What controls the thickness of continental crust in the Archean?

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ABSTRACT

Exposed continents are one of Earth's major characteristics. Recent studies on ancient ocean volume and exposed landmasses suggest, however, that the early Earth was possibly a water world where any significant landmass was unlikely to have risen above sea level. In the modern Earth, the thickness of continental crust seems to be controlled by sea level and the buoyancy of continental crust. Simply applying this concept to the Archean would not explain the absence of exposed continents, and we suggest that a third element that is currently insignificant was important during the early Earth: the strength of continental upper crust. Based on pressure imbalance expected at continent-ocean boundaries, we quantify the conditions under which rock strength controls the thickness of continental crust. With the level of radiogenic heat production expected for the early Earth, continents may have been too weak to have maintained their thickness against a deep ocean.

INTRODUCTION

Earth is covered by two different kinds of crust, oceanic and continental crust. The normal oceanic crust is ~7 km thick (White et al., 1992), whereas the continental crust is ~40 km thick on average (Christensen & Mooney, 1995). The thickness of oceanic crust is a simple function of mantle potential temperature beneath mid-ocean ridges (McKenzie & Bickle, 1988); a hotter mantle melts more, resulting in thicker oceanic crust. For the thickness of continental crust, we do not expect such a genetic relationship with the state of the present-day mantle, because the formation and evolution of continental crust are more complex, involving a variety of tectonic processes over a timescale of billions of years. Instead, as suggested by Hess (1962), the long-term thickness of continental crust may be regulated by a combination of isostasy and sea level. Because the continental crust is less dense than the oceanic crust, isostasy allows the former to be thicker than the latter. As the continental crust thickens, its surface eventually exceeds the sea level, above which erosion becomes important. In the hypothesis of Hess (1962), therefore, the thickness of continental crust is controlled by the ocean volume; a deeper ocean allow the existence of thicker continental crust. This effect of ocean on continental height is well understood in geology. The mean height of continents with respect to the sea level, known as the continental freeboard (Wise, 1974), has been close to zero at least back to two billion years ago (Korenaga et al., 2017).

Recent geochemical studies exploring the emergence of continents (Bindeman et al., 2018; Johnson & Wing, 2020) have raised the possibility that most of the Archean was a water world. A study on deep water cycle has also suggested that the Archean ocean could have been more voluminous, with its surface reaching 4-6 km above mid-ocean ridges (Korenaga et al., 2017). If the thickness of continents is regulated only by isostasy and sea level, however, exposed continental crust seems inevitable even with a deep ocean. This is especially so because the mass of continents is likely to have reached the present-day level in the early Earth (Korenaga, 2018). Even with more gradual continental growth, continents can still grow to sea level through thickening by orogeny.

In this study, we investigate the possibility that, in addition to sea level and isostasy, the thickness of continental crust is also regulated by the strength of crustal rocks. Although it has already been suggested that the Archean crust is too weak to have supported high mountains (Rey & Coltice, 2008), previous studies have been limited to topographic variations *within* continents. Here we consider the integrity of continents at a greater spatial scale, that is, with respect to adjacent ocean basins. Our results suggest that, whereas strength does not seem to be a limiting factor today, ductile flow could have been more important in the Archean.

In what follows, we first estimate differential stresses applied on the continental crust by performing isostatic calculations. We also estimate temperature in the upper crust by calculating the geotherm for both the modern and Archean Earth. As the concentration of heat-producing elements in the Archean continental crust is still debated, we compare three cases: no secular change in crustal composition (Guo & Korenaga, 2020), 75% less heat production (Condie, 1993), and 50% less heat production (Ptáček et al., 2020). From these stress and temperature conditions, we evaluate the representative deformation rate of continental upper crust at the present and in the Archean. Finally, we discuss how the Archean continental height might have been controlled and what this implies for the early Earth landscape.

MODEL SETTING

The topography of continental crust and its overall thickness are controlled by different mechanisms. The former is supported by variations in crustal thickness, for which the strength of lower crust is important (e.g., Rey & Coltice, 2008). Higher radiogenic heat production and higher mantle temperatures in the Archean could have weakened and even melted the lower crust (Galer & Mezger, 1998), thereby limiting the surface relief. We expect a different mechanism to control the average thickness of continents (cf. England & Molnar, 1997). At the boundary of continental and oceanic domains, the horizontal flow of continental lower crust would be limited by the surrounding oceanic lithosphere, leaving the continental upper crust to control the average crustal thickness (Fig. 1). As the top portion of continental crust is bounded only by ocean, its ability to support its own weight should determine the continental height. The strength of continental upper crust is also better constrained by geological observations. The strength of continental lower crust depends on its water content, lithology, and the depth distribution of radiogenic heating (Burov & Diament, 1995), all of which are poorly constrained in deep time. In contrast, the silica content of the continental upper crust is likely to have stayed relatively constant through Earth history (Keller & Harrison, 2020). Crustal thickness may also be regulated by delamination as thickened crust could stabilize garnet in the lower crust, but delamination is not dynamically feasible under most conditions (Mondal & Korenaga, 2018).

In our model, pressure difference at the continent-ocean interface drives crustal flow, the rate of which is controlled by the rheology of continental upper crust. A temperature profile through the continental crust is calculated with the secular evolution of radiogenic heat production and mantle potential temperature (see Supplementary Materials (SM) for details). The flow law for the upper crust is based on our reanalysis of published deformation experiments on quartz aggregates. When discussing the strength of crust or mantle, it is customary to calculate yield stress for an assumed strain rate. In our model setting, however, stress driving deformation is predetermined by the aforementioned pressure difference, so instead of yield stress, we calculate strain rates corresponding to the pressure difference.

Pressure Difference Between Continental and Oceanic Domains

As the continental upper crust is denser than the surrounding seawater, the pressure difference at the continent-ocean boundary drives the horizontal flow of continental crust into the ocean (Fig. 1); this pressure difference exists at both passive and active margins. As the oceanic lithosphere is denser than the continental crust, the pressure difference is reversed at greater depths, and the continental lower crust is supported by the surrounding oceanic lithosphere. Therefore, the overall thickness of continental crust is contingent on the free-flowing continental upper crust. In our model, the compensation depth is taken to be at the base of continental lithospheric mantle (Fig. 1), and the corresponding isostatic balance is given by:

$$h_{\mathrm{cc}}\rho_{\mathrm{cc}} + h_{\mathrm{cm}}\rho_{\mathrm{cm}} = h_{\mathrm{w}}\rho_{\mathrm{w}} + h_{\mathrm{oc}}\rho_{\mathrm{oc}} + h_{\mathrm{om}}\rho_{\mathrm{om}} + h_{\mathrm{m}}\rho_{\mathrm{m}}, \quad (1)$$

where h and ρ denote thickness and density, respectively, and subscripts "cc", "cm", "w", "oc", "om", and "m" refer to continental crust, continental lithospheric mantle, water, oceanic crust, oceanic lithospheric mantle, and asthenospheric mantle, respectively. Here the continental structure is compared with the oceanic structure at the mid-ocean ridge, to provide a minimum ocean depth at the continent-ocean boundary. More realistic cases with older seafloor are considered in the discussion section. Our isostasy model is built on the continental freeboard model of Korenaga et al. (2017), in which the thicknesses and densities of different layers and their temporal evolution are estimated from a range of geological and geophysical observations as well as theoretical considerations (see SM for a summary of key model parameters). In this study, we vary the thickness of continental crust as a free parameter, and to provide a conservative estimate on pressure difference, the top of continental crust is assumed to be at or below sea level (Fig. 1).

Strength of Upper Continental Crust

The rheology of continental upper crust is usually considered to be controlled by that of wet quartz aggregates (Kohlstedt et al., 1995; Bürgmann & Dresen, 2008). However, several different flow laws have been published (e.g., Luan & Paterson, 1992; Gleason & Tullis, 1995; Rutter & Brodie, 2004a,b; Fukuda et al., 2018), and predicted crustal strength varies substantially with the choice of quartz flow law. Because crustal strength plays a central role in our analysis, we have critically examined major deformation data on quartz aggregates, using the nonlinear inversion method developed by Korenaga & Karato (2008) and amended by Mullet et al. (2015) (see SM for details). We focus on ductile deformation because the available pressure difference at the continent-ocean boundary does not exceed the brittle strength of rocks, assuming a friction coefficient of 0.8 (Byerlee, 1978).

Our reanalysis shows that, as opposed to what Fukuda et al. (2018) suggested, there is little experimental support for grain boundary sliding being the dominant deformation mechanism of quartz aggregates. A combination of diffusion and dislocation creep is shown to be sufficient to explain the published quartz deformation data, with dislocation creep being most relevant under geological conditions. Our reanalysis further suggests that the deformation data of Gleason &Tullis (1995) provide statistically the most reliable flow law. In this study, therefore, we use the flow law derived from their data using our inversion.

RESULTS

Strain-rate profiles expected at the edges of continental crust were calculated for crustal thicknesses between 35-60 km under present-day and Archean conditions. Representative calculations are shown in Fig. 2 for a thickness of 40 km. The present-day strain rate does not exceed the geological deformation rate of 10^{-15} s^{-1} (corresponding to the deformation time scale of ~30 Myr for 100% strain) even at greater depths. In contrast, the Archean crust reaches the geologic deformation rate at relatively shallow depths, suggesting that the continental upper crust was capable of significant flow in the Archean, possibly limiting the continental height.

We focus on the strain rate at the base of the ocean (Fig. 3), because it is the highest strain rate achievable in the upper crust adjacent to the ocean. The strain rate continues to increase below the base of the ocean, but it is unlikely to be realized because the crust is bounded by oceanic lithosphere at those depths. The strain rate calculated under present-day conditions is much smaller than the geological strain rate of 10^{-15} s^{-1} for an almost entire range of crustal thickness considered. For the early Archean, on the other hand, the relative buoyancy between the continental and oceanic domains would place mid-ocean ridges at ~6 km below sea level (Korenaga et al., 2017; Rosas & Korenaga, 2021), for which the continental crust had to be >50 km thick to reach sea level (Fig. 3). Such crustal thickness is sufficient to produce a geologically significant strain rate. This secular change in crustal strength arises partly because the crustal geotherm was hotter in the past due to higher radiogenic heat production and higher mantle temperature (Fig. 2) and partly because a slightly deeper part of the upper crust becomes relevant for a deeper ocean in the past (Fig. 3).

DISCUSSION

So far, we focused on crustal strain rates expected at the depth of mid-ocean ridges. However, the depth of ocean basins at the continent-ocean interface can be greater because of seafloor subsidence. Thus, at realistic continent-ocean boundaries where old seafloor is adjacent to continents, we can expect higher crustal strain rates because strain rates generally increase with depth (Fig. 2). For comparison, strain rates expected in the case of a seafloor subsidence of 3 km, which is observed for present-day seafloor older than ~100 Ma, is also shown in Fig. 3 (dashed lines). Even with this consideration, crustal strain rates are much lower than the geological deformation rate (10^{-15} s^{-1}) under present-day conditions. Thus, it is reasonable to conclude that crustal strength does not play any role in controlling continental thickness at present. Under Archean conditions, on the other hand, the geological deformation rate is reached when crustal thickness is greater than ~45 km, and this suggests a difficulty of growing continents to reach sea level when the oceans is deeper than ~5 km at mid-ocean ridges (Fig. 3). The stability of the Archean continental crust can be increased by lowering crustal heat production, and with only 50% heat production, the crust may be able to grow up to ~55 km (Fig. 3), allowing it to emerge above sea level even when the mid-ocean ridge is ~6 km deep.

A real continent-ocean boundary would not be as sharp as depicted in Fig. 1, and a more gradual transition reduces pressure difference. With pressure difference halved, for example, the predicted strain rate would be lowered by around one order of magnitude given the stress dependence of crustal rheology. This difference is smaller than the uncertainties associated with flow-law predictions, seafloor subsidence, or crustal heat production (Fig. 3). Another realistic complication not captured by our model is the effect of a cold skin layer of the continental crust adjacent to the ocean. Such a cold layer would deform in a brittle regime, with a yield strength being ~56% of lithostatic pressure, assuming a friction coefficient of 0.8 and hydrostatic pore fluid pressure. As our model geometry conforms to the homogeneous stress condition, for which the Reuss average is appropriate (Karato, 2008), upper-crustal flow would still be controlled primarily by the ductile strength of a weaker interior as long as the cold skin layer is relatively thin (see SM).

When we aim to understand the early Earth landscape, the importance of a theoretical approach becomes apparent because relevant geological data are scarce. Given the available experimental constraints on the strength of continental upper crust, it seems difficult for continents to emerge from a deep ocean expected for the early Archean, unless the crust is depleted in heat-producing elements. At the same time, our results also suggest that continental crust could easily be thickened up to ~40 km even in the Archean (Fig. 3), so exposed continents are possible if the ocean was not so deep. Because the history of ocean volume is unlikely to be monotonic (Miyazaki & Korenaga, 2022), the possibility of exposed land, which plays an essential role in some major hypotheses for the origin of life (e.g., Deamer, 2019), could depend critically on a subtle balance between the growth of ocean and continents (Korenaga, 2021). Given an ocean depth, our model can predict the maximum crustal thickness, and this capability will be vital in estimating the extent of exposed land in the early Earth.

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FIGURE CAPTIONS

Figure 1. Schematic drawing for compositional stratifications in the top few hundred kilometers of Earth: (A) continental domain and (B) oceanic domain. Different layers have different thicknesses and densities as indicated. These layers are drawn not to scale. Dotted lines indicate expected crustal flow based on horizontal pressure imbalance and rock rheology. Dashed rectangle denotes the main focus of our modeling (Fig. 2).

Figure 2. Modeling strain-rate profiles for present-day and Archean crust with a thickness of 40 km. (A) and (B) show predicted pressure differences as a function of depth for present-day and Archean Earth, respectively. (C) shows present-day (blue) and Archean (red) geotherms, and (D) shows the corresponding strain-rate profile (shading indicates interquartile range corresponding to the uncertainty of quartz flow law). For comparison, the Archean geotherms with 75%

(orange) and 50% (green) heat production are shown in (C), and Archean strain-rate predictions according to the original flow law of Gleason and Tullis (1995) (black dashed) and the field-based flow law of Hirth et al. (2001) (black dotted; see SM for its uncertainty) are shown in (D).

Figure 3. (top) Depth of mid-ocean ridges (h_w) corresponding to continental thicknesses, according to the freeboard model of Korenaga et al. (2017). (bottom) Crustal strain rate, as a function of continental thicknesses, at the base of the ocean (e.g., at ~2.5 km depth at present in Fig. 2D): present-day (blue) and Archean (red), with shading for interquartile range. Median strain rates at depth of 3 km below the ridge depth are shown in dashed. Also shown are two more Archean predictions, at depth of 3 km below ridge depth, with 75% (red dot-dashed) and 50% (red dotted) heat production.

[1]GSA Data Repository item 201Xxxx, Reanalysis of the deformation data of quartz aggregates and the details of strain rate modeling, is available online at www.geosociety.org/pubs/ft20XX.htm, or on request from editing@geosociety.org.

Figure 1

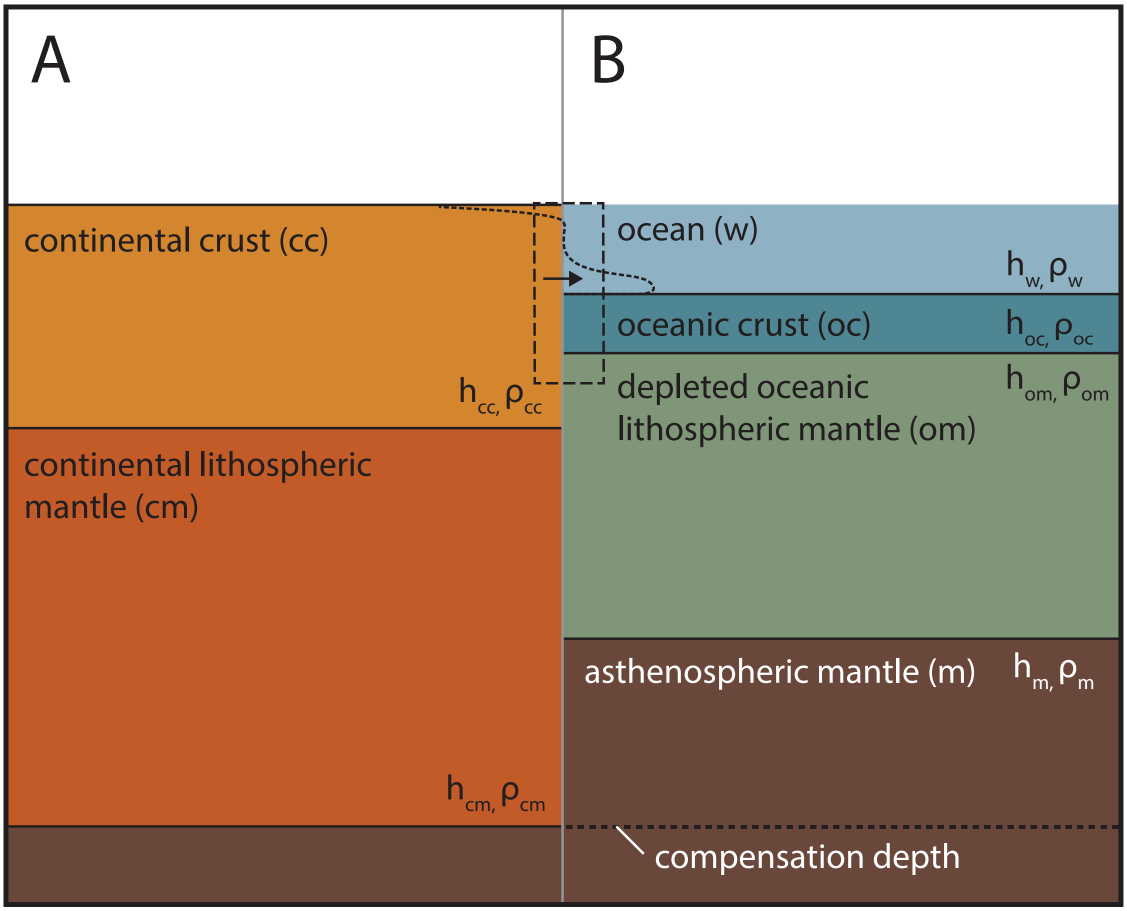

A
B
continental crust (cc)
h_{cc}, ρ_{cc}
continental lithospheric mantle (cm)
h_{cm}, ρ_{cm}
ocean (w)
h_{w}, ρ_{w}
oceanic crust (oc)
h_{oc}, ρ_{oc}
depleted oceanic lithospheric mantle (om)
h_{om}, ρ_{om}
asthenospheric mantle (m)
h_{m}, ρ_{m}
compensation depth

Figure 2

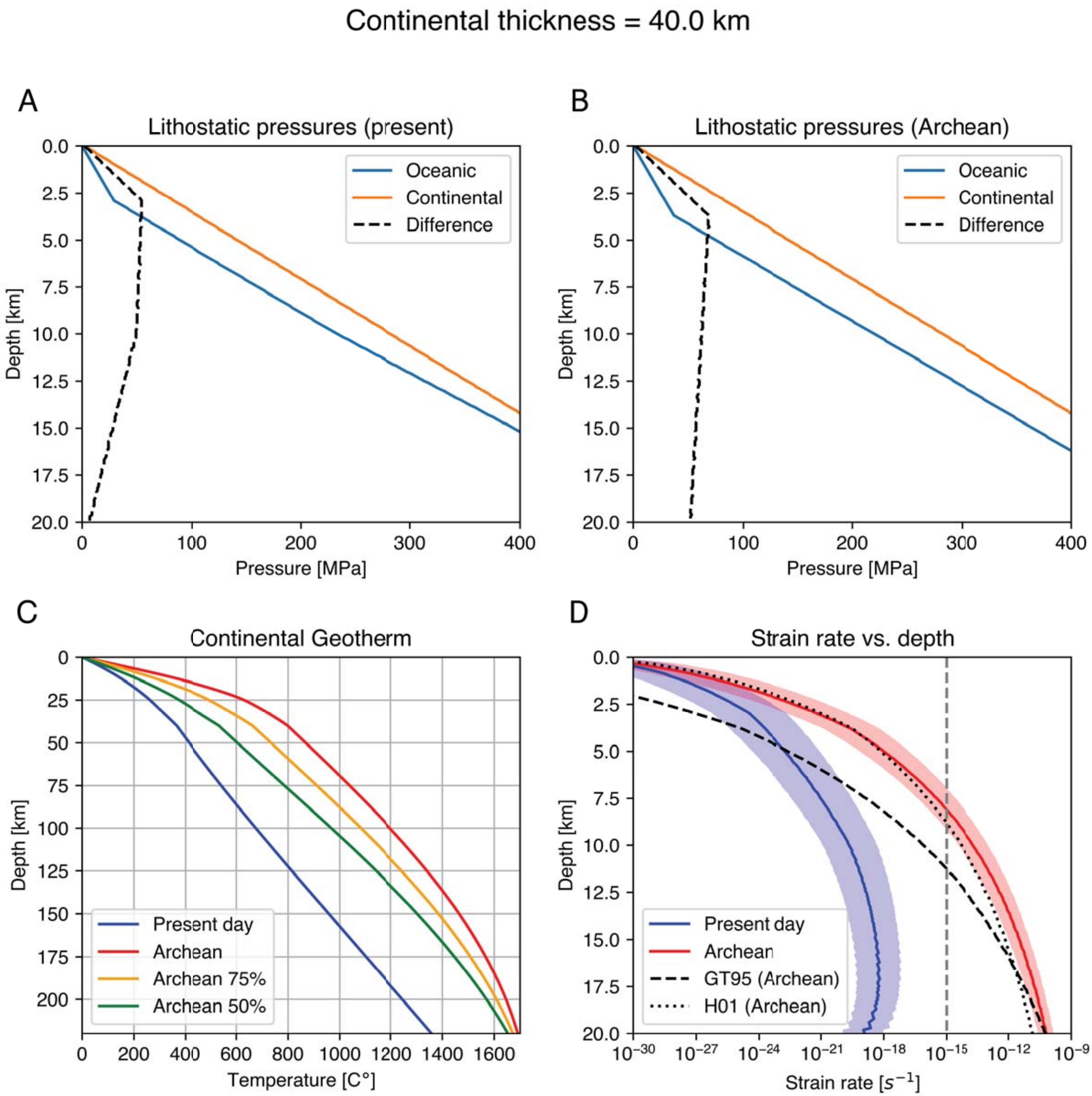
