

Evidence for a very thick Kaapvaal craton root: implications for equilibrium fossil geotherms in the early continental lithosphere

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Abstract

The presence of diamonds of Archaean age within the cratonic lithosphere is highly surprising because their formation implies a cool early mantle lithosphere despite the early Earth's hotter state. There remains a widely (albeit not exclusively) held concept that the prevailing geothermal gradient within Archaean cratons was not much different from today. This 'Archaean paradox' is supported by pressure and/or temperature (PT) estimates of potentially ancient diamond inclusion (DI) falling close to modern cratonic geotherms. Solutions to the paradox include more effective Archaean heat loss mechanisms (e.g. heat pipes) and deeper lithospheric mantle roots whose fossil equilibrium geotherms mimic modern cratonic geotherms at the depths from which most xenoliths are sampled.

Here we demonstrate the effects of a thermally evolving Earth on equilibrium geotherms within the Kaapvaal craton of South Africa. We modelled geotherms from the present-day conditions backward in time to 3.0 Ga at thermal equilibrium in the lithosphere by adjusting three dominant variables: (i) time-corrected crustal heat production; (ii) a secularly cooling mantle potential temperature (T_p) in the underlying convecting mantle; and, for illustrative purposes, (iii) variable depth of the lithosphere-asthenosphere boundary (LAB) or variable heat flow at the LAB.

Computed model geotherms were compared to DI PT data from the Kaapvaal craton, as well as PT conditions recorded in mantle xenoliths of variable age. It is demonstrated that equilibrium geotherms in the lithosphere cannot be reconciled with PT conditions recorded in Kaapvaal craton DI nor with those recorded in Proterozoic xenoliths if either the lithosphere has remained of constant thickness or held a constant basal heat flux since 3.0 Ga. Rather, provided that PT conditions recorded by DI from the Kaapvaal craton reflect equilibration along fossil ancient

equilibrium geotherms, it is required that the early lithosphere was substantially thicker than what is preserved today. The inferred former thickness of Kaapvaal craton could have reached up to ~350 km if the prevailing mantle T_p at the time of diamond encapsulation was high and they are of Archaean age. If, however, the Archaean asthenosphere T_p was more modest or if the diamonds formed later than the Archaean the inferred LAB depths are shallower (~250-275 km), but still substantially deeper than present-day.

Although only the Kaapvaal craton has the necessary DI PT constraints for this exercise, the fundamental nature of radioactive heat production rate and mantle cooling mean that our findings could imply that Earth's continental lithosphere in general attained its maximal vertical extent early in the Archaean and that cratons have been variably eroded and weakened since that time.

Introduction

Diamond is stable in the Earth's mantle at elevated pressure and low to moderately high temperature. At a depth of 150 km, diamond stabilises below $\sim 1250^{\circ}\text{C}$ (Day, 2012) conditions that are found mostly in the present-day deep cratonic lithosphere. It is evident that some peridotitic (P-type) diamond was already stable within the lithosphere back in the Archaean with samples from ca. 2.7 Ga lamprophyres, volcanoclastics and metaconglomerates in the Wawa region of in the Superior craton (Stachel et al., 2006; Miller et al., 2012). Furthermore, dating of P-type diamond inclusions (DI) commonly yield Archaean ages in most cratonic regions (Howell et al., 2020). This dictates that portions of deep lithosphere must have been relatively cool as far back as the Archaean. This finding, combined with evidence for modest Archaean continental geotherms (e.g., Burke and Kidd, 1978; Boyd et al., 1985) is surprising in view of the 2-3 times higher radioactive heat production (Fig. 1a). Indeed, most current models of terrestrial mantle evolution propose that 2.7 Ga before present, the mantle potential temperature (T_p) was 150-200 $^{\circ}\text{C}$ hotter than today (Herzberg et al., 2010). As a result, temperatures in the cratonic lithosphere would be expected to have been substantially higher in the Archaean, inhospitable conditions for diamond formation (Ballard and Pollack, 1988). Solutions to this the so-called 'Archaean paradox' (e.g. Lenardic, 1998) include greater early heat loss through the oceanic lithosphere (e.g., Bickle, 1978; Burke and Kidd, 1978; Lenardic, 1998), heat loss from heat pipes (i.e., via advection; Moore and Webb, 2013), lithosphere emplaced in an initially (cool) disequilibrium state (e.g., Michaut et al., 2009; Moore and Lenardic, 2015) or formation of a much thicker early lithosphere (Ballard and Pollack, 1988).

A prevailing view that fossil Archaean cratonic geotherms were 'similar' to modern ones, has encouraged comparison of pressure-temperature (*PT*) conditions recorded by inclusions trapped in diamond inclusions (DI) to equilibrium geotherm 'families' (e.g., Hasterok and Chapman, 2011) or models (e.g., Mather et al., 2011) parameterised at the present day. The contribution of radioactive decay of ^{40}K , ^{232}Th , ^{235}U and ^{238}U to the surface heat flux 1 Ga ago was only modestly higher than today's ~5 TW (Fig. 1a), perhaps explaining the similarity in *PT* data recorded in xenoliths from Proterozoic and Cretaceous kimberlites in South Africa (Fig. 2a-b) and justifying plotting DI *PT* conditions onto modern geotherms. However, from 1 to 3 Ga the difference in radioactive heat production is much more pronounced, with resulting radiogenic contributions to the heat flow at the Earth's surface almost doubling from 25 TW at 1 Ga to 48 TW at 3.0 Ga (Fig. 1a). This great difference in radioactive heat production invalidates the concept of using modern geotherms to evaluate the formation conditions of Archaean DIs.

This is illustrated simply by considering the heat production within the cratonic crust alone in a continental lithosphere losing heat purely by conduction. A theoretical present-day cratonic crust of 35 km thickness differentiated into an upper crust of 17 km thickness of a density of 2750 kg/m³ with internal heat generation of 1.4 $\mu\text{W}/\text{m}^3$ and a lower crust of 18 km thickness of a density of 2860 kg/m³ with an internal heat generation of 0.4 $\mu\text{W}/\text{m}^3$ would presently contribute 31 mW/m² to surface heat flux, compatible with observations. The theoretical crustal contribution to surface heat flow can be calculated back through time using, K, Th and U abundances and K/U and K/Th ratios of 1.0 and 0.26 in the upper crust (where K is in wt % and Th and U in ppm) derived from the global average of 3.5 to 2.5 Ga Archaean upper crust (Condie, 1993), and K/U and K/Th of 2.4 and 0.56 for the

lower crust xenolith database of Rudnick and Fountain (1995). This demonstrates that during the Mesoproterozoic crustal contribution to the surface heat flux alone would have surpassed typical present-day cratonic surface heat fluxes worldwide of $\sim 36\text{--}50\text{ mW/m}^2$ (Jaupart and Mareschal, 2007). At 3.0 Ga the calculated crustal contribution to the surface heat flux in this example is $\sim 75\text{ mW/m}^2$. Hence, during the Archaean, a mantle lithosphere experiencing equilibrium conductive heat loss cannot have had geothermal gradients comparable with modern day. Burke and Kidd (1978) proposed ‘similar’ geothermal gradients in the *crust* during the Archaean with a modest $6\text{ }^\circ\text{C/km}$ decrease from the Archaean to the present day. In a 37 km thick crust this would still equate to a 222°C difference at the Moho. Even this modestly different crustal geothermal gradient in the Archaean would manifest as a differing geothermal gradient in the mantle, which depending on prevailing basal heat flux, could vary substantially from that at the present day (Fig. 1c). Many other studies have proposed higher Archaean crustal geothermal gradients to explain the preponderance of high- T granulites in Archaean metamorphic belts.

Hence, qualitative inferences drawn from comparison of DI PT data with modern geotherms, and particularly when inclusions are projected onto those geotherms (where estimates of either P or T are missing), about diamond sampling intervals (e.g., Nimis et al., 2020), formation environments (e.g., Korolev et al., 2018) and the thermal conditions of the cratonic root in the Archaean (e.g., Miller et al., 2012), could be subject to error.

This contribution provides theoretical fossil equilibrium geotherms for a secularly cooling Earth using the example of the cratonic lithosphere of South Africa. The modelling considered the two dominant variables: crustal heat production through time (Fig. 1b) and the evolving T_p in the underlying convecting mantle (Fig.

1a). We generated models by varying the depth of the lithosphere-asthenosphere boundary (LAB) or the heat flux at the LAB. The resulting equilibrium geotherms were compared to *PT* conditions recorded in mantle xenoliths of near-modern and Proterozoic age in addition to P-type DIs of at least 1.15 Ga age but presumed to be of earlier (Palaeoproterozoic or Archaean) age (Richardson et al., 1993). From this evaluation, conclusions are drawn regarding the depth of the LAB in cratons earlier in Earth's history and the chances of survival of Archaean diamonds.

