A COMBINED GEOCHEMICAL AND GEODYNAMIC PERSPECTIVE ON THE THERMOCHEMICAL EVOLUTION OF MARS.
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Introduction: Understanding the thermal-chemical evolution of the terrestrial planets is critical to obtain information about the current and past states, e.g., habitability. In this particular case, Mars is an ideal candidate to study this process: Mars' surface has largely been preserved from antiquity [1-3], there has been ongoing igneous activity into the present [1-5], and the planet has been the subject of numerous missions. From these missions in conjunction with the study of martian meteorites, there is sufficient data to better constrain the thermochemical history of Mars.

Previous thermal history inferences come from two approaches: Geochemical and geodynamic. A previous geochemical approach used the martian meteorites to determine the mantle potential temperature (T_P) through time [6]. Overall, they found a general trend of cooling through time, with the notable exception of the olivine-phyric shergottites, which were postulated to be sourced from a mantle plume. Some previous geodynamic approaches used various models to place constraints on core size, surface heat flow, etc. [7-9]. Independently, geodynamic models are often nonunique, and the geochemical models may have either high degree of uncertainty or lack specific thermal evolutionary context. Here we seek to combine both geochemical and geodynamic models in an iterative, boot-strap approach to investigate the thermal evolution of Mars.

Methods: Our initial geochemical and geodynamic models are independent. Each approach is inherently coupled via a common output, for our purposes here: thermal profiles of the crust and mantle (areotherms). These techniques are complementary as they use much of the same input data including: the concentrations of the Heat Producing Elements (HPE: K, Th, U) [10,11], thermal conductivity (k), density, and crustal thickness. It should be noted that both approaches produce *average* areotherms for the planet, which does not preclude hotter (e.g., plume) or colder regions of the planet.

Geochemical models. Here we used an iterative, robust approach to calculate areotherms through the conductive crust and lithosphere [12-14]. This model also requires *a priori* knowledge of the surface heat flow (q_0), so we used representative values of 50 mW/m^3 for 4 Ga [7] and 25 mW/m^3 for today [9], and for simplicity we assumed a linear decrease with time. This approach has the advantage of variable vertical resolution, e.g., we can distribute the HPEs homogeneously throughout the crust and lithosphere or in thinner discrete layers. We started with the assumption that the initial primordial crust was 20 km thick and enriched in HPEs [15]. We then assumed either linear growth to 50 km

[16], or growth via mantle melting where the melt is extracted from the mantle and added to the crust proportionally to the calculated melt fraction (F). The melting approach used the mantle adiabats [6,17] and the mantle solidus [18] to determine F and pressure (P) of melting. We used least squares regression to parameterize previous experimental data [19,20], which measured melt composition as a function of P and T and calculated the equivalent F. This allowed us to calculate melt composition as a function of both P and F. In turn, this also allowed us to vary the concentrations of HPEs in each layer which was carried forward into the next timestep. Additionally, after the areotherm models were complete, we used the T and compositional outputs as inputs to Perple_X [21] to calculate the density and seismic velocities (V_P and V_S) of the coupled crust and lithosphere.

Geodynamic models. We used CitcomS [22] to calculate volumetrically averaged mantle areotherms for boundary constraints in the areotherm modeling. CitcomS is a robust and well benchmarked code that utilizes fully spherical 3D and dynamic domains of the whole mantle, as opposed to the 1D models required by the geochemical calculations. These profiles employed a constant boundary temperature at the surface ($T = $ -53 ºC) and at the CMB ($T = $ 1547 ºC); with an adiabatic gradient of 0.18 ºC/km [17]) and a basally defined Rayleigh number of 3x10^6. We considered variable internal heating rates (Q), a proxy for time and chemical depletion, and core fractions (f) ranging from 0.54 (S-bearing) to 0.4 (S-poor) of the total planet radius. From these profiles, we determined average q_0 and F as a function of time.

Geochemical + Geodynamic output. To begin to work toward self-consistency, we used the CitcomS q_0 outputs and age estimation as inputs to the geochemical areotherm calculation to evaluate the two approaches.

Results: The geochemical and geodynamic approaches to areotherm calculation (Fig. 1) result in slightly different areotherms depending on the assumptions, with the geodynamic results (Fig. 1A) indicating melting was still occurring at 1.9 Ga, whereas a slightly more rigorous geochemical calculation based on the geodynamic q_0 (Fig. 1C) indicates that no melting was occurring at that time. This is in contrast to the purely geochemical approach, where our preferred model (that matches the high concentrations of HPEs indicated by the InSight data [23]) indicates melting to the present day (not shown), consistent with the observation of recent volcanic activity. The difference results primarily from the different q_0 values, as the

CitcomS models run colder than the geochemical models.

The geochemical model that includes the calculation of F and melt compositions, were run through Perple_X that calculated density and V_S profiles through time (Fig. 2). Overall, all properties decrease with depth through the crust and mantle lithosphere and increase with depth in the mantle. The crust-lithosphere boundary (Moho) is very pronounced with a sharp increase in all properties, while the thermal lithosphere-asthenosphere boundary (LAB) is relatively more subdued kink. We are testing layered models in accordance with InSight results.

