CHEMICAL DIFFERENTIATION OF A CONVECTING PLANETARY INTERIOR: CONSEQUENCES FOR A ONE PLATE PLANET SUCH AS VENUS

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Abstract. Partial melting to generate the crust of a planet creates compositionally buoyant residual mantle. In the absence of mantle flow associated with plate tectonics, this buoyant, refractory layer may collect at the top of the mantle with important implications for the evolution of the interior and surface. In this study models of the thermal and chemical evolution of a planetary interior demonstrate the possible consequences of a chemically buoyant depleted mantle layer. As the depleted layer thickens the melting temperature at the top of the underlying convecting mantle also increases and the density difference across the thermal boundary layer, $T_{\text{interior}} - T_{\text{a}}$ (see Figure 1), and $h$ is the thickness of the convecting mantle, corresponding to the overall mantle depth (3000 km) less the thickness of crust and depleted mantle present at any stage of the evolution. The viscosity for thermally activated solid state creep is temperature dependent. However, for simplicity, we assume a constant mantle viscosity and consider values in the range $\mu = 5 \times 10^{20}$ to $10^{22}$ Pa-s. Future studies will need to examine the possible effect of temperature dependent viscosity. We adopt a 500°C surface temperature for Venus. Other physical parameters are taken to have values that would be reasonable for the mantle of Venus (cf. Turcotte and Schubert, 1982). The radioactive heat production decreases with time due to the decay of the U, Th and K. This is represented as a simple exponential decay with a half-life of $3 \times 10^9$ years. The magnitude of the heat production rate is expressed in terms of $H_{2}$ (2.5 x 10^{-11} W/kg), the value that would be required to provide all of the earth's present heat flux by radioactivity (Turcotte and Schubert, 1982).

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.

$$C_{\text{p}} \left( T_{\text{interior}} - T_{\text{m}} \right)/2L$$

where $T_{\text{m}}$ is the melting temperature, $(T_{\text{interior}} - T_{\text{m}})/2$ is the degree of melting that this mantle undergoes, and 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 $C_{\text{p}}(T_{\text{interior}} - T_{\text{a}})/C_{\text{S}}$, where $T_{\text{m}}$ is the melting temperature, $C_{\text{S}}$ is the specific heat of a solid, and $C_{\text{p}}$ is the pressure dependent solidus that we used is $T_{\text{m}}(C) = 1100 + 1.75 z$ where $z$ is the depth in km (Takahashi and Aubele, 1986).
**Fig. 1.** Planetary structure considered in models of thermal and chemical internal evolution. These models examine the role of a chemically buoyant basalt-depleted mantle layer and crustal recycling on planetary evolution.

With a prescribed initial interior temperature and heat generation rate, the time evolution of this model is calculated numerically. This includes the thermal evolution of the convecting interior along with the generation and recycling of crust and the creation and instability of the depleted mantle layer. The time step for the integration in the following results is 1 Myr. Accuracy was checked by reducing the time step by a factor of two and observing no significant change in the numerical solutions.

**Results**

We have explored the behavior of this model over a range of values for the initial interior temperature, the rate of heat production, $T_{\text{flow}}$, and $T_{\text{ge}}$. We consider initial temperatures in the range of 1800 to 2600 °C and rates of heat production from 0.25 $H_e$ to 0.75 $H_e$. If approximately 15% of the earth's heatflow is derived from continental crust (cf. Turcotte and Schubert, 1982) and about 20% of the remaining mantle heat flux reflects secular cooling, the value 0.7 $H_e$ may be a good estimate for the earth's mantle.

Figure 2 shows model results as functions of time for a 2200°C initial interior temperature, heat production rate of 0.5 $H_e$, mantle viscosity $\mu=10^{21}$ Pa-s, $T_{\text{flow}}=1100$°C, and $T_{\text{ge}}=800$°C. The interior temperature at first decreases with time reflecting both secular cooling and decreasing radioactive heating. The crust initially thickens but quickly reaches a thickness of about 70 km at which crustal recycling begins.