Methodology

Geotherm calculation

The present day geothermal gradient in the cratonic lithosphere, defined by conductive equilibrium, is expressed by:

$$k \frac{d^2T}{dz^2} = -A$$

[1]

Where, k is thermal conductivity, A is heat production, T is temperature and z depth. For a specified lithospheric structure (e.g., heat production, layer divisions and thicknesses, density), thermal conductivity, and surface heat flux, 1D equilibrium geotherms can be calculated downward from the Earth's surface ($T = 0^\circ\text{C}$) using a discretised form of the heat equation (in layers of arbitrary thickness) for the upper crust, lower crust and lithospheric mantle for a Cartesian geometry;

$$T_{i+1} = T_i + \frac{q_i}{\lambda_i} \Delta Z_i - \frac{A_i}{2\lambda_i} \Delta Z_i^2$$

[2]

and,

$$q_{i+1} = q_i - A_i \Delta Z_i$$

[3]

where A_i , λ_i , ΔZ_i , are inter-layer heat production (in $\mu\text{W}/\text{m}^3$), inter-layer thermal conductivity (in $\text{W}/\text{m}\cdot\text{K}$) and layer thickness (in km), respectively. The vertical column is divided into arbitrarily thin layers (0.1 km), for which properties can be considered constant. Temperature, T_{i+1} ($^{\circ}\text{C}$) and the heat flux, q_{i+1} (mW/m^2) at the bottom of each layer are determined from the temperature, T_i , and heat flux, q_i at the top of each layer. As we use a T -dependent thermal conductivity model, a Newton-Raphson iterative scheme is employed to solve for temperature T_{i+1} and the intra-layer thermal conductivity K_{i+1} downward from $T = 0^{\circ}\text{C}$ at the Earth's surface (e.g., Hasterok and Chapman, 2011). The thermal conductivity model used in this study is presented in the supplementary material. The assumed Cartesian geometry is valid only when the lithosphere is thin relative to the radius of the planet, as is the case for the Earth.

The vertical extent of the lithosphere can either be set as an input parameter by iterating the surface and thus mantle heat flux or allowed to emerge as a result from a fixed mantle heat flux. The intercept of the calculated lithospheric geotherm and the mantle adiabat, which is itself a reflection of T_p (Rudnick et al., 1998), provides the depth to the LAB in the model. As explained earlier, T_p in our models was varied to explore the effect of secular cooling of the convecting mantle. The model yields an evolving Urey ratio, presently of 0.38 (Fig. 1a). This thermal model has been argued to provide a good fit to petrological estimates of T_p through Earth's past for non-arc basalts of known age (Herzberg et al., 2010).

In the strictest sense, the intercept between a conductive geotherm and the mantle adiabat does not coincide with a 'real' rheological boundary or the LAB. In reality, a thermal boundary layer (TBL) exists as part of the convecting upper mantle in which the dominant mechanism of heat transfer switches from conduction to advection (McKenzie et al., 2005). As such the intercept of a conductive geotherm and the adiabat will provide an overestimated depth to the LAB. However, besides simplicity, an added benefit of projecting the conductive geotherm on the adiabat is that it eliminates secular uncertainties on parameters used to calculate the TBL (e.g., mantle viscosity) and therefore ensures consistent comparisons of geotherms through time.

Applicability of equilibrium geotherms

Due to the long tectonic quiescence of cratons, the thermal conditions within their roots are widely assumed to approximate a state of thermal equilibrium. However, thermal transients have occurred. Firstly, in an evolving Earth heat generation decayed with time, meaning that thermal equilibrium is never truly attained (Michaut and Jaupart, 2007). In the cratonic lithosphere, following isolation from convecting mantle, the maximum crustal radiogenic temperature component at the LAB would be reached after 1-2 Gyr (Jaupart and Mareschal, 2007). Thus estimates of the surface, Moho and basal heat fluxes in our ancient equilibrium models (2 to 3 Ga) will likely be overestimated, whilst those closer to the present-day (0 to 1 Ga) underestimated. Radiogenic heat production within the lithospheric mantle, widely agreed to be very low $<0.02 \mu\text{W}/\text{m}^3$ (e.g., Michaut et al., 2007), would have further amplified thermal transients but are not relevant to our model with no internal lithospheric mantle heat generation. Existing secular geotherm models accounting

for these long-term thermal transients require a lithosphere of fixed thickness and cannot be employed here as we explore differing depths to the LAB.

Episodic geological events have also punctuated the otherwise quiescent tectonic and magmatic history of cratons. The Kaapvaal craton's thermal state was perturbed by the emplacement of the 2.05 Ga Bushveld Igneous Complex (BIC), the Jurassic Karoo flood basalts, and the Mesozoic kimberlites that brought cratonic xenoliths to the surface. Importantly, the *PT* arrays constructed from Kaapvaal cratonic xenoliths in Cretaceous kimberlites conform with equilibrium, rather than perturbed, geotherms that are wholly consistent with present day surface heat flux measurements, despite postdating Karoo magmatism by just 60-100 Ma. The only exceptions are metasomatized or deformed samples that are displaced from equilibrium geotherms (e.g., Mather et al., 2011), suggesting that, although Karoo intraplate lavas are widely dispersed at surface, their thermal perturbation of the lithosphere was confined leading to enhanced lateral, rather than vertical heat transfer, followed by quite rapid decay of heat anomalies (Jaupart and Mareschal, 2007). This inference could be supported by the lack of present-day heat flow anomalies over the Deccan traps (65 Ma) and the Parana Basin (120 Ma) in two other shield regions (Hurter and Pollack, 1996; Roy and Rao, 2000). We therefore approximated the modelled thermal state of the bulk of the Kaapvaal lithosphere under equilibrium conditions (with heat flux at the Moho equal to that at the LAB), recognising that a minor present-day heat flow component could still derive from Karoo magmatism. Earlier major events Umkondo and Bushveld LIPs could also have had similar effects. Our secular equilibrium geotherms could underestimate the thermal effects of such events. In practice we argue that long wavelength transients arising from crustal heat source reorganisation and the potential secular cooling of

the convecting mantle (Michaut and Jaupart, 2007) are greater sources of model uncertainty.

Constraining the near-modern thermal state of the Kaapvaal Craton

To anchor secular modelling, we calculated an ‘initial’ geotherm representing close to present-day (~85 Ma) thermal condition of the Kaapvaal lithosphere (Fig. 2a). The input parameters are provided in the supplementary information for the calculation are consistent with what is widely considered ‘typical’ of the present-day Kaapvaal lithosphere (Table S1). Internal heat production of the lithospheric mantle itself is the most difficult parameter to determine due to conflicting estimates of K , Th and U concentration in mantle xenoliths arising from ubiquitous contamination of mantle xenoliths and secondary addition of heat-producing elements during ascent in their host kimberlite. This explains the dichotomy in the measured heat production of mantle xenoliths sampled by basalts and kimberlites (Rudnick et al., 1998), with the former yielding overwhelmingly low heat production values of $< 0.02 \mu W/m^3$. Xenoliths transported by the vastly more incompatible element-rich kimberlites such much wider range and generally higher values up to $0.1 \mu W/m^3$ (Rudnick et al., 1998). Instead of arbitrary correction for contamination, we preferred an internal lithospheric mantle heat production of $0 \mu W/m^3$, dictating that at any one time the Moho heat flux is equal to that through the LAB (i.e., in steady state).

The geotherm was fitted to PT equilibration conditions in mantle xenoliths from Cretaceous kimberlites, extracted from the database of Wu and Zhao (2011), which provides a good geographical cover of the Kaapvaal craton (for information on thermobarometry see the supplementary information). The only Cretaceous PT data not considered were those of mantle xenoliths from Finsch, due to their peculiar

offset towards lower temperatures at a given depth (Fig. 3a). Possible interpretations for these unusual data are discussed later.

The preferred near-modern geotherm (Fig. 2a) was obtained iteratively and yields the following parameters: a surface heat flux of 47 mW/m², a Moho and LAB heat flux of 15.8 mW/m² and a lithospheric thickness of 195 km, calculated with a modern T_p of 1350 °C (Herzberg et al., 2010). The model surface heat flux matches the independently derived 47 ± 7 mW/m² average of surface heat flux measurements taken from 144 localities over the geographical region of the Kaapvaal craton (Jones, 2017). The geotherm produces a very reasonable fit to mantle xenolith *PT* data for xenoliths of < 150 km depth where a conductive equilibrium state is widely inferred. Inspection of the residual plot between modelled and calculated xenolith temperature (Fig. 3a) reveals a cluster of deep xenoliths at >150 km, that irrespective of location (i.e., Western vs Eastern Kaapvaal vs Lesotho) plots at temperatures 100-200°C above the model geotherm. These most likely represent thermally perturbed xenoliths close to the LAB, recording transient (i.e., non-steady state) thermal conditions.

We opted to constrain the geotherm via close approximation to the known surface heat flux of the Kaapvaal craton while achieving a semi-quantitative fit to *PT* data for xenoliths of < 150 km depth. An alternative is to statistically minimise the misfit between the geotherm and the *PT* array from non-perturbed xenolith, requiring classification into pristine and thermally perturbed groups. This carries the risk of accidental inclusion of thermally perturbed data, which would artificially decrease the depth to the LAB. Regardless, our near-modern Kaapvaal craton geotherm is similar to previous estimates from xenolith studies (e.g., Mather et al., 2011) and independent estimates of LAB depth from geophysical studies (e.g., Ravenna et al.,

2018). Therefore, we consider it to be a robust starting point for exploring fossil geotherms.

