs, then each flow will have interacted with the surface and the influence of volatiles will be much more volumetrically significant. Present estimates of crustal thickness (about 10–20 km in the plains; Zuber, 1987; Grimm and Solomon, 1988) combined with the large thickness necessary to produce melting by vertical recycling under most geotherms (Head, 1990b) (Figure 7) suggest that surface reactions will not be a volumetrically significant factor in altering the basic basaltic composition of the crust for vertical recycling processes. This influence may be important locally, however, such as at zones of underthrusting (Head, 1990a) where the thin surface deposits exposed to the atmosphere may be remelted (Figure 5b).

On the basis of the phase equilibria discussions, "island arc" convergent zones

on Venus (assuming that subducted basalts are anhydrous to somewhat carbonated and that a significant percentage of the crustal column has been exposed at the surface) should gradually release small volumes of CO_2 to the overlying mantle wedge. Melting temperatures of peridotite are depressed, but only slightly at pressures less than 19 kb (70 km), and if melts are produced at all, they are SiO_2-undersaturated basalts (i.e., alkali basalts). This should lead to significantly different crustal and morphologic manifestations on Venus than on Earth. On Earth, subduction and dehydration of oceanic basaltic crust releases H_2O into the mantle wedge ultimately producing large volumes of andesite and a range of SiO_2-saturated igneous rocks from basalt to rhyolite. These volcanic products build composite volcanoes (island arcs) on the overriding slab and, combined with the intrusive phases, contribute to the production of crustal material that is less dense than the subducting basalt. This phenomenon produces the well-known and morphologically distinctive terrestrial island arc systems at convergent plate boundaries (Figure 9). Occasionally, because of their lower density and relatively high topography, these island arcs are accreted to the continents to contribute to the growth of continental crust. On Venus, on the other hand, melting of peridotite and CO_2 fluid should produce SiO_2-undersaturated magmas and its differentiates. Such differentiates, which include trachytes and phonolites, have occasionally poured out of large fissures on Earth to form voluminous plateau lavas and even shield volcanoes (Williams, 1982). These SiO_2-enriched volcanics are not characteristic of island arcs, but rather are common in continental rift environments (e.g., Hess, 1989). Island arcs on Venus, therefore, might consist of trachyte and phonolite rather than andesite and dacites as on Earth (Figure 9).

If the subducted crust on Venus is largely anhydrous, there will be no additional melting of the mantle wedge because there are no fluid components to reduce the solidus of peridotite. The solidus of eclogite, however, is about 100–150 °C lower than that of peridotite, so that melting of the basaltic crust could potentially occur
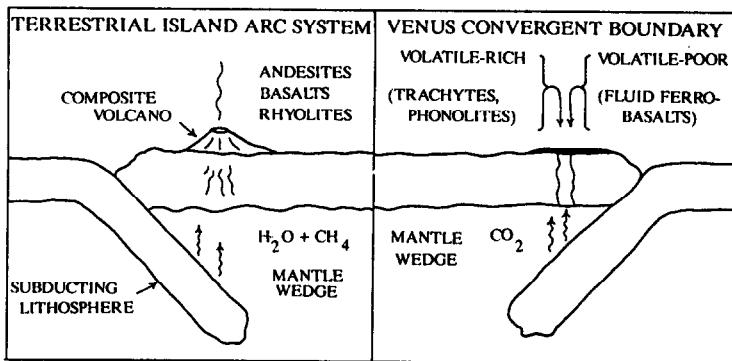

Fig. 9. Generalized characteristics of island arc convergent boundary environments and their Venus analogs.

even if the temperature of the mantle wedge remains above the solidus. The depth at which the eclogite component begins to melt will depend on the subduction rate but thermal equilibration of the slab with the surrounding mantle is rapid under the low descent rates assumed for the Venus lithosphere (Burt and Head, 1989). A slab subducted with a geotherm of 10 °C/km (Figure 7) crosses the basalt-eclogite transition at about 70 km depth, but does not cross the eclogite solidus and remains unmelted. The subduction zone would thus not have a volcanic arc. A slab subducted at geotherms thought to characterize Venus (14–23 °C/km; Solomon and Head, 1990) (Figure 7) crosses the solidus at between 30–60 km depth. The basalt crust associated with the slab would undergo nearly total melting in the case of the steepest geotherm (about 25°C/km), whereas only partial melting would occur in a cooler mantle environment (i.e., a geotherm of about 15° C/km; Figure 7). However, even in the 15 °C/km case, transition to complete eclogite will not occur, and at least partial melting of the basalt crust will dominate. In all of these melting cases, basaltic magmas and their differentiation products would dominate the volcanic arc. Ferrobasalts, rather than SiO_2-rich liquids, would be the most voluminous differentiation product.

