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CHEMICAL DIFFERENTIATION OF A CONVECTING PLANETARY INTERIOR: CONSEQUENCES FOR A ONE-PLATE PLANET SUCH AS VENUS. E. M. Parmentier and P. C. Hess, Department of Geological Sciences, Brown University, Providence RI 02912, USA.



Fig. 1.

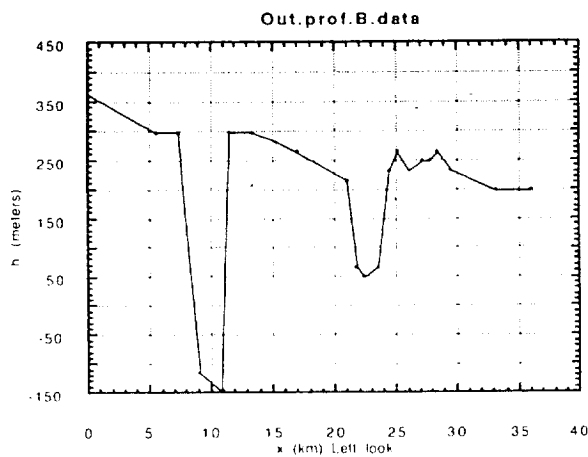


Fig. 2.

on the plateau and the cliff walls in the two images (Fig. 1). The layover "shifts" the features closer to the apparent edge of the wall relative to the oppositely illuminated image. Figure 1 also shows one "single-point" depth measurement, illustrating the large parallax displacement between the left- and right-looking data in areas of moderate relief. At the bottom of each scene is a sample profile trace across the trough and part of the channel connecting the collapsed source with the trough. Figure 2 is a plot of this data using the cycle 1 image sample values for placement on the x axis. Vertical height values are relative values, and are a function of the parallax displacement of features relative to the projection of the two FMIDRs.

References: [1] Moore H. J. et al., this volume. [2] Parker T. J. et al. (1991) *LPSC XXII*, 1035-1036. [3] Baker V. R. et al. (1992) *JGR*, in press.

Partial melting of the interior of a planet to generate its crust must inevitably leave behind compositionally buoyant residual mantle. This basalt-depleted mantle is chemically less dense than undepleted mantle due to its reduced Fe/Mg and dense Al-bearing minerals such as garnet. The chemical density difference is substantial: for 20% melt extraction the density decrease is as large as that due to a 500°C temperature increase. The melting temperature of this depleted residual mantle is also increased. Deep mantle circulation driven by cold strong sinking lithosphere associated with plate tectonics may mix the depleted mantle back into the mantle. In the absence of plate tectonics, less mixing will allow a buoyant depleted layer to collect at the top of the mantle [1,2].

Chemically depleted mantle forming a buoyant, refractory layer at the top of the mantle can have important implications for the evolution of the interior and surface. On Venus, the large apparent depths of compensation for surface topographic features [3] might be explained if surface topography were supported by variations in the thickness of a 100-200-km thick chemically buoyant mantle layer or by partial melting in the mantle at the base of such a layer. Long volcanic flows seen on the surface [4] may be explained by deep melting that generates low-viscosity MgO-rich magmas. The presence of a shallow refractory mantle layer may also explain the lack of volcanism associated with rifting [5]. As the depleted layer thickens and cools, it becomes denser than the convecting interior and the portion of it that is hot enough to flow can mix with the convecting mantle. Time dependence of the thickness of a depleted layer may create episodic resurfacing events as needed to explain the observed distribution of impact craters on the venusian surface [6].

We consider a planetary structure like that shown in Fig. 1 consisting of a crust, depleted mantle layer, and a thermally and chemically well-mixed convecting mantle. The thermal evolution

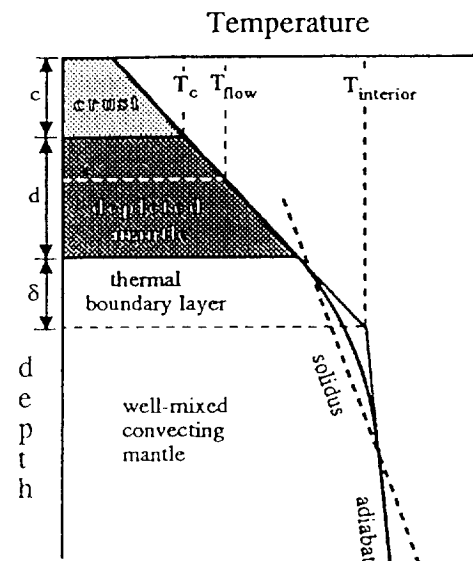


Fig. 1. Planetary structure considered in models of thermal/chemical internal evolution. These models examine the role of a chemically buoyant basalt-depleted mantle layer and crustal recycling on planetary evolution.