Constraints from Proterozoic mantle xenolith samples and diamond inclusions

There are two key groups of empirical fossil *PT* constraints for validation of secular geotherms. The first is from the ~1.15 Ga Premier kimberlite (Wu et al., 2013). Its mantle xenoliths hold information about the past thermal state of the Kaapvaal lithosphere at 1.15 Ga. Analogously to the near-modern starting geotherm, a preferred equilibrium geotherm of the Kaapvaal craton at 1 Ga (Fig. 2b) was constructed and this should closely approximate the *PT* array from mantle xenoliths from Premier (Fig. 2b). The second source of relevant *PT* information is from DIs from the Premier kimberlite (Nimis et al., 2020) that likely record thermal conditions in the lithosphere prior to 1.15 Ga because a significant proportion of P-type DIs worldwide have Archaean or Palaeoproterozoic entrapment ages (Howell et al., 2020), although younger entrapment also exists (e.g., Richardson et al., 1993; Koornneef et al., 2017). At Premier, harzburgitic DIs have been used to infer growth in the Archaean whilst lherzolitic DIs have yielded an isochron age of ~1.9 Ga (Richardson et al., 1993). This age may be more relevant because the *PT* estimates of DI used here are obtained from clinopyroxene. Regardless, the DIs must be at least 1.15 Ga old (i.e., the age of the kimberlite), and probably had grown in multiple events, but interestingly, with the coolest *PT* data forming one tight quasilinear array, that could be consistent with growth at similar time and thermal conditions.

There is a clear and marked difference in *PT* conditions of Premier DIs and peridotite xenoliths, with the former plotting well below both the near-modern and the

model 1 Ga Kaapvaal geotherms (Fig. 2c) as noted by previous workers (e.g.,
Korolev et al., 2018; Nimis et al., 2020).

Results

The presentation of initial results was motivated by the common comparison of DI *PT* data with modern day geotherms. In such comparisons, it is implicit that the lithosphere has remained at fixed thickness and experienced an unchanged basal heat flux through time. Keeping both these parameters constant is geologically not possible for a lithosphere in thermal equilibrium, but fixing one parameter serves to illustrate how equilibrium geotherms would change through time. Initial results of geotherm calculations are thus presented as two classes of models. In the first, the heat flux across the Moho (and thus LAB) is fixed, allowing the lithospheric thickness to vary. In the second, the LAB depth is fixed, allowing the heat flux across the Moho (and LAB) to vary. The geotherms were generated by iteratively changing the surface heat flux until it reproduces the Moho heat flux (first class of models) or lithospheric thickness (second class of models) at each time window. All models were completed for both an evolving (secularly cooling) convecting asthenosphere and for one with a fixed temperature corresponding to the modern T_p of 1350 °C. For all models, comparison with diamond *PT* data and the evolution of the 'diamond window' (*PT* conditions in which formation of diamond can occur) is shown to afford easy assessment of compatibility with formation and preservation of diamond within cratonic lithosphere since the Archaean. After a brief evaluation of the initial models, results for more realistic parameter combinations are presented.

Results from models with fixed LAB heat flux and variable lithospheric thickness

The fossil cratonic geotherms with a fixed Moho heat flux of 15.8 mW/m^2 (in 0.5 Ga steps up to 3.0 Ga) are shown for the secularly cooling and fixed T_p in figures 4a and 4b, respectively. The key observation is that with the Moho/LAB heat flux held constant, the effect of increased internal heat production within the crust backward in time leads to significantly higher temperatures in the lithospheric root. This ultimately derives from the fact that, as energy transfer from the underlying mantle is constant through time in this scenario, the increased heat production in the crust must be accommodated by increased temperatures in the cratonic mantle, such that at a depth of 120 km, the temperature at 3.0 Ga would have been $\sim 400^\circ\text{C}$ hotter than today, regardless of whether the asthenosphere T_p was 1350°C or hotter. In this scenario, diamond is not stable at any depth prior to ~ 2.0 Ga in the case of secularly cooling mantle, and ~ 1.65 Ga, in the case of the fixed modern T_p (Fig. 4i and 4ii). This prediction is in clear conflict with the ubiquity of Archaean diamonds from cratonic lithosphere worldwide, including those in the Kaapvaal craton (e.g., Richardson et al., 1984).

The second clear outcome of the models with fixed Moho and LAB heat fluxes, regardless of T_p , is that the LAB is at its minimum depth in the Archaean, but subsequently, in response to the decay of radioactive heat production in the crust, the lithosphere is predicted to have cooled and thickened (Table 1). Secular thickening of the lithosphere since the Archaean disagrees with the vast majority of Re-Os ages of depleted cratonic peridotites that show Archaean stabilisation when unperturbed by later metasomatism (e.g., Pearson and Wittig, 2008).

A third observation is that the 1.0 and 1.5 Ga modelled lithosphere structures do a very poor job at reproducing the *PT* data from mantle xenoliths from the ~1.15 Ga Premier kimberlite (Figs. 5a; 5b). This is particularly problematic for the model of a secularly cooling asthenosphere in which the LAB is < 180 km in the Mesoproterozoic. Even for the xenoliths that equilibrated at shallower depths (i.e., above the modelled LAB) there is a notably poor fit of the 1.0 and 1.5 Ga modelled geotherms with the *PT* data for mantle xenoliths from Premier, apparently having equilibrated at lower temperature at any given pressure.

The final observation is that none of the modelled geotherms at any time interval are at all consistent with DI thermobarometry. The recorded *PT* conditions in DIs are consistently strongly offset below the model geotherms (Figs. 4a; b). This observation is valid regardless of the true age of the DI.

Results from models with fixed lithospheric thickness and variable basal heat flux

Models with fixed lithospheric thickness, in this example 195 km (Figs. 4c; d), display more modest increases in fossil lithosphere temperature. Whether with fixed or secularly cooling convecting mantle, a diamond window persists into the Archaean. At a qualitative level this is consistent with the presence of Archaean diamonds in the Kaapvaal mantle lithosphere (Figs. 4iii; 4iv).

However, as before, comparison with xenolith *PT* data from the ~1.15 Ga Premier kimberlite cannot be reconciled with the geotherm model of a fixed LAB depth for a secularly cooling convecting mantle due to the consistently hot modelled geotherm relative to conditions recorded in the xenoliths (Fig. 5c). It is worth noting that the Premier kimberlite is emplaced near the BIC and is known to have erupted

through anomalously thick crustal crust (~45 km) compared to that of the wider Kaapvaal craton. The thickened crust near Premier most likely resulted from mafic underplating in the lower crust associated with the emplacement of the BIC (Youssof et al., 2013). Such underplating could result in thermal conditions deviating from those predicted by our models which use data mostly from the Kimberley region of the Western Kaapvaal and Lesotho, with a total crustal thickness ranging from ~36 to 40 km in thickness (Youssof et al., 2013). However, the effect of the localised differences in crustal structure on the thermal conditions in the underlying mantle was likely secondary, because lateral variation of heat production and hence also crustal thickness is effectively attenuated by the large distance over which diffusion occurs (Hawkesworth and Jaupart, 2021). Regardless, a thicker lower crust at Premier and the effects linked to the cooling of such an underplate would only lead to greater temperature in the lithosphere and this increase would simply make models with a fixed lithospheric thickness (or fixed Moho heat flux above) even less reconcilable with mantle xenolith or diamond *PT* data.

Furthermore, whilst the Mesoproterozoic model geotherms at 1.0 and 1.5 Ga with a fixed T_p and LAB depth do provide an acceptable fit to xenolith *PT* data (Fig. 5d) the *PT* conditions recorded by Premier DI do not fit to the model for the Archaean (Fig. 4d). Indeed, across the entire time range explored as above, there is no window that would provide the temperatures low enough at the recorded depths where diamonds could have formed in the modelled Kaapvaal craton (Fig. 4c), even with an optimistically 'cool' modern-day T_p (Fig. 4d).

Together these models demonstrate that neither a cratonic lithosphere of fixed modern-day thickness nor fixed Moho and LAB heat fluxes are compatible with the thermal evolution of the Kaapvaal craton constrained by the presence of

Archaean age diamond, DI *PT* equilibration, and Premier xenolith *PT* data. Their position in *PT* space is consistently below any geotherm (i.e., offset to lower T for a given P) derived from our models. This clearly indicates that the thermal gradient within the lithospheric mantle must have been lower at the time of their formation. Any transient heat inputs associated with large igneous events (e.g., BIC) would only exacerbate this problem. The most straight-forward explanation for a lower geothermal gradient is the former existence of a thicker lithosphere with a consequent lower LAB heat flux (Ballard and Pollack, 1988). The following section presents models for this scenario.