It is possible, however, to produce calc-alkaline silicic volcanism under some circumstances even if the subducted crust is anhydrous. If the Venusian mantle is highly reducing, C-H-O fluids are dominated by methane rather than H_2O and CO_2. If the Venus crust is highly oxidized (Pieters *et al.*, 1986), introducing oxygen by subducting oxidized crust into methane-bearing mantle creates H_2O which allows the mantle wedge to melt and could lead to terrestrial-like calc-alkaline volcanism.

If this general characterization of convergent plate boundaries on Venus is true, and if the condition was established early in the history of Venus, it can be predicted that any ancient "continental" or "felsic" crust on Venus was not tholeiitic to tonalitic, as on Earth (Taylor and McLennan, 1985), but may range to alkali basalt and its differentiates if CO_2-dominated melting of subducted crust is dominant. In this case, SiO_2-rich rocks would approach trachyte to phonolite rather than tonalite compositions. On the other hand, if the subducted crust is volatile poor, basalt magmas and their differentiates, SiO_2-poor ferrobasalts, would be the rule and produce regional volcanic plains unlike the typical edifices associated with terrestrial "island arcs". The formation of "continental" or "felsic" crust, in the terrestrial sense, would thus not be favored on Venus. Rather, anomalously thickened plains units or local plateaus of thickened crust perhaps similar to Iceland would be basaltic in composition. The paucity of H_2O on Venus, however, would mean that melting of the basal sections of the basaltic crust would not be enhanced, and therefore, the production of rhyolite lava as is observed on Iceland (Basaltic Volcanism on the Terrestrial Planets, 1981) would not be favored.

MANTLE PLUMES/HOT SPOTS

Mantle that is advected passively beneath a spreading center (Figure 1) under Venusian conditions begins to melt at about 20 kb and first yields picrites (Hess

and Head, 1990) and then SiO_2-saturated boninites at shallower levels (Hess and Head, in preparation). The volcanic expression of off-axis intraplate mantle plumes (Figure 1) however, is quite different and is influenced by the potential temperature of the diapir and the thickness of the mechanical lithosphere. A study of terrestrial intraplate hot spots by Campbell and Griffiths (in press) showed that volcanism is characterized by massive outpourings of flood basalts which begin abruptly and then rapidly decay in intensity within a period of a few million years. Field data and model calculations show that the plume head is large, perhaps 1000 km or more in diameter, and extraordinarily hot near its axis (Figure 10). In these models high temperature picrite melts (MgO $\sim$15%) are derived from the hot interior whereas more voluminous tholeiite basalts are produced from the cooler regions of the plume head. If the Venusian plumes are 100 to 150 °C hotter than their terrestrial counterparts, then the rising plume should begin to melt at about 1600–1700 °C and at about 100 km (Figure 10). The melts entrained in the hot core would resemble basaltic komatites whereas the cooler parts of the plume generate MgO-rich tholeiites. Pressure-release melting continues as the plume head rises until interrupted by the mechanical lithosphere at about 40 km. Approximately 30–40% of the plume core and smaller fractions of the cooler plume head are molten. As the plume head rises, it buoyantly lifts the overlying mantle and places the lithosphere under tension. Stretching, thermal thinning and rifting of the lithosphere results in the abrupt draining of magma from the plume head and voluminous flooding. The large volume of melt, its high temperature and its low viscosity, create large individual flows of potentially great areal extent.