of the convecting spherical planetary interior is calculated using energy conservation: the time rate of change of thermal energy in the interior is equated to the difference in the rate of radioactive heat production and the rate of heat transfer across the thermal boundary layer (cf. [7]). Heat transfer across the thermal boundary layer is parameterized using a standard Nusselt number-Rayleigh number relationship $Nu = (Ra/Ra_{crit})^{1/3}$. The radioactive heat production decreases with time corresponding to decay times for the U, Th, and K. The planetary interior cools by the advection of hot mantle at temperature $T_{interior}$ into the thermal boundary layer where it cools conductively. The crust and depleted mantle layers do not convect in our model so that a linear conductive equilibrium temperature distribution is assumed as shown in Fig. 1.

The rate of melt production is calculated as the product of the volume flux of mantle into the thermal boundary layer and the degree of melting that this mantle undergoes. The volume flux of mantle into the thermal boundary layer is simply the heat flux divided by amount of heat lost in cooling mantle to the average temperature in the thermal boundary layer. The degree of melting is calculated as the temperature difference above the solidus, approximately $(T_{interior} - T_m)/2$ where T_m is the melting temperature (see Fig. 1), divided by the latent heat of melting. A maximum degree of melting, 25% in the results described below, is prescribed corresponding to the maximum amount of basaltic melt that the mantle can initially generate. As the crust thickens, the pressure at the base of the crust becomes high enough and the temperature remains low enough for basalt to transform to eclogite. We assume that basalt does not transform to eclogite below a prescribed kinetic blocking temperature T_{ge} , which we vary over the range $700^\circ - 900^\circ C$. Basalt that transforms to eclogite is assumed to sink into the convecting mantle and to mix instantaneously with it, both thermally and chemically. Heat-producing elements fractionated into the crust are thus returned to the mantle. It is not yet clear based on surface images alone whether or how crust is recycled on Venus, but the large amount of crust predicted by our models suggests that recycling must occur.

The stability of the depleted mantle layer depends on its density relative to that of the convecting interior. We take $\Delta\rho/\rho_0 = \alpha(T_{interior} - T_{depleted}) - \beta X$ where $T_{depleted}$ is the average temperature in the part of the depleted layer that can flow ($T > T_{flow}$), X is the average degree of melting, and the coefficient β accounts for the chemical density variations described above. We take $T_{flow} = 1100^\circ C$. The stability conditions for the depleted layer are then simply that if $\Delta\rho > 0$, the depleted mantle instantaneously mixes with the convecting mantle, while if $\Delta\rho < 0$, the layer is stable and does not mix at all with the convecting mantle. The latter condition, in particular, is oversimplified because it does not account for the possibility that viscous stresses in the convecting mantle can entrain buoyant material if its $\Delta\rho$ is not sufficiently large [8].

With a prescribed initial temperature and heat generation rate, the time evolution of this model is calculated numerically. One result of such a calculation, in Fig. 2, shows temperatures, crust and depleted layer thicknesses, and degree of melting as functions of time. The interior temperature decreases only slightly in this particular case because the rate of heat production is nearly sufficient to balance secular cooling. The crust initially thickens but quickly reaches a thickness of about 60 km at which crustal recycling begins. The depleted mantle layer at first thickens because the layer is below the temperature T_{flow} . However, as the crust thickens, the temperature of the depleted layer becomes greater than T_{flow} . In addition, the temperature difference between the interior and the depleted layer is large enough that the depleted layer is