Results from models designed to reproduce DI *PT* arrays

Additional geotherms (Fig. 6) were modelled by iterating the surface heat flux to achieve a reasonable coincidence with the coolest recorded DI *PT* conditions. Target model times were 1.15, 1.9 and 3.0 Ga, to coincide with potential diamond growth events in the Kaapvaal. The 1.15 Ga event could coincide with the emplacement of the Premier kimberlite and the Umkondo LIP (Hanson et al., 2004; Wu et al., 2013) and a diamond forming event in the adjacent Limpopo belt (Koornneef et al., 2017). The 1.9 Ga event would correspond to a diamond forming event recorded in Iherzolitic DI from Premier (Richardson et al., 1993). Whilst the 3.0 Ga event could correspond to the stabilisation of the Kaapvaal mantle lithosphere taken here as the suturing of the western and eastern blocks (Schmitz et al., 2004), which is also consistent with inferred Archaean growth of harzburgitic diamond at Premier (Richardson et al., 1993) and in the adjacent Limpopo belt (Koornneef et al., 2017).

As before, the model geotherms account both for the run-down in heat producing elements in the crust and the potential for a secularly cooling mantle or

one of fixed modern T_p . In all the modelled geotherms (parameters listed in Table 2), the heat flux across the Moho and LAB is substantially lower than in the earlier versions and in all cases, with the resulting lithosphere thicker than today.

Looking first at the 1.15 Ga model geotherm, it is found that if the diamonds that erupted at Premier formed only shortly before emplacement above a secularly cooling mantle, a LAB at 273 km would be required (Fig. 6b) to explain the DI PT array. This would imply, perhaps unreasonably, very substantial (~50 km) and rapid removal of lithosphere to yield the shallower LAB depth required by the xenolith data (Fig. 2b). This seems improbable and instead, the DI far more likely record PT conditions from earlier times. It is worth noting that a subset of clinopyroxene from the Premier kimberlite yield temperatures that are substantially 'hotter' for a given depth, in closer agreement with the thermal conditions recorded by xenoliths at 1.15 Ga (Fig. 2c). It is possible that this subset of DI were derived from post-Archaean growth events, as observed for a suite of DI dated from the nearby Venetia kimberlite. The Venetia DIs yield two separate Sm-Nd isochron regression ages, one at 2.95 ± 0.07 Ga and another at 1.15 ± 0.11 Ga (Koornneef et al., 2017). This latter age overlaps emplacement of kimberlites at Premier and the Umkondo LIP (Hanson et al., 2004; Wu et al., 2013).

Looking next at the 1.9 and 3.0 Ga geotherms best fitting the DI PT conditions, we find that the predicted original LAB was very deep, at 294 and 345 km, respectively (Table 2), for the widely used secularly cooling mantle model of Herzberg et al. (2010). Accordingly, at some point in the past, the Kaapvaal craton could have been significantly thicker. By contrast, using a modern T_p (i.e., no secular cooling of the asthenosphere), more modest lithospheric thicknesses of 256 and 278

km emerge at 1.9 and 3.0 Ga, respectively, but still thicker than the modern day thickness of 195 km.

Discussion

A thicker lithosphere: Evaluating the evidence from diamond inclusions

The first consistent finding of our study is that PT conditions recorded in DI plot below geotherms approximating present-day conditions as well as those for a lithosphere of fixed thickness or fixed base heat flux through time (Fig. 4). The lower thermal gradient recorded by DI is indicative of a geotherm with a lower basal heat flux and hence greater LAB depth in the Archaean and Palaeoproterozoic – provided that most formed at that time. A lower continental basal heat flux in the Archaean supports proposals that the a greater proportion of Earth's heat was lost through the oceans (e.g., Bickle, 1978; Burke and Kidd, 1978; Ballard and Pollack, 1988; Lenardic, 1998). If the diamonds do record PT in the Archaean with a thicker lithosphere, estimates of the temperature at the Moho remain relatively low $\sim 650^{\circ}\text{C}$ consistent with previous estimates of the maximum permissible temperatures in the lower crust ($\sim 800^{\circ}\text{C}$) during the Archaean (Burke and Kidd, 1978).

Evidence for a modestly lower LAB heat flux and thicker Kaapvaal lithosphere in the more recent geological past comes from mantle xenoliths from Finsch and Premier. For a secularly cooling convecting mantle, the LAB depth at 1.15 Ga is calculated to have been 215 km at Premier, 20 km thicker than at 85 Ma (Fig. 2a; 2b) and consistent with other estimates that suggest a slightly thicker lithosphere at 1.15 Ga underlying Premier (e.g. LAB ~ 225 km, Tappe et al., 2021). This could mean that over the last ~ 1 Ga alone, ~ 20 -30 km of refractory lithosphere may have been removed from the base of Kaapvaal craton, but only if a warmer T_p is assumed.

Recent and more rapid erosion at the base of the lithosphere of the Kaapvaal craton may also be consistent with the anomalously 'cool' *PT* estimates obtained from mantle xenoliths from the Finsch group II kimberlite, which was emplaced at 118 ± 3 Ma, earlier than the majority of xenoliths from late Cretaceous (~85 Ma) kimberlites (e.g., Tappe et al., 2018). The position of the Finsch mantle xenoliths in *PT* space fits along a geotherm of a lithosphere of 215 km thickness (but of near present-day heat production), which could represent conditions prior to late Cretaceous LAB erosion (Fig. S6).

Some caution is warranted when comparing absolute *PT* estimates from combinations of various thermobarometers owing to potential artefacts (see the supplementary information for a full discussion). However, direct comparisons of xenolith *PT* data from Premier using either cpx or opx-cpx-grt thermobarometry provide relative *PT* estimates suggesting no systematic offset (Fig. 3d-e) and the supplementary material for further discussion. Despite any remaining uncertainties, one firm conclusion from this study is that it is not valid to project DI temperature estimates from diamonds missing either *T* or *P* estimates onto present day steady state geotherm families, unless the diamond is demonstrably young. Although a common practice, this type of projection should be discontinued as it is possible to produce erroneous depth and temperature information and estimates of the vertical extent of the diamond 'window' (e.g., Fig. 6d).

Uncertainties about the former depth of the lithosphere

The major finding from this study is consistent with predictions of earlier proposals (notably Ballard and Pollack, 1988 and Artemieva and Mooney, 2002) that the vertical extent of the cratonic lithosphere could have been substantially greater in the

past. While the new geotherms calculated here are one step towards more realistic proxies for the early continental lithosphere, they fall short of capturing the full dynamism of cratonic thermal and LAB depth evolution. There are several sources of uncertainty. Firstly, our time-dependent thermal models assume thermal equilibrium and neither take into account the expected thermal disequilibrium introduced by changing internal heat production in the lithosphere nor thermal transients introduced by change in T_p (Michaut and Jaupart, 2007; Michaut et al., 2009). Secondly, changes in thickness of the crust (Condie, 1993) and distribution of heat crustal producing elements caused thermal transients not captured by the model. Thirdly, and most significantly, the former cratonic LAB depth is very strongly dependent on T_p . We evaluated T_p with two end-member models. Using the very conservative constant T_p of 1350°C, our model yields a LAB depth of 278 km at 3 Ga. When T_p is allowed to decrease through time, the calculated LAB depths are much greater (Table 2); 345 and 294 km at 3 and 1.9 Ga, respectively. Furthermore, petrological data used by Herzberg et al. (2010) for their preferred mantle thermal history are permissive of alternative cooling trajectories. Our model uses a present day Urey ratio of 0.38 but if it was lower (e.g., ~0.23), our calculations would underestimate mantle T_p farther back in time, and particularly so in the Archaean (i.e., by ~100°C at 3.0 Ga; Herzberg et al., 2010). Hence, we would underestimate the LAB depth. Finally, it should be stated that the sensitivity of the geotherms to mantle T_p and crustal heat production exert the greatest effect on fossil LAB depth estimates.

Notwithstanding these uncertainties, a substantially deeper LAB for the Kaapvaal craton earlier in its history seems unavoidable. Although our geotherms are specific for this craton and rely on the availability of *DI PT* constraints, it is likely

that the finding of formerly thicker cratonic lithosphere is more widely applicable. We note that more limited DI inclusion *PT* data from other cratons (e.g., Stachel and Harris, 2008) are comparable and that the key parameters imposing deeper LAB, mantle T_p and radioactive heat production rates, are fundamental rather than local. It is interesting to note that some non-plate tectonic models for an early hot Earth predict the development of deep cold lithosphere (Moore and Webb, 2013). In such models, progressive burial of cool surface rocks by catastrophic volcanic resurfacing events (Moore and Webb, 2013) forms a strong cool lithosphere. If the estimates of high Palaeoarchaeon T_p (e.g., Herzberg et al., 2010) are accurate, heat loss via volcanic resurfacing and ensuing deep cool lithosphere could be the expected consequence of the cooling of the Earth's asthenosphere and the surviving cratons would be representative of the early continents rather than being a biased picture of survival of the tectonically fittest.

Mechanisms of lithosphere erosion

Possible mechanisms for thinning of the cratonic lithosphere include thermomechanical erosion via vigorous convection or basal drag and/or interaction with upwelling mantle plumes, with erosion from the former gradual and the latter likely episodic. Erosion and/or fertilization has destroyed the cratonic lithosphere below several cratons (e.g., North China; Wu et al., 2019) and more localised thinning is indicated by recent seismic imaging of the Kaapvaal and Zimbabwe cratons (Celli et al., 2020). In these examples, lithospheric erosion significantly post-dated craton formation and stabilisation.

