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Effect of Crustal Heating and Insulation on Heat Transport in Mercury

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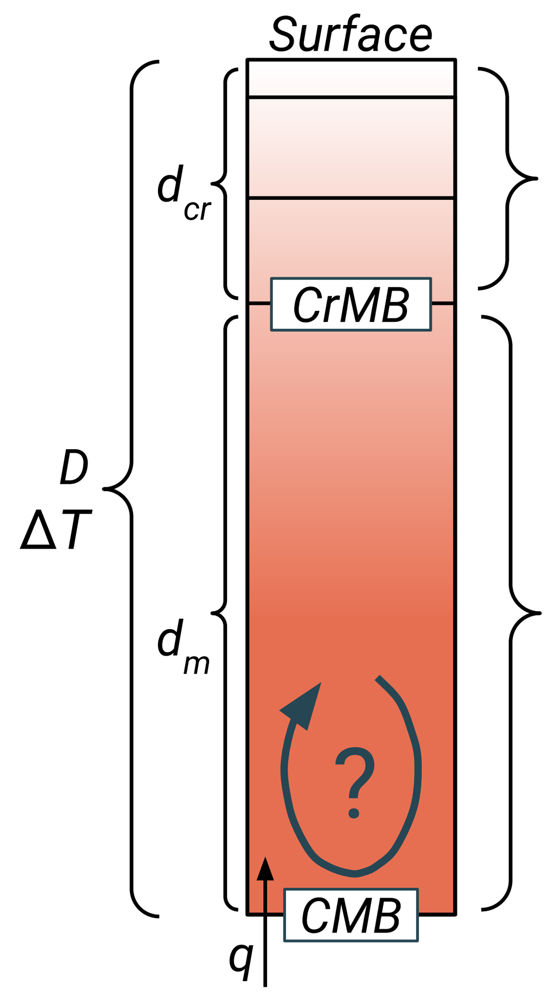
The model



Conductive crust with *n* homogeneous layers. Layer *i* has:

- Thickness d_i
- Thermal conductivity k_i
- Volumetric heating h_i

Includes buoyant sulfide layer.

Silicate mantle may convect if conditions permit.

- **No volumetric heating or melting**
- Temperature-dependent viscosity
- Heat flux across convecting layer ~ $Ra^{1/3}$

Rectangular geometry, steady state.

Layered heat conduction

We solve the heat equation to relate bottom heat flux *q* to total temperature drop ΔT and layer properties:

$$\frac{D}{[k]} = \sum_{i=1}^{n} \frac{d_i}{k_i}$$

Layer conductivities and thicknesses are combined into **"effective conductivity" [k].**

$$q = \frac{[k]}{D}(\Delta T - H)$$

H has units K, equivalent to higher surface temperature. X_i is a value between 0 and 1 representing how much insulation is above layer *i*.

$$H = \frac{D}{[k]} \sum_{i=1}^{n} h_i d_i X_i$$

$$X_i = \frac{\frac{d_i}{2k_i} + \sum_{j=i+1}^{n} \frac{d_i}{k_i}}{D/[k]}$$

Values for Mercury

$\frac{d}{k}$: Lower crust: ~10,000; Mantle: ~100,000; 5 km regolith: ~25,000

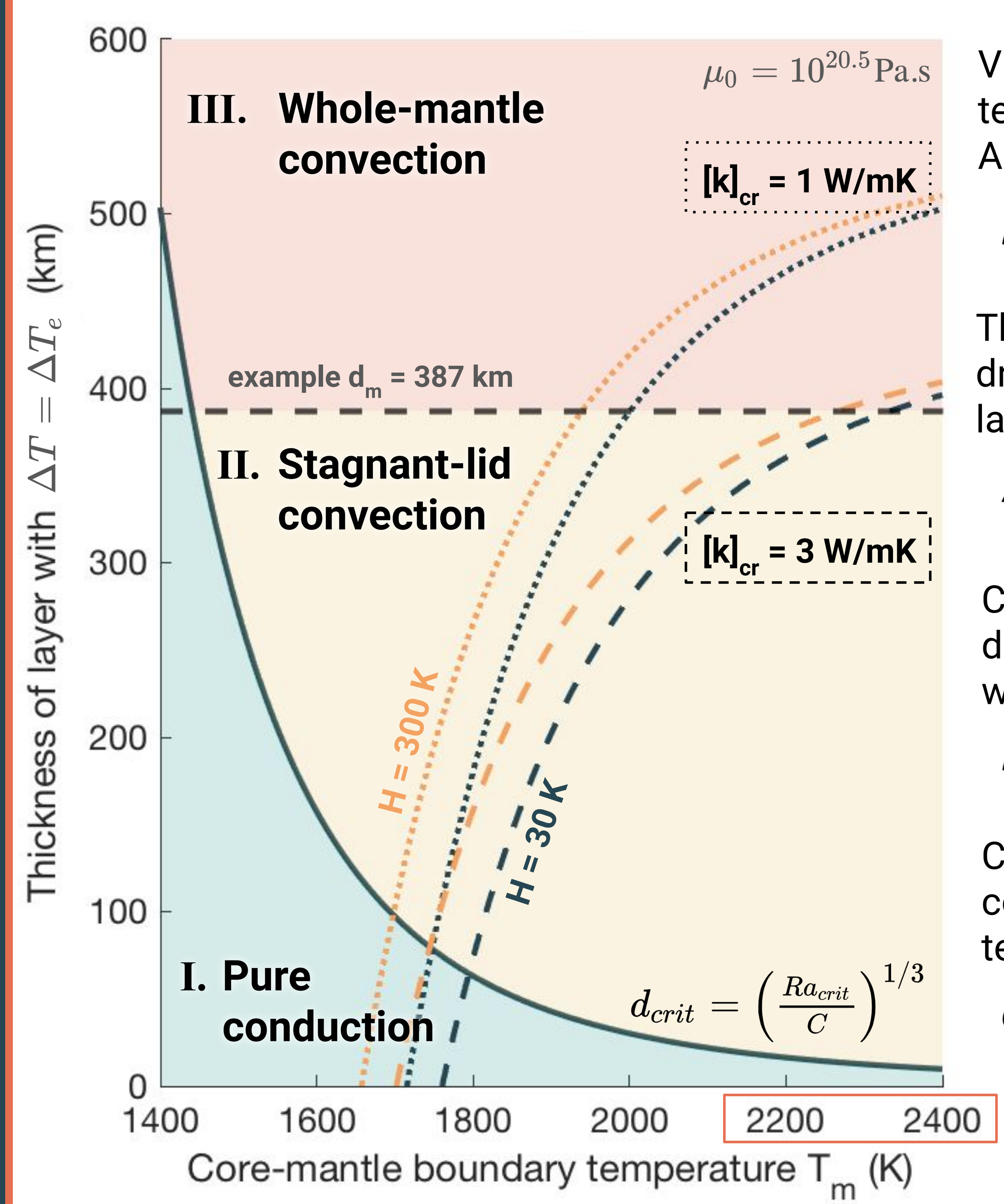
$[k]$: Crust: 1-3 (for 0-5 km megaregolith.); Crust+mantle: ~3

H: Crust: 10-50 K now, 10x more in earliest history (using GRS surface concentrations[6] and 0-5 km megaregolith.); Sulfide layer: 0-600 K now (depending on bulk radiogenic element concentrations and partitioning.)

Airless bodies such as Mercury and the Moon have a thick, strongly insulating megaregolith[1], suggested to have slowed planetary cooling.[2,3] Mercury may also have persistent sulfide layering in its mantle[4], which may insulate or (if it sequestered radiogenic elements) generate heat.

Regimes of heat transfer

Three regimes of heat transfer are defined by the thickness of the warm layer with $\Delta T = \Delta T_e$ relative to d_m and the critical thickness for convection d_{crit}.



Viscosity is strongly temperature-dependent, Arrhenius rheology.

$$\mu = \mu_0 \exp\left[\frac{A}{R}\left(\frac{1}{T_m} - \frac{1}{T_0}\right)\right]$$

The maximum temperature drop across the convecting layer is dictated by viscosity.[7,8]

$$\Delta T_e = \frac{\mu}{\frac{d\mu}{dT}}\Big|_{T_m} = a_{rh}\frac{RT_m^2}{A}$$

Convecting layer thickness depends on temperature as well as layer properties.

$$k_m \frac{D}{[k]} - d_{crit}\left(\frac{\Delta T - H}{\Delta T_e} - 1\right)$$

Critical thickness of convective layer depends on temperature and viscosity.

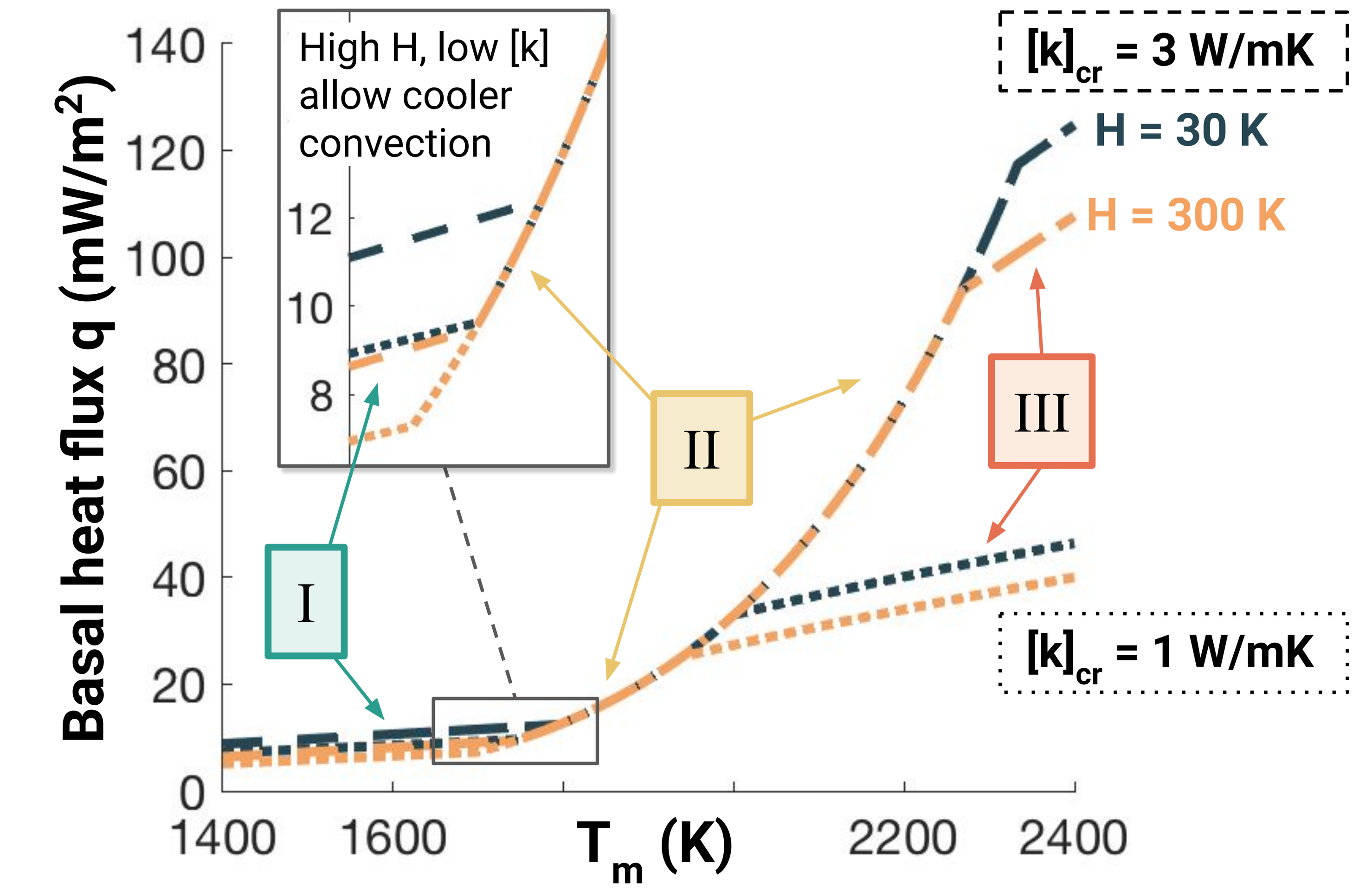
$$C = \frac{Ra}{d^3} = \frac{\rho_m \alpha_m g \Delta T_e}{\kappa \mu}$$

Crustal properties affect the regime expected for a particular temperature.

A crust with layered properties can be described with two terms: [k]/D and H. Crustal properties weakly affect heat flux except at high temperature, but strongly affect upper mantle temperature. In this model, regolith would only have slowed planetary cooling if lower mantle temperatures were very high or very low for long.

Conclusions: heat flux

The crust affects heat transport within each regime (mostly) as well as the system's regime at T_m.



$$q_I = (\Delta T - H)\frac{[k]}{D}$$

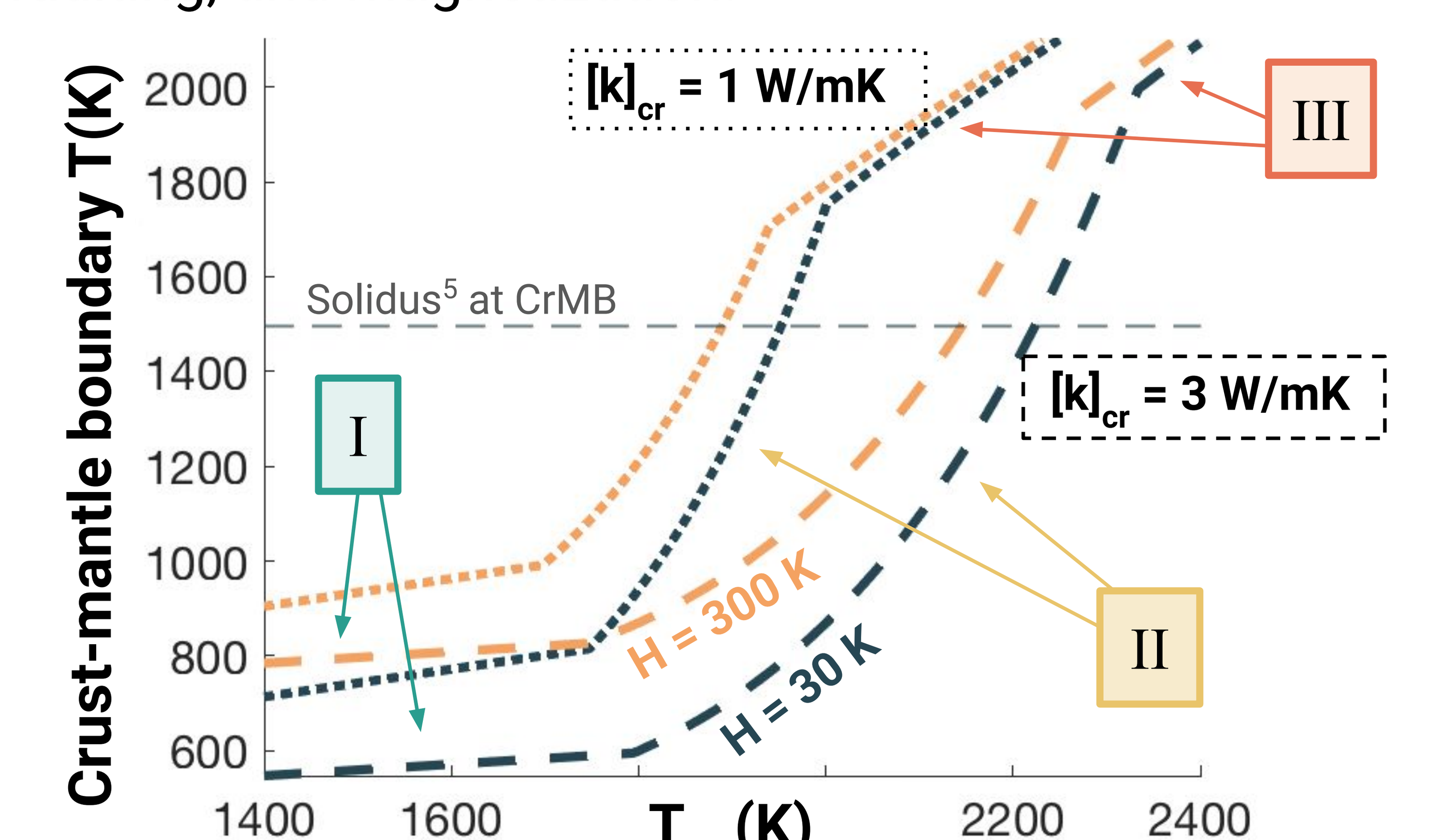
$$q_{II} = \Delta T_e \frac{k_m}{d_{crit}}$$

$$q_{III} = \frac{\Delta T - H}{\frac{d_{cr}}{[k]_{cr}} + \frac{d_{crit}}{k_m}}$$

Crustal properties have the most effect at high and low temperatures, but NONE in stagnant lid regime.

Conclusions: temperature

Crustal properties strongly affect upper mantle/lower crust temperature, with implications for volcanism, faulting, and magnetization.



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