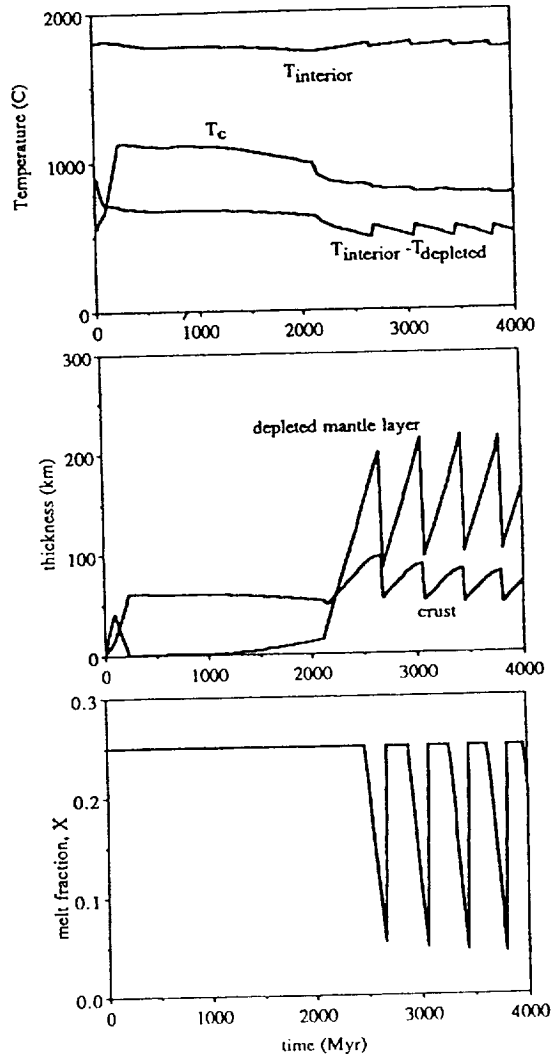


Fig. 2. Planetary evolution including the effects of crustal recycling and the dynamics of a chemically buoyant basalt-depleted mantle layer. Temperatures (top), thicknesses of crust and depleted mantle layers (middle), and melt fraction (bottom) illustrate the evolution that is discussed in the text. In this example the initial interior temperature is $1800^\circ C$, the rate of heat production is 0.5 of a steady-state Earth, $T_{flow} = 1100^\circ C$, and $T_{ge} = 900^\circ C$.

denser than the interior. Thus for a period of the evolution a depleted layer does not accumulate. However, as the interior cools, a growing thickness of depleted material is cooler than T_{flow} . As $T_{interior} - T_{depleted}$ decreases, negative thermal buoyancy becomes less than the positive compositional buoyancy. The depleted layer then thickens rapidly (at slightly more than 2000 m.y. in Fig. 2). As the depleted layer thickens, the thermal boundary layer and region of melting move deeper into the interior. As the melting temperature at the base of the thermal boundary layer increases, due to this increase in the depth of melting, the amount of melting decreases as shown in Fig. 2. The average compositional buoyancy of the depleted layer thus decreases, until the thermal buoyancy results in a net negative buoyancy. The layer then mixes into the convecting interior, the thermal boundary layer and region of melting rise to lower pressure, larger amounts of melting resume, and the depleted layer thickens once more. This process repeats itself at very regular intervals of about 400–450 m.y. This time interval is remarkably similar to the inferred resurfacing age of Venus.

We have explored behavior of this model for a range of the parameters. The time-averaged thickness of the depleted layer is controlled in part by T_{flow} : decreasing T_{flow} to 1000°C reduces this thickness by about 50%. Decreasing T_{ge} to 700°C reduces the time-average thickness of the crust by about 20 km. However, the cyclical variation in depleted layer thickness, along with the accompanying fluctuation in crustal thickness, is a robust feature of the models. Varying the initial temperature by $\pm 200^\circ\text{C}$ and radioactive heating, expressed as a fraction of that required to explain all the Earth's present-day heatflow by radioactivity, by a factor of 2 influences the onset time of this behavior, but the period of the variation remains on the order of 300–500 m.y. for the complete range of conditions considered.

The parameterized convection model assumes that instabilities are globally synchronous. If the instability described by this model is global in scale, it may take the form of episodic plate spreading and subduction. But studies of impact crater densities on Venus [6] and the distribution of volcanic features [4] suggest that resurfacing may occur in patches rather than globally. Localized volcanic resurfacing on Venus may be a consequence of local instability of the lithosphere or alternatively may mean that large, exceptionally hot plumes penetrate even thick, buoyant lithosphere. The Archean greenstone belts on Earth, which are flooded by high MgO volcanics, require similar mechanisms of formation. Komatiites, for example, require potential temperatures of at least 1800°C [9,10] and mean depths of melt segregation of 160–330 km [11], yet average mantle temperatures in the Archean are thought to be only 100°–150°C higher than present [12]. These contradictions are best explained by a model in which komatiites form only in plumes, whereas more typical terrestrial basalts form at spreading centers. The chemical differentiation of Venus described in this study almost demands that komatiite-to-picrite volcanics form the dominant portion of the venusian crust.