The most recent episode of thinning of the Kaapvaal lithosphere (between 1.15 Ga and 85 Ma) could be consistent with foundering related to mantle plume

543 interaction over the last 200 Ma (e.g., Ravenna et al., 2018; Celli et al., 2020). The
544 Kaapvaal craton has anomalously high elevation (up to ~1.5 km in the central
545 regions) that appears related to lithospheric thinning (Ravenna et al., 2018).
546 Significant crustal erosion and denudation, which could imply uplift resulting from
547 lithospheric thinning, has occurred since 120 Ma on the Kaapvaal craton, with
548 erosion of ~500 m of the upper crust in the Kimberley region and a further ~850 m
549 post 85 Ma (Hanson et al., 2009). Such significant erosion clearly contrasts with the
550 generally low denudation rates of cratons worldwide (e.g., Blackburn et al., 2012). It
551 also contrasts with evidence that the elevation of the Kaapvaal craton prior to the
552 Cretaceous showed limited emergence above sea-level since the Archaean (Ballard
553 and Pollack, 1988). This is consistent with the modest decrease in LAB depth from
554 the Proterozoic to the Cretaceous inferred from mantle xenoliths. The greater
555 reduction in LAB depth inferred from DI *PT* data suggests that the majority of
556 lithospheric erosion occurred prior to 1.15 Ga. This, coupled to the lack of evidence
557 that the Kaapvaal craton was ever substantially above sea-level prior to the
558 Cretaceous (Ballard and Pollack, 1988), might suggest that most of the thinning of
559 the lithospheric mantle must have occurred gradually rather than in catastrophic
560 events. A possible mechanism could be basal drag. If mantle viscosities in the
561 Archaean were significantly lower as would be expected for a convecting mantle
562 some 100-200°C warmer than today, it is possible that basal drag was less
563 significant earlier in the Archaean, irrespective of whether plate velocities were
564 higher. We suggest as mantle viscosity increased with secular Proterozoic cooling
565 of the convecting mantle, greater erosion of the lithosphere could have occurred
566 then, resulting in the fairly common ~200-250 km thick lithosphere found globally
567 (Hasterok and Chapman, 2011).

Numerical modelling of mantle convection with chemically distinct continental roots also supports a bimodality in their stability at thicknesses of 300-350 km and ~220 km (Doin et al., 1997). Whilst a lithosphere of > 300 km is in excess of the deepest xenoliths from ~240-250 km depth found to date from cratonic lithosphere, the absence of evidence alone does not disprove a thicker early lithosphere (Artemieva, 2011), particularly as kimberlites sampled mantle xenoliths non-randomly and generally from shallower depths than diamonds, at least at Premier (e.g., Fig. 2a and 2c). Most simply though, the rarity of xenoliths from > 200 km depth globally might simply result from the fact that kimberlites have overwhelmingly formed in the Phanerozoic (Tappe et al., 2018) long after most of the basal erosion could have occurred. On this basis, we propose that an original Kaapvaal lithosphere of >300 km thickness is not beyond the realm of possibility. Substantial upward migration of the LAB since craton formation could be an attractive explanation for the recovery of majorite bearing garnet, or exsolutions thereof, in DIs from the Kaapvaal and other cratons. Majorite is not stable within the current cratonic keel (at least in the Kaapvaal) and majoritic garnet DI apparently formed from depths of ~250 km (i.e., in excess of present day LAB estimates) to as deep as the mantle transition zone (Harte, 2010). The Jagersfontein kimberlite also contains rare peridotite xenoliths with omphacitic exsolution lamellae in pyrope garnet indicative of formation at similar depths between 300-400 km depth (Haggerty and Sautter, 1990). It is possible that rather than being sub-lithospheric in origin, majorite bearing xenoliths and DI could be vestiges of the long-removed much deeper portions of the original cratonic lithosphere.

A thicker lithosphere in the Archaean is the most parsimonious explanation for the chemistry of Archaean basaltic lavas that require a high T_p , the position of DI PT

well below the modern cratonic geotherm, and the existence of DI and xenoliths with chemistries that could only have formed below the base of the current lithosphere. In summary, whilst cratons may have attained their maximal vertical extent early in their history, most likely during the Archaean at a time culminating with their greatest mechanical stability, they have been gradually eroded and weakened ever since.

Figure captions

Figure. 1: a) Global heat flux (in TW) at the Earth's surface as a function of the contribution of radioactive decay within the Earth through time as well as the evolution of the T_p of the convecting mantle through time corresponding to a modern-day Urey ratio of 0.38 (Herzberg et al., 2010). b) Internal heat production of the upper and lower crust as a function of radioactive decay through time, adapted from Arevalo et al. (2009). Heat production of upper crust calculated from the map model of Condie (C93; 1993). Heat production of lower crust calculated from the lower crustal xenolith database of Rudnick and Fountain (RF95; 1995), assuming an Archaean intermediate composition. c) Illustration of how a 220°C warmer Moho temperature in the Archaean can generate substantially differing or similar geotherms in the lithosphere at diamond sampling depths dependant on the prevailing basal heat flux (black and grey lines). This situation equates to a 6°C/km decrease in the crustal gradient through time to an assumed present day Moho temperature of 500°C. A present day geotherm (red) of ~190 km thickness with a basal heat flux of 15 mW/m² is shown for comparison, as are the stability fields of graphite and diamond (Day, 2012). Mantle T_p defined at the present day (1350°C) and Archaean (1580°C) from Herzberg et al. (2010). See text for discussion.

Figure 2: Semi-quantitative best-fitting geotherms for mantle xenoliths from Cretaceous (a) and Premier kimberlites (b), compared with clinopyroxene diamond inclusions within Premier kimberlite (c). Xenolith data from Wu and Zhao, (2011) and diamond data from Nimis et al. (2020). Graphite diamond transition after Day (2012).

Figure 3: Residual plots (a-c) for the semi-quantitative best fitting geotherms constructed from xenolith data for close-to-the-present-day across the Kaapvaal craton (~85 Ma) and for the Premier kimberlite (1.15 Ga) (d-e). For discussion of the thermobarometry see the supplementary information.

Figure 4: Modelled geotherms for the Kaapvaal craton accounting for the decrease in heat production within the Earth from 3.0 Ga to the close-to-present-day (85 Ma) under varying conditions. Models a) and b) used a fixed LAB heat flux, models c) and d) used a fixed lithospheric thickness. Models a) and c) have a secularly cooling convecting mantle temperature with decrease in mantle T_p since 3.0 Ga (Herzberg et al., 2010). Models b) and d) have a constant convecting mantle through time demonstrated by a fixed mantle T_p . Panels i, ii, iii and iv display the depth to LAB and graphite-diamond transition (Di-in) as a function of time for geotherm models a, b, c and d, respectively. Note that in none of the models does any secular geotherm approach the position of diamond inclusions from the Kaapvaal craton and in the case of models a) and b) no diamond window exists into the Archaean. Diamond data from Nimis et al. (2020).

Figure 5: Geotherm models at 1.0 Ga and 1.5 Ga taken from Figure 4 compared to PT conditions recorded in mantle xenoliths from the ~1.15 Ga Premier kimberlite. Models a) and b) used a fixed LAB heat flux, models c) and d) calculated with a fixed lithospheric thickness of 195 km obtained from the best fit geotherm model from

mantle xenoliths entrained within Cretaceous kimberlites. Graphite diamond transition after Day (2012). Diamond data from Nimis et al. (2020).

Figure 6: Effect of LAB depth on geotherms. Geotherms in panel a) display the semi-quantitative ‘initial’ geotherms chosen for close-to-the-present-day and at 1.15 Ga, fitted to mantle xenolith *PT* estimates. Geotherms in models a), b) and c) show geotherms from hypothetically thicker lithosphere at 1.15, 1.9 and 3.0 Ga, respectively, that provide a qualitative fit to the coolest *PT* conditions recorded in Premier diamonds. Also shown are a range of geotherms for specific basal heat fluxes (grey). d) displays the best fitting geotherms from models a), b) and c) compared with olivine invariant lines from the popular Al-in-olivine thermometer (Bussweiler et al., 2017), the red shaded region shows where the xenolith geotherm and geotherms fitted to the ‘coolest’ diamond inclusion data diverge significantly. Diamond data from Nimis et al. (2020).

Table Captions

Table. 1: Model results for secular models with fixed LAB heat fluxes or fixed lithospheric thickness both with and without a secularly cooling underlying convecting mantle. Where Q_s is the surface heat flux, H_{uc} and H_{lc} is the internal heat production in the upper and lower crust, respectively. T_p is the mantle potential temperature as demonstrated by a mantle adiabat at specific time intervals (τ) in Ga. LAB = lithosphere-asthenosphere boundary (see text for discussion). Di-in marks the intersection of the geotherm and the diamond stability field in the lithospheric mantle after Day (2013) and window corresponds to the thickness of lithospheric mantle hospitable for diamond stability.

Table 2: Model results for hypothetically thicker lithosphere at differing times in Earth's history. Abbreviations as used in Table. 1.