As the crust thickens the pressure at the base of the crust becomes high enough while the temperature remains low enough for basalt to transform to dense eclogite. Gabbro transforms to eclogite over a pressure range (Ringwood, 1975), but we assume for simplicity that the transformation occurs at a pressure $P$ (MPa) = 2 $(T - 100)$ where $T$ is in °C. We also assume that basalt does not transform to eclogite below a prescribed kinetic blocking temperature $T_{\text{ge}}$, which we vary over the range 600-800°C. Basalt which transforms to eclogite is assumed to sink through the depleted layer and mix instantaneously with the convecting mantle, 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 how crust is recycled on Venus, but the large amount of crustal production 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 $\Delta p/\rho_0 = \alpha(T_{\text{interior}} - T_{\text{depleted}}) - \beta X$ where $T_{\text{depleted}}=(T_d + T_{\text{flow}})/2$, the average temperature in the part of the depleted layer that can flow ($T > T_{\text{flow}}$, see Figure 1), $X$ is the average degree of melting, $\alpha$ is the coefficient of thermal expansion (3 x $10^{-5}$°C), and the coefficient $\beta$ (7.5 x $10^{-3}$) accounts for the chemical density variations described above. The stability conditions for the depleted layer are then simply that if $\Delta p > 0$ the depleted mantle instantaneously mixes with the convecting mantle while if $\Delta p < 0$ 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 with a small $\Delta p$. Laboratory experiments (Olson and Kincaid, 1990) suggest, however, that entrainment does not occur if $|\Delta p| \geq \alpha \Delta T$ where $\Delta T$ is the temperature difference across the thermal boundary layer.

**Fig. 2.** Planetary evolution including the effects of crustal recycling and the dynamics of a chemically buoyant basalt-depleted mantle layer. Temperatures (a), thicknesses of crust and depleted mantle layers (b), melt fraction and crustal production rate (c), and surface thermal gradient (d) illustrate the evolution that is discussed in the text. In this example the initial $T_{\text{interior}}$ is 2200°C, radioactive heat production rate is 0.5 $H_e$ (see text for definition), the mantle viscosity $\mu=10^{21}$ Pa-s, $T_{\text{ge}}=800$°C, and $T_{\text{flow}}=1100$°C.
The depleted mantle layer at first thickens because the layer is below the temperature $T_{\text{flow}}$. However, as the crust thickens, the temperature of the depleted layer becomes greater than $T_{\text{flow}}$. And, the temperature difference between the interior and the depleted layer is large enough that the depleted layer is 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 $T_{\text{flow}}$, and the depleted layer gradually accumulates (beginning at about 1600 Myr). As $T_{\text{interior}} - T_{\text{depleted}}$ decreases, the negative thermal buoyancy of the depleted layer becomes less than its positive compositional buoyancy. The depleted layer then thickens rapidly (at about 3600 Myr). As the depleted layer thickens, the thermal boundary layer and region of melting move deeper into the interior. The melting temperature at the base of the thermal boundary layer increases, due to this increase in the depth of melting; and, therefore, the amount of melting and the crustal production rate decrease. 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 sequence repeats at very regular intervals of about 400-450 Myr.

Figure 3 shows the evolution of crust and depleted mantle layer thicknesses for a range of heat production. (a) heat production rate = 0.25 $H_{\text{C}}$, (b) heat production rate = 0.75 $H_{\text{C}}$. In each case all other parameters have the standard values given in Figure 2. These results should be compared with those in Figure 2(b). Note, that the time scale in the bottom diagram is twice the length of that in the top diagram and in Figure 2.

Models with different initial interior temperatures, but with a heat production rate and other parameters the same as in Figure 2, show that the initial interior temperature, like the rate of heat production, controls the time in the evolution at which instability begins. For initial interior temperatures of 1800°C and 2600°C the onset time is about 3 Gyr and 4 Gyr, respectively. Higher initial temperatures require more cooling and so longer for instability to occur. Except for the difference in onset time, the evolution looks very similar to that shown in Figure 2. The thicknesses of the crust and depleted layer and the frequency of instability are relatively independent of initial interior temperature.