Conclusions/Implications: Overall, initial results of the geodynamic and geochemical approaches to calculating areotherms are in general agreement, though vary in the details. This integration needs further refinement. The results from the geochemical model that includes melting suggest that the mantle lithosphere is slightly denser than the mantle, though that difference decreases with time, though this would suggest some minor overturn/mixing early in the planet's history.

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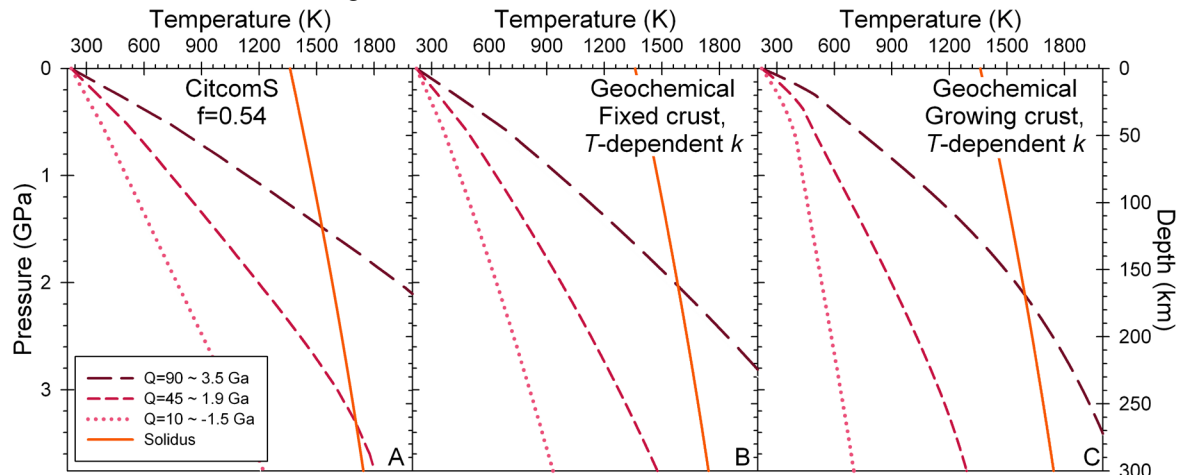
Figure 1. Comparison of the **A)** geodynamic (f=0.54 ≈ 1830 km radius core) and **B,C)** geochemical areotherm calculations. **A)** CitcomS calculations used a fixed HPE concentration [10], a fixed crustal thickness of 53 km, and a fixed k of 4 W/mK, the output was a q_0 (36, 20, 11 mW/m^2) and time (3.5, 1.9, -1.5 Ga). Those outputs were used as inputs to the geochemical method. **B)** This calculation also used a fixed HPEs [10] and crustal thickness of 53 km but varied k as a function of T [13,14]. **C)** This calculation used fixed HPE concentrations for the crust [11] and depleted mantle lithosphere 50% of [10], the same k as in B, and a linearly growing crust starting at 20 km.
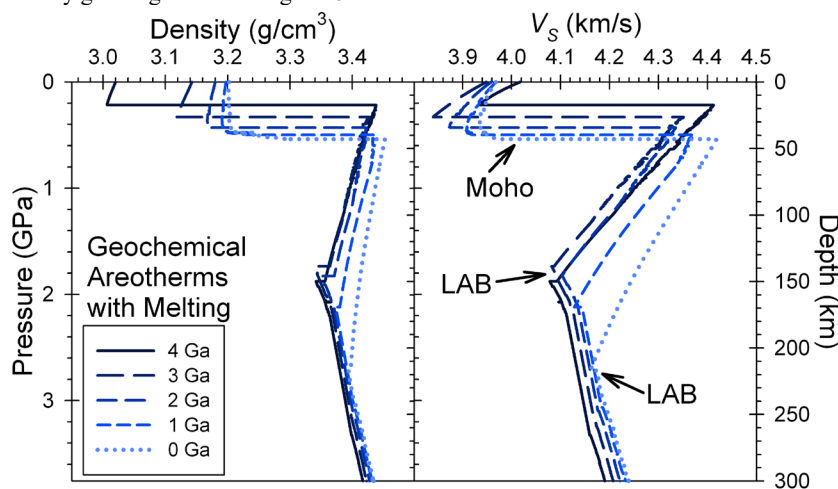
Figure 2. Geochemical areotherm + Perple_X results through time (not related to CitcomS results). These areotherms were calculated with a full melting model, therefore the crust and lithosphere thickness and HPE distribution (and overall composition) were changing with each timestep related to the F calculation. The average crust K concentration at the modern day (0 Ga) for this particular model is ~5,000 ppm, which is similar to the 3 layer crust model of InSight [23]. The crustal density does not take impact gardening or weathering products into account, this is the initial basalt density.