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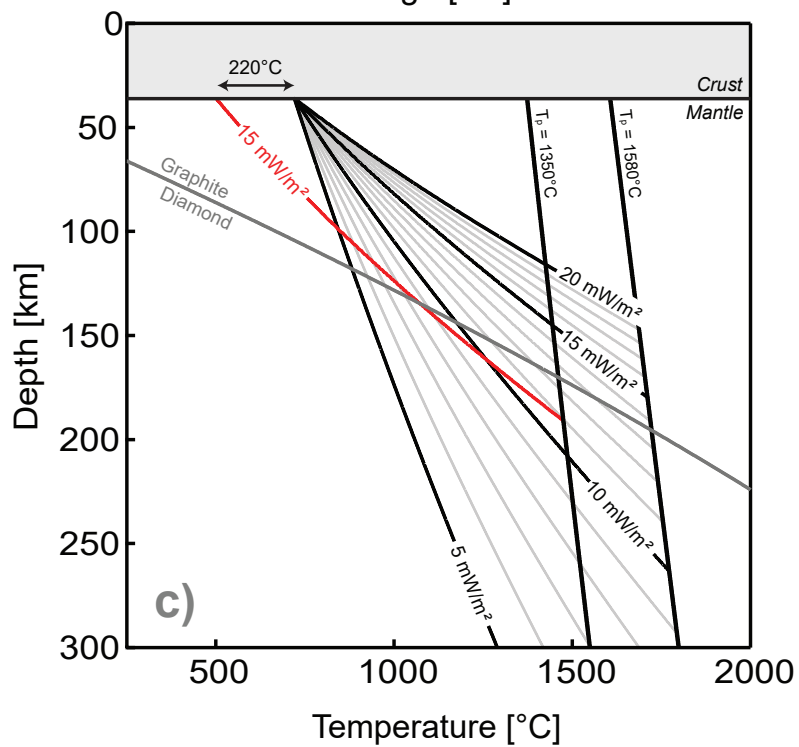
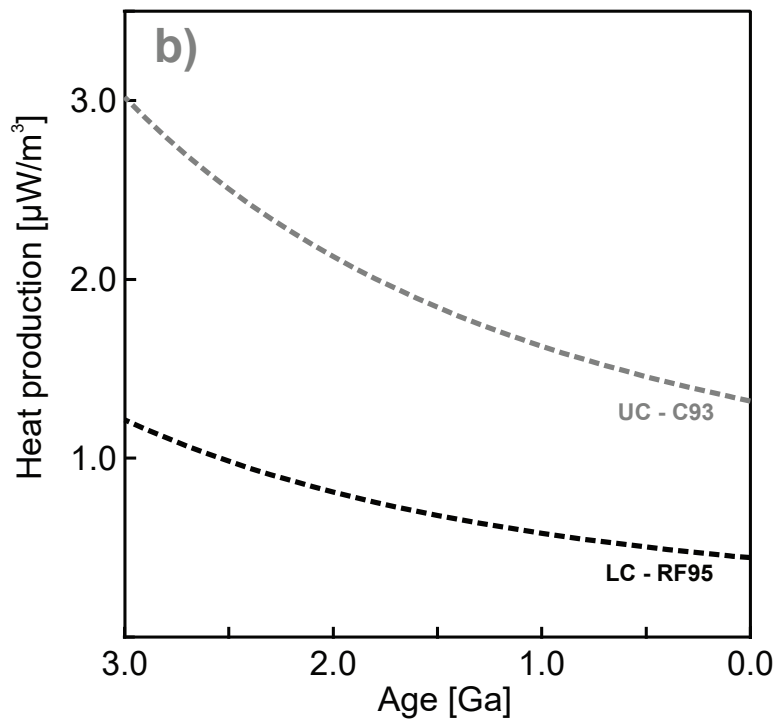
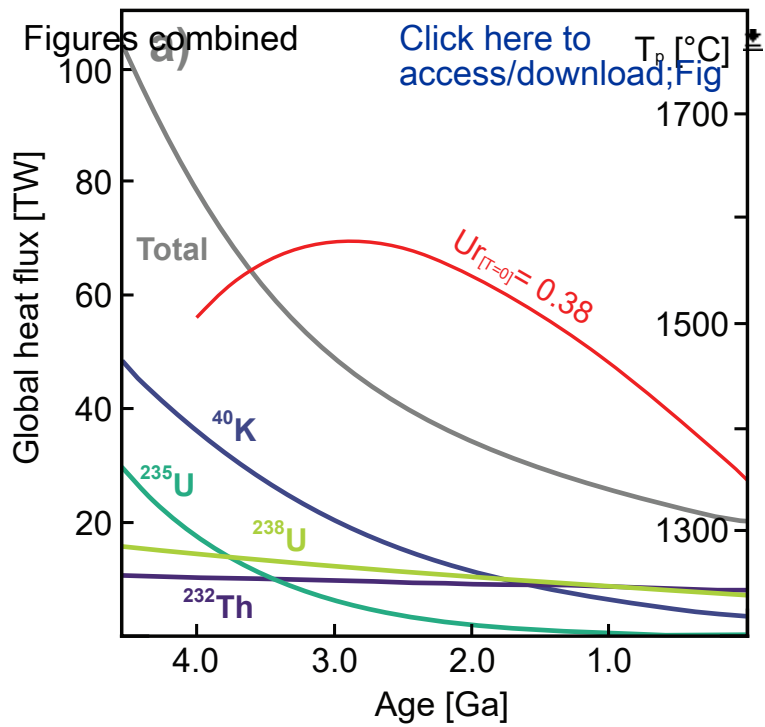
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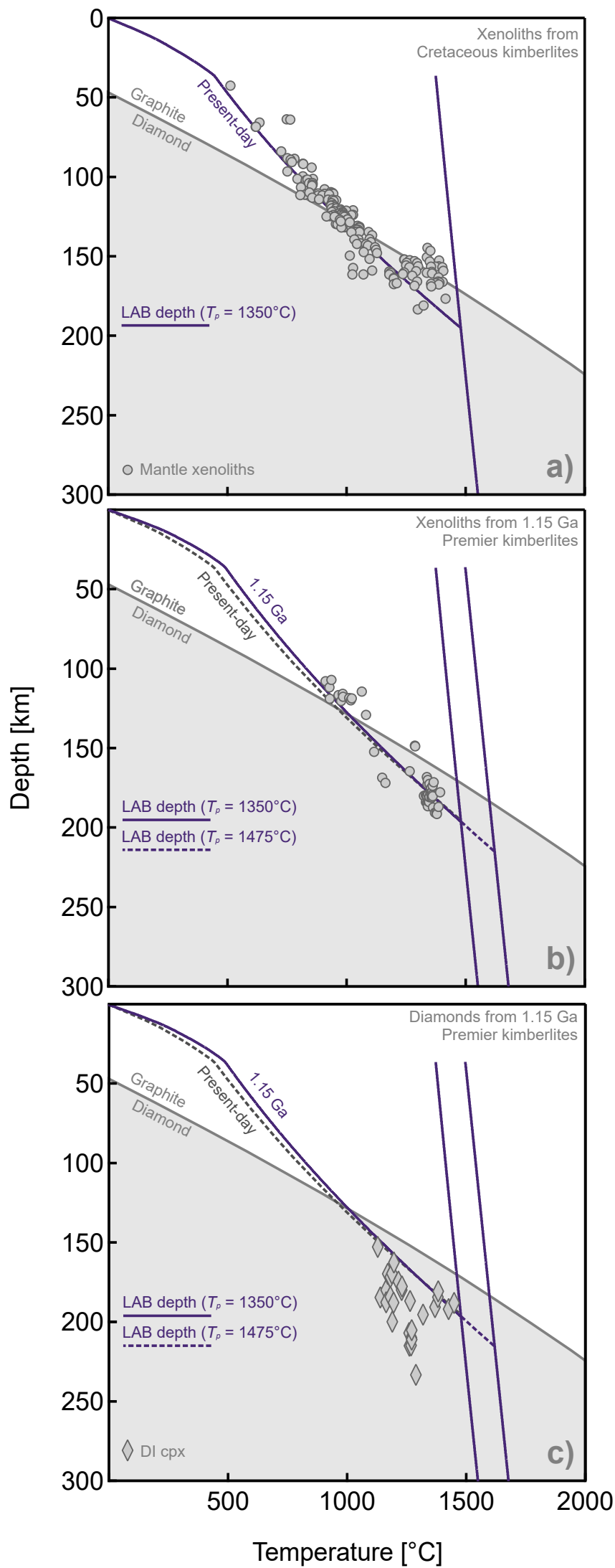
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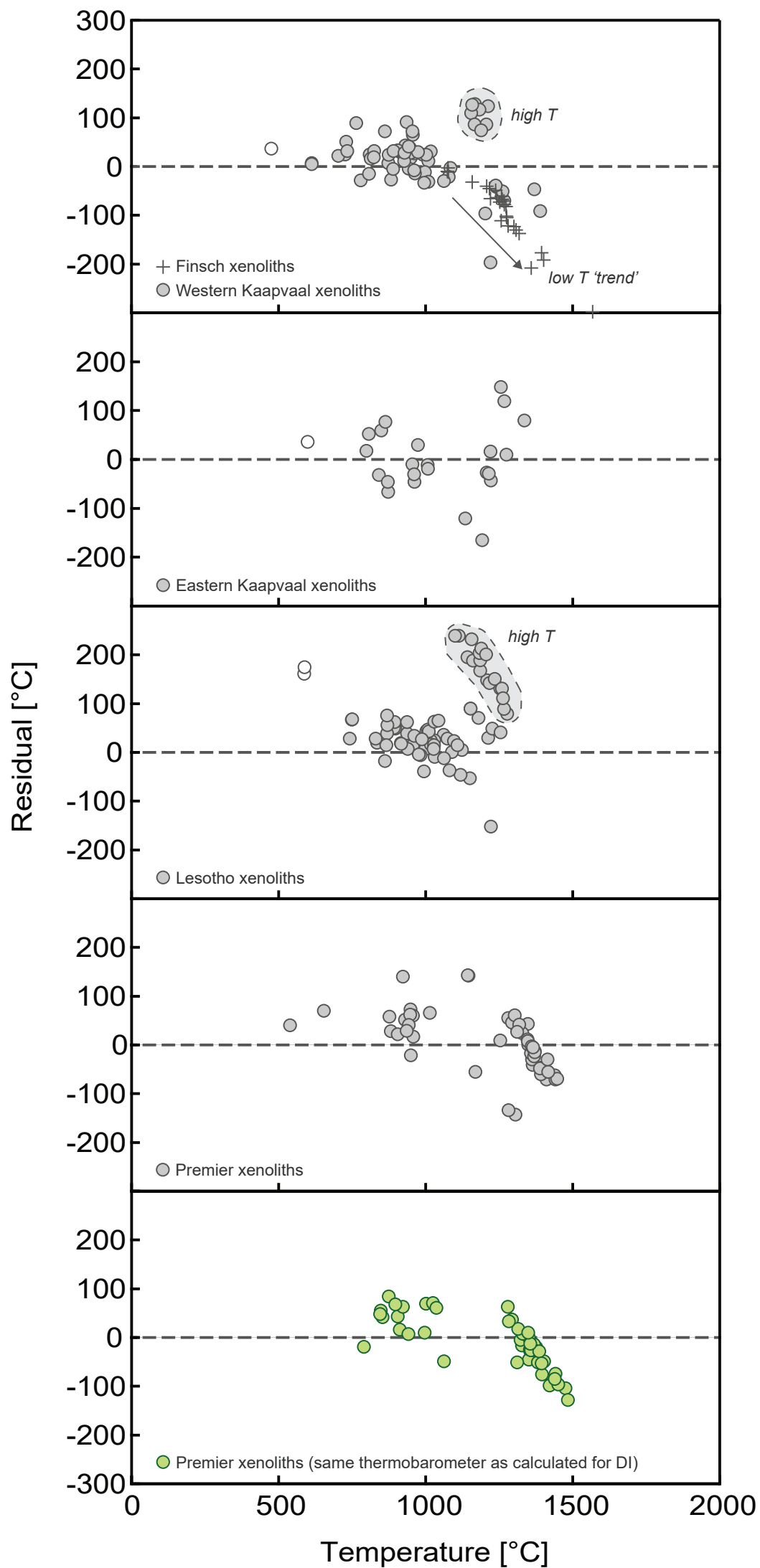