Anomalously hot plumes impinging upon a thin lithosphere do not favor the formation of the SiO_2-undersaturated basalt associations that are so characteristic of intraplate oceanic volcanism on Earth such as on Hawaii or the Azores (Basaltic Volcanism on the Terrestrial Planets, 1981). Several conditions are required to form such basalts from a CO_2-poor peridotite. Alkali basalts are formed by small degrees of melting at pressures exceeding about 15 kb (e.g., Falloon and Green, 1988). Larger degrees of melting at lower pressures produce olivine tholeiites or
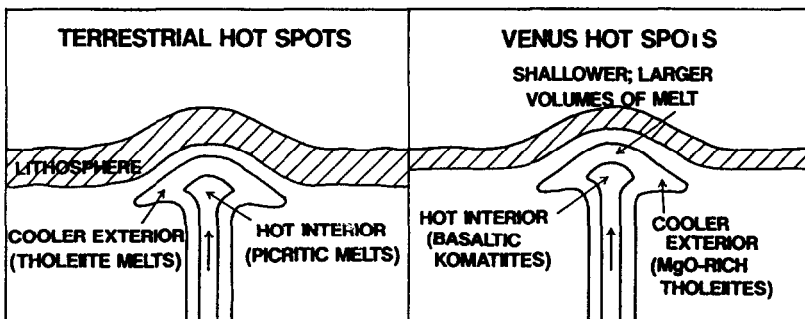

Fig. 10. Comparison of melting relationships for a mantle plume-hot spot on Earth and Venus.

even quartz tholeiites. The addition of CO_2 at low pressures does little to alter these conclusions. Mantle plumes on Venus are probably hotter than those on Earth and, because of the thin lithosphere, rise and continue to melt to higher levels in the mantle. The combination of these parameters results in large volumes of melt that have equilibrated with the mantle at relatively low pressures. These melts range from komatiite to tholeiite in composition. Perhaps only the coolest parts of the plume, which have the lowest melt fraction, will yield SiO_2-undersaturated basalts since the pressure at the base of the lithosphere (10–12 kb) is barely in the stability field of these magmas (Figures 6, 7). Nevertheless, alkali basalts and other SiO_2-undersaturated magmas are more likely to be developed in regions of thickened lithosphere.

Present-day basaltic volcanism on Earth is dominated by olivine tholeiite basalts containing less than about 15% MgO. Mantle adiabats are relatively cool and cross the mantle solidus at depths corresponding to 20 to 30 kb. Much more MgO-rich magmas, komatiites, were common in the Archean but are very rare in the Phanerozoic or even early Precambrian. The reason is that the extraordinary thermal conditions needed to generate komatiites (T > 1900°C and P > 80 kb) require mantle adiabats that are 400–500°C hotter than the average modern mantle at comparable depths (e.g., Nisbet and Walker, 1982; Hess, 1990). This implies the existence of an ancient "magma ocean" (Nisbet and Walker, 1982) or very hot mantle plumes rising from the core-mantle boundary (Campbell *et al.*, 1989; Hess, 1990).

Present-day basaltic volcanism on Venus should be dominated by MgO-rich olivine tholeiite to picritic basalts. If these predictions are correct, and if the thermal and chemical assumptions are accurate, then it is likely that komatiites played a significant role in the early evolution of crust on Venus. This may imply that the early crust was denser and possibly thicker than at present. The greater density, coupled with a more vigorously convecting mantle, could have more readily initiated subduction and crustal recycling.

7. Conclusions

1. *Sources of melts and petrogenetic styles.* On the basis of what is presently known about Venus, three broad sources of melts and petrogenetic styles are likely to occur:

(a) Primary magmas derived from partial melting of a peridotitic mantle (which from peridotite will produce quartz tholeiites to komatiites depending on the P-T conditions, basanites to kimberlites at depths greater than about 70 km if significant CO_2 is involved).

(b) Magmas derived from remelting of anhydrous basaltic crust (which will produce highly fluid ferrobasalts at shallow depths and possibly trondjhemites, andesites, and basaltic andesites at greater depths).

(c) Magmas derived from remelting of carbonated and/or hydrated basaltic crust

(which will produce slightly less SiO_2-rich near-solidus liquids if CO_2 is present and silica-enriched melts at depths greater than about 70 km if H_2O is present).