We have also considered different values of $\mu$, $T_{\text{flow}}$, and $T_{\text{ge}}$ with all other parameters except the one varied the same as in Figure 2. Decreasing the viscosity from $10^{21}$ Pa-s to $5 \times 10^{20}$ Pa-s increases the cooling rate and causes instability of the depleted layer earlier in the evolution (about 2 Gyr), similar to the effect of decreasing the heating rate or the initial interior temperature as described above. The average thickness of the depleted layer is sensitive to $T_{\text{flow}}$. Decreasing $T_{\text{flow}}$ from 1100°C to 1000°C reduces the thickness of depleted layer that is cooler than $T_{\text{flow}}$. The total layer thickness decreases by about 50%. The period of the depleted layer instability is also reduced to about 350 Myr. Decreasing $T_{\text{ge}}$ from 800°C to 600°C reduces the average thickness of the crust by about 30 km late in the evolution. This thinner crust results in lower temperatures in the depleted layer and a thicker layer since more depleted mantle is cooler than $T_{\text{flow}}$. The fluctuation in crustal thickness is much smaller.
Discussion

Our parameterized convection model does not address the spatial scale of depleted mantle layer instability or of crustal recycling. Instabilities of the depleted mantle layer are either global in scale or globally synchronous at smaller scales. Although the distribution of volcanic features [Head, et al, 1992a] and the number of volcanically modified impact craters appear to be nonrandom, the global population of impact craters cannot be distinguished from a completely spatially random distribution [Phillips, et al, 1992]. This is consistent with global resurfacing [Phillips, et al, 1992; Schaber, et al., 1992].

Our model implicitly assumes that crustal recycling and depleted mantle layer instability occur independently by different mechanisms. Crustal recycling occurs as dense eclogite forming at the base of the crust sinks through the depleted mantle layer to mix with the deeper convecting mantle. Alternatively, the sinking eclogite could disrupt the depleted mantle layer. If this type of whole-lithosphere instability is global in scale, it may take the form of episodic plate tectonics [Turcotte, 1992]. Two and three-dimensional models of mantle dynamics and their application to interpret geologic features on Venus will be needed to resolve the scale at which instabilities develop and the possible interaction between instabilities arising in the crust and deple ted mantle layer. The origin of many surface features on Venus remains debatable; however, no evidence of plate tectonic features has been identified [e.g. Solomon, et al., 1992].

In addition to the possible implications mentioned previously, a thick depleted mantle layer has interesting consequences for magma genesis and for the thickness of the elastic lithosphere. Depths of melting in our model exceeds the combined thickness of the crust and the depleted lithosphere cooler than Tflow. The composition of melts generated at these great depths would range from picrite to komatite [Hess and Head, 1990]. Such magmas occur primarily in the Archean greenstone belts on earth. The large thickness of basaltic crust predicted by these models would generate crustal roots well within the garnet granulate and eclogite facies. Partial melting of these roots would generate SiO2-rich, highly viscous magmas, even in the absence of H2O or other fluids [Hess and Head, 1990]. The existence of these siliceous magmas may be expressed in the steep-side, pancake domes observed on Venus [Pavri et al, 1992].

Recent studies of topography adjacent to surface loads have inferred an elastic lithosphere up to 40 km thick [Sandwell and Schubert, 1992]. If rock remains elastic to temperatures of 750°C, as in the terrestrial oceanic lithosphere, this large elastic thickness implies a thermal gradient as low as 6°C/km. In our model, a buoyant depleted mantle layer through which heat is transferred by conduction results in thermal gradients approaching values this low when the depleted layer is thick (Figure 2d).

Perhaps the most interesting characteristic of our model is the cyclical variation in depleted layer thickness that it predicts. This, along with the accompanying fluctuation in crustal production rate, is a robust feature of the models. Varying the initial temperature by ±400°C and radioactive heating by a factor of three influences the onset time of this behavior, but the initial period of the variation remains on the order of 300-500 Myr for the complete range of conditions considered. Taken at face value, this time interval is remarkably similar to the inferred resurfacing age of Venus.

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References


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