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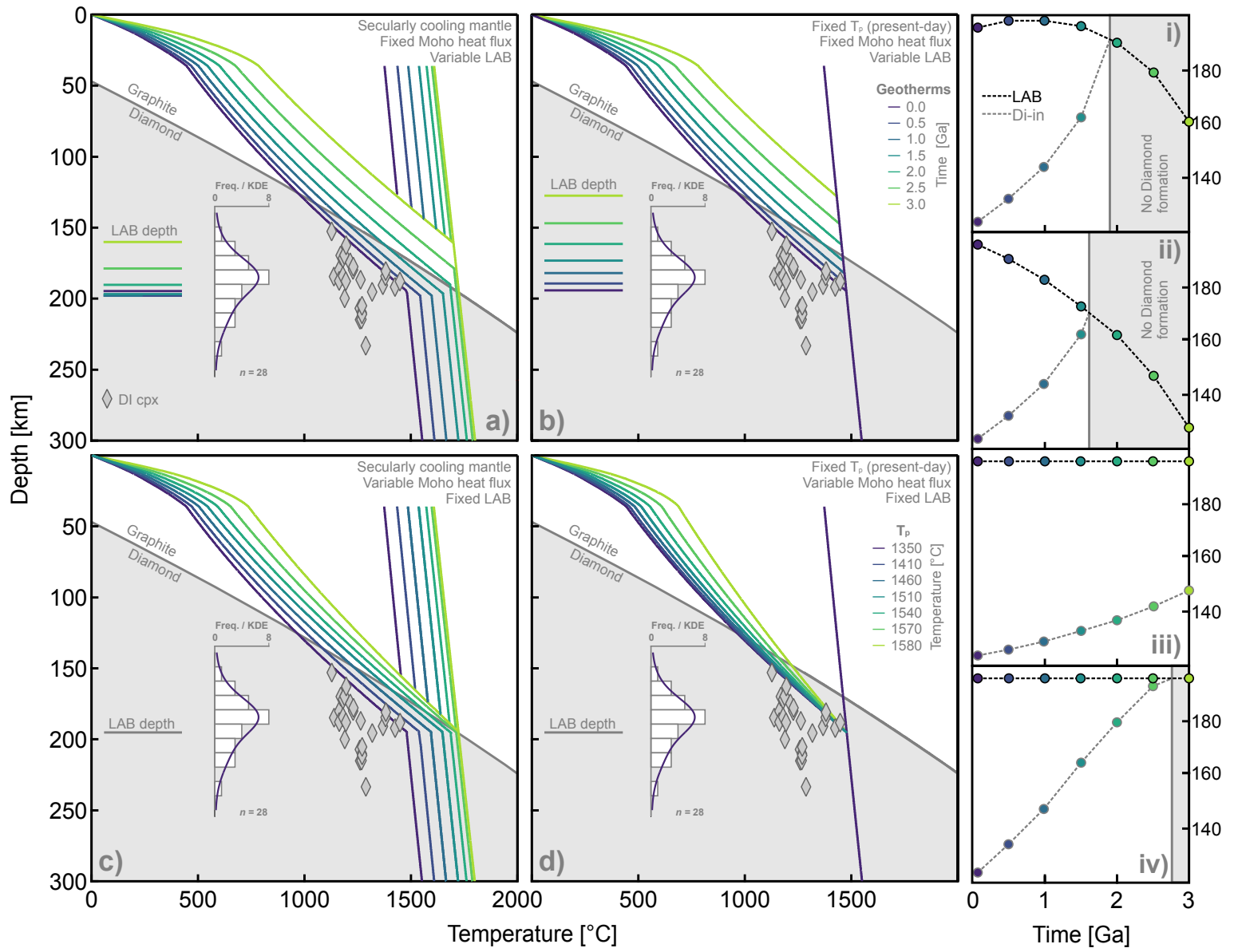
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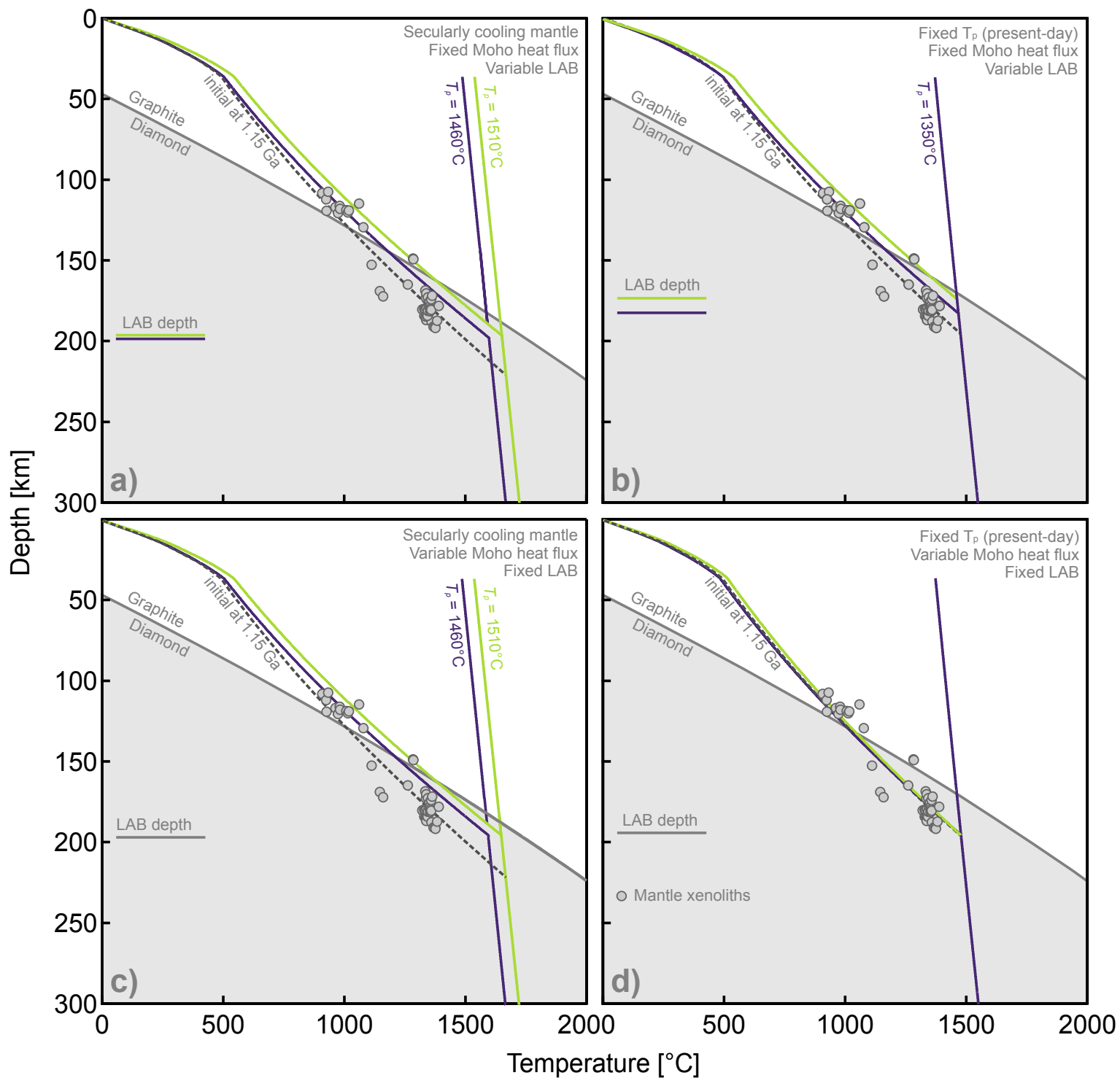
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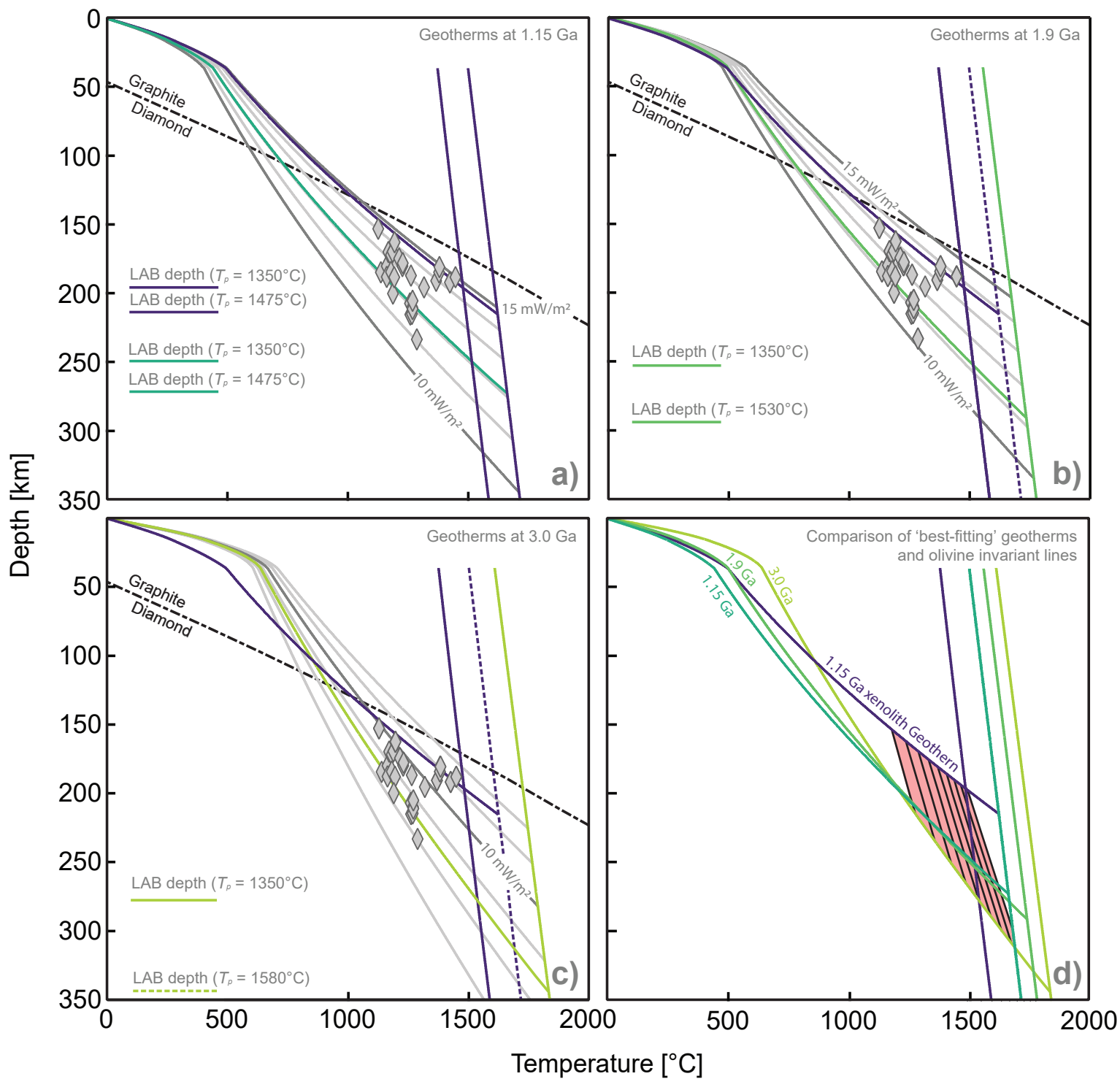