Thus, a wide range of magma types is possible on Venus depending on which of these environments is dominant and the depth of melting. On the basis of these predictions, this paradigm may help in the interpretation of depth of melting and environments of melting through the recognition of various volcanic deposits in different environments.

2. *Spreading-center environments.* Volcanic activity on Venus should be broadly similar to that on Earth in any divergent plate boundary environment assuming that the main difference between Earth and Venus is higher upper mantle temperatures on Venus. Differences would include greater degrees of melting resulting in picritic basalts and thicker basaltic crust.

3. *Convergent boundary and hot spot environments.* Volcanic activity on Venus could be very different from that on Earth in potential convergent plate boundary environments and intraplate or 'hot spot' environments. The type of volcanism developed above convergent margins will depend on the fluid content of the subducted crust and on the geotherm established in the downwelling regions. Unless H_2O is involved, calcalkaline volcanics, particularly of the andesite-dacite variety, are not anticipated. Basalt and its differentiation products, ferrobasalts under dry conditions and perhaps syenites under CO_2-dominated melting, should characterize the volcanic arc. Hot spot volcanism created on thin lithosphere and from extra-hot mantle plumes should be dominated by picrite basalts, and extensive flood basalt volcanism. SiO_2-undersaturated volcanics should be rare.

4. *Planetary crust formation and evolution.* Planetary crusts are of three principal types (Taylor, 1989): (1) *primary*, which form as a result of accretional heating (e.g., the lunar highland crust); (2) *secondary*, which form as a result of partial melting of planetary mantles (e.g., the lunar maria, and the terrestrial oceanic crust); and (3) *tertiary*, which form by the reprocessing of secondary crusts (e.g., continental crust on Earth). On the basis of our analysis, we conclude that primary and secondary crusts on Venus should be similar to that of the Earth, but that the tertiary crust on Venus is likely to be different than that on Earth because of the lack of water and the production of ferrobasalts rather than a range of silica-rich magmas.

5. *Venera lander analyses and mechanisms of crustal formation.* Both hot spot and crustal spreading petrogenetic environments are capable of producing the range of compositions measured by the Venera landers, and could be candidates for formation of the crust on Venus. What is needed to assess their relative importance is a better understanding of the implications of estimates of present crustal thickness and volumes and how this links to melting and recycling (see discussion in Head, 1990b), and to the volume of crust expected throughout the thermal evolution of the planet.

6. *Relationship to Earth history.* The Archean on Earth was distinguished by the widespread occurrence of komatiites within the basal volcanic sections in

greenstone belts (Nisbet, 1987). The high MgO contents, the near-chondritic inter-element ratios of certain elements, and the extraordinarily high, one-atmosphere liquidus temperatures require that most komatiites are products of very large degrees of melting in the Archean. Some have even argued that komatiites orig-inate from a magma ocean (Nisbet and Walker, 1982; Hess, 1990) or from ultra-hot diapirs rising from deep within the mantle (Campbell and Griffiths, in press; Hess, 1990). Although komatiites make up a small fraction of the volcanic piles in greenstone belts, this may reflect the difficulty of moving high-density liquids through the less dense "continental" crust. It is quite possible that komatiite volcanism was more pervasive on Earth in its early history but that most were extruded in oceanic crustal environments and ultimately subducted into the mantle. Venus apparently does not have an extensive "continental" or "felsic" crust, and since Venus is almost as massive as Earth, and was as hot or even hotter, it follows that komatiite volcanism was and perhaps still is extensive. Some of the long and voluminous lava flows observed on the surface (Head and Wilson, 1986; Wilson *et al.*, 1990) may represent such deposits. It is possible, therefore, that the present surface and volcanic processes on Venus may bear some resemblance to the pre-Archean on Earth, before nuclei of continental "felsic" masses were firmly established.

7. *Future studies.* Additional sample analysis missions are needed to distinguish between various hypotheses for the nature of melting, for the remelting of crustal material and assessment of the role of volatiles, and for the extent of magmatic differentiation during magma ascent and eruption. Specifically, surface chemical analyses at possible spreading centers (Western Aphrodite Terra), hot spots (central Beta Regio), and zones of convergence, crustal thickening, and possible underthrusting, subduction, and remelting (Western Ishtar Terra) would help to assess these factors.

Acknowledgments

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