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VENUS STEEP-SIDED DOMES: RELATIONSHIPS BETWEEN GEOLOGICAL ASSOCIATIONS AND POSSIBLE PETROGENETIC MODELS. B. Pavri and J. W. Head III, Department of Geological Sciences, Brown University, Providence RI 02912, USA.

Introduction: Venus domes are characterized by steep sides, a circular shape, and a relatively flat summit area. In addition, they are orders of magnitude larger in volume and have a lower height/diameter ratio than terrestrial silicic lava domes [1]. The morphology of the domes is consistent with formation by lava with a high apparent viscosity [2]. Twenty percent of the domes are located in or near tessera (highly deformed highlands), while most others (62%) are located in and near coronae (circular deformational features thought to represent local mantle upwelling). These geological associations provide evidence for mechanisms of petrogen-

esis and several of these models are found to be plausible: remelting of basaltic or evolved crust, differentiation of basaltic melts, and volatile enhancement and eruption of basaltic foams.

Development of Models: Hess and Head have shown that the full range of magma compositions existing on the Earth is plausible under various environmental conditions on Venus [11]. Most of the Venera and Vega lander compositional data are consistent with tholeiitic basalt [3–6]; however, evidence for evolved magmas was provided by Venera 8 data consistent with a quartz monzonite composition [7]. Pieters et al. have examined the color of the Venus surface from Venera lander images and interpret the surface there to be oxidized [8].

Preliminary modeling of dome growth has provided some interpretations of lava rheology. Viscosity values obtained from these models range from 10^{14} – 10^{17} Pa·s [9], and the yield strength has been calculated to be between 10^4 and 10^6 Pa [1], consistent with terrestrial silicic rocks. The apparent high viscosity of the dome lavas suggests that the domes have a silicic composition or must augment their viscosity with increased viscularity or crystal content.

Petrogenetic Models: Sixty-two percent of the Venus domes are associated with coronae, circular features that have been proposed as sites of mantle upwelling, and 20% of the domes are located near tessera, relatively high areas of complex deformed terrain. We have investigated several models that are consistent with these geologic associations. The first case involves the differentiation of basalt in a magma reservoir in the crust, perhaps produced by partial melting within a mantle plume. The second case is melting at the base of thickened basaltic crust, and the final case is volatile exsolution and enhancement within a basaltic magma reservoir. The association of domes with tessera might be explained by crustal remelting, while the association with coronae may be consistent with chemical differentiation of a magma reservoir or the exsolution and concentration of volatiles in the reservoir before eruption.

Chemical Differentiation: High-silica magmas can be produced under reducing or oxidizing conditions, and regardless of whether the crust is wet or dry. If water is present, crystal fractionation of a basaltic magma will produce intermediate to silicic magmas. Differentiation of dry oxidized basalt in a magma reservoir can also produce silica-rich magma, as well as a suite of intermediate composition magmas [10]. The production of immiscible silica-rich melts and ferrobasalts occurs under reducing conditions, but no intermediate magma is produced [10].

Crustal Remelting: The melting of dry tholeiite basalt at pressures of 15–25 kbar or above will result in SiO_2 -rich magmas for <20% partial melting [11]. Depths of 53–88 km are necessary so that melting occurs in the eclogite facies where garnet is present as a low-silica phase in the residue. For higher degrees of melting, andesites or basaltic andesites will form. The presence of water would allow the formation of high silica melts at shallower depths since amphibole could replace garnet as a low silica residue. An excess of water would also reduce the viscosity of a high silica melt, making it easier to transport. The volume of crustal melting required to produce one dome would be reduced considerably if tessera represents evolved crust, as proposed by Nikolayeva [7].

Volatile Exsolution/Basalt Foam: Volatile enhancement represents an alternative mechanism for increasing magma viscosity. In this model, magma viscosity is increased by two mechanisms. First, as more vesicles form in the magma, the bubbles have difficulty moving past one another and second, the liquid has difficulty moving along the thin interbubble walls as the vesicles