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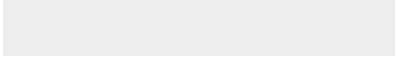


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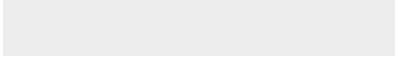


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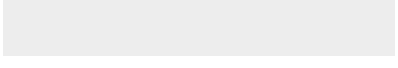


Figure 4 (high-resolution)



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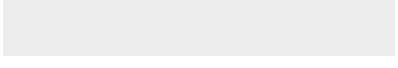


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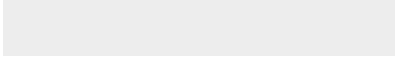


Figure 6 (high-resolution)



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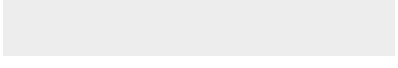


Table 1

Model (Fig. 4)	Time (τ)	Q_s (mW m ⁻²)	$Q_{\text{Moho or LAB}}$ (mW m ⁻²)	H_{uc} ($\mu\text{W m}^{-3}$)	H_{lc} ($\mu\text{W m}^{-3}$)
4a	0.085	47.0	15.8	1.34	0.45
	0.5	50.0	15.8	1.46	0.5
	1	54.4	15.8	1.63	0.58
	1.5	60.0	15.8	1.85	0.68
	2	67.2	15.8	2.13	0.81
	2.5	76.7	15.8	2.5	0.98
	3	90.0	15.8	3.02	1.22
4b	0.085	47.0	15.8	1.34	0.45
	0.5	50.2	16.0	1.46	0.5
	1	54.6	16.0	1.63	0.58
	1.5	60.1	15.9	1.85	0.68
	2	66.9	15.5	2.13	0.81
	2.5	75.6	14.7	2.5	0.98
	3	87.7	13.5	3.02	1.22
4c	0.085	47.0	15.8	1.34	0.45
	0.5	50.0	15.8	1.46	0.5
	1	54.4	15.8	1.63	0.58
	1.5	60.0	15.8	1.85	0.68
	2	67.2	15.8	2.13	0.81
	2.5	76.7	15.8	2.5	0.98
	3	89.6	15.8	3.02	1.22
4d	0.085	47.0	15.8	1.34	0.45
	0.5	49.7	15.5	1.46	0.5
	1	53.6	15.0	1.63	0.58
	1.5	58.6	14.4	1.85	0.68
	2	65.0	13.6	2.13	0.81
	2.5	73.5	12.6	2.5	0.98
	3	85.5	11.2	3.02	1.22

T _p (°C)	LAB (km)	Di-in (km)	Window (km)
1350	195	124	72
1410	198	132	66
1460	198	144	54
1510	196	162	34
1545	190	198	0
1570	179	-	0
1580	161	-	0
1350	195	124	71
1410	195	135	60
1460	195	148	47
1510	195	165	30
1545	195	180	15
1570	195	193	2
1580	195	-	0
1350	195	124	71
1350	190	132	58
1350	182	144	38
1350	173	162	11
1350	162	-	0
1350	147	-	0
1350	128	-	0
1350	195	124	71
1350	195	126	69
1350	195	129	66
1350	195	133	62
1350	195	137	58
1350	195	142	53
1350	195	148	47

Table 2

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Model parameters (variable T_p)	Time (τ = Ga)	Q_s (mW m ⁻²)	$Q_{\text{Moho or LAB}}$ (mW m ⁻²)
Best-fitting Premier emplacement geotherm	1.15	54.9	14.7
	1.15	52.3	12.6
Thicker lithosphere	1.9	60.8	11.2
	3	82.7	8.5
Model parameters (fixed T_p)			
Best-fitting Premier emplacement geotherm	1.15	54.9	14.7
	1.15	52.3	12.0
Thicker lithosphere	1.9	60.8	11.2
	3	82.7	8.5