Supporting Information for "Prolonged multi-phase magmatism due to plume lithosphere interaction as applied to the High Arctic Large Igneous Province"

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Introduction

This Supporting Information contains additional information on the chosen mantle temperature profile and the solidus used in calculating melting, and how this relates to whether numerical models have to be compressible or not.

Text S1.

As described in the main text, the models used in this study are incompressible. This assumption is commonly used in upper mantle models, since it has been shown that the effect of compressibility is small in the upper mantle where pressures are low (Albers & Christensen, 1996). The usage of the Boussinesq approximation decouples the conservation equations of mass and momentum by assuming that the density is not pressuredependent, which makes the equations easier and cheaper to solve. Therefore, the advantage of reduced computational costs is typically larger than the error introduced by the simplification.

However, pressure does not only affect the density, but also the temperature of the mantle. Compression increases the temperature linearly with depth, resulting in an adiabatic temperature increase across the mantle on top of the heating from the core and radioactive decay following the equation:

$$T = T_{\mathrm{pot}} \exp \alpha g d / C p \tag{1}$$

with the mantle potential temperature T_{pot} (the extrapolation of the mantle adiabat to the surface), the thermal expansivity α, the gravitational acceleration g, the depth d, and the specific heat capacity C_p. Values for *alpha* and C_p from the models are $3.5 \cdot 10^{-5}\,1/\mathrm{K}$ $1250\,\mathrm{J/(kg{\cdot}K)}$, respectively. This temperature increase should not be included when using the Boussinesq approximation. While our models do have a slight increase of temperature with depth (see Figure S1a), this increase is smaller than the mantle adiabat (Figure S1b) and follows the setup of Heyn and Conrad (2022), whose models we build on. The slight temperature increase serves the purpose of decreasing the formation time of the plume at the beginning of the model. However, as can be seen in Figure the difference between the melting region for a compressible case and our chosen setup are insignificant. Both the onset depth of melting as well as the melting interval are comparable, with the minor differences being smaller than the uncertainty in the parameters involved in calculating melting. Using a constant temperature profile corresponding to the mantle potential temperature of our profile (Figure S1c) would also result in a very similar melting region. In all three cases, the temperature is used to calculate the density and pressure at each depth, and the pressure is then used to calculate the solidus. For the case of the incompressible profile with constant temperature within the upper mantle, the compressible solidus is further corrected for the mantle adiabat.

In total, we ran 13 models, 6 with stagnant plate and 7 with moving plate. The initial LAB geometries for each 6 cases are:

- 1 step: one step in lithosphere thickness at 500 km from the impigning plume, stepping from 100 km thick basin to 150 km thick continental lithosphere (see Figure 8b); $(\partial T/\partial p)_s = 7.8 \cdot 10^{-8}$ and $(\partial T/\partial p)_m = 7.8 \cdot 10^{-8}$
- 1 step deeper: same as above, but with lithosphere thicknesses of 150 km and 200 km, respectively; $(\partial T/\partial p)_s = 7.8 \cdot 10^{-8}$ and $(\partial T/\partial p)_m = 5.8 \cdot 10^{-8}$
- 2 steps: two consecutive steps 500 km apart, with the first step at the position of plume arrival. Lithosphere thickness increases from 100 km to 150 km to 200 km (Figure 4a and 5a); $(\partial T/\partial p)_s = 7.8 \cdot 10^{-8}$ and $(\partial T/\partial p)_m = 6.8 \cdot 10^{-8}$
- 2 steps, symmetric around a thinner lithosphere (Figure 8c). The plume impinges on the first step; $(\partial T/\partial p)_s = 7.8 \cdot 10^{-8}$ and $(\partial T/\partial p)_m = 7.8 \cdot 10^{-8}$
- 1 ramp: a smooth transition of 500 km width between a 50 km thin basin and a 100 km thick craton (Figure 7a), with the plume hitting at the beginning of the ramp; $(\partial T/\partial p)_s = 10.8 \cdot 10^{-8}$ and $(\partial T/\partial p)_m = 9.0 \cdot 10^{-8}$
- 1 ramp deeper: same as above, but with a transition between a 100 km thick basin and a 200 km thick craton; $(\partial T/\partial p)_s = 7.8 \cdot 10^{-8}$ and $(\partial T/\partial p)_m = 7.8 \cdot 10^{-8}$

In addition, we ran one model with moving plate and constant LAB depth of 100 km and a $\partial T/\partial p = 7.6 \cdot 10^{-8}$. The only differences between the models are the LAB depth and the pressure gradients of the solidus given above and arked by subscript s for stagnant and m for moving plate.

References

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Heyn, B. H., & Conrad, C. P. (2022). On the relation between basal erosion of the lithosphere and surface heat flux for continental plume tracks. *Geophysical Research Letters*, *49*(7), e2022GL098003. Retrieved from https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2022GL098003 (e2022GL098003 2022GL098003) doi: 10.1029/2022GL098003

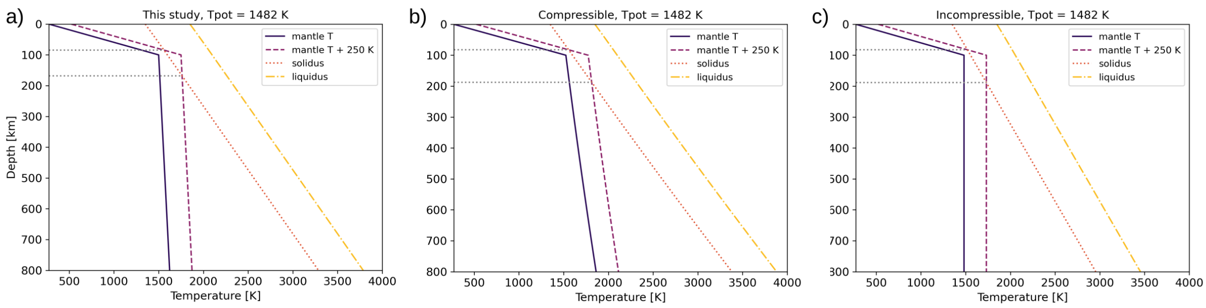

Figure S1. Mantle temperature profiles and solidus profiles for (a) our models, (b) a compressible case with the same mantle potential temperature, and (c) an incompressible case in which the mantle temperature below the lithosphere corresponds to the mantle potential temperature. In all three cases, the surface solidus is set to 1350 K and the pressure gradient is $7.8 \cdot 10^{-8}$ K/Pa. The liquidus is assumed to be 500 K hotter than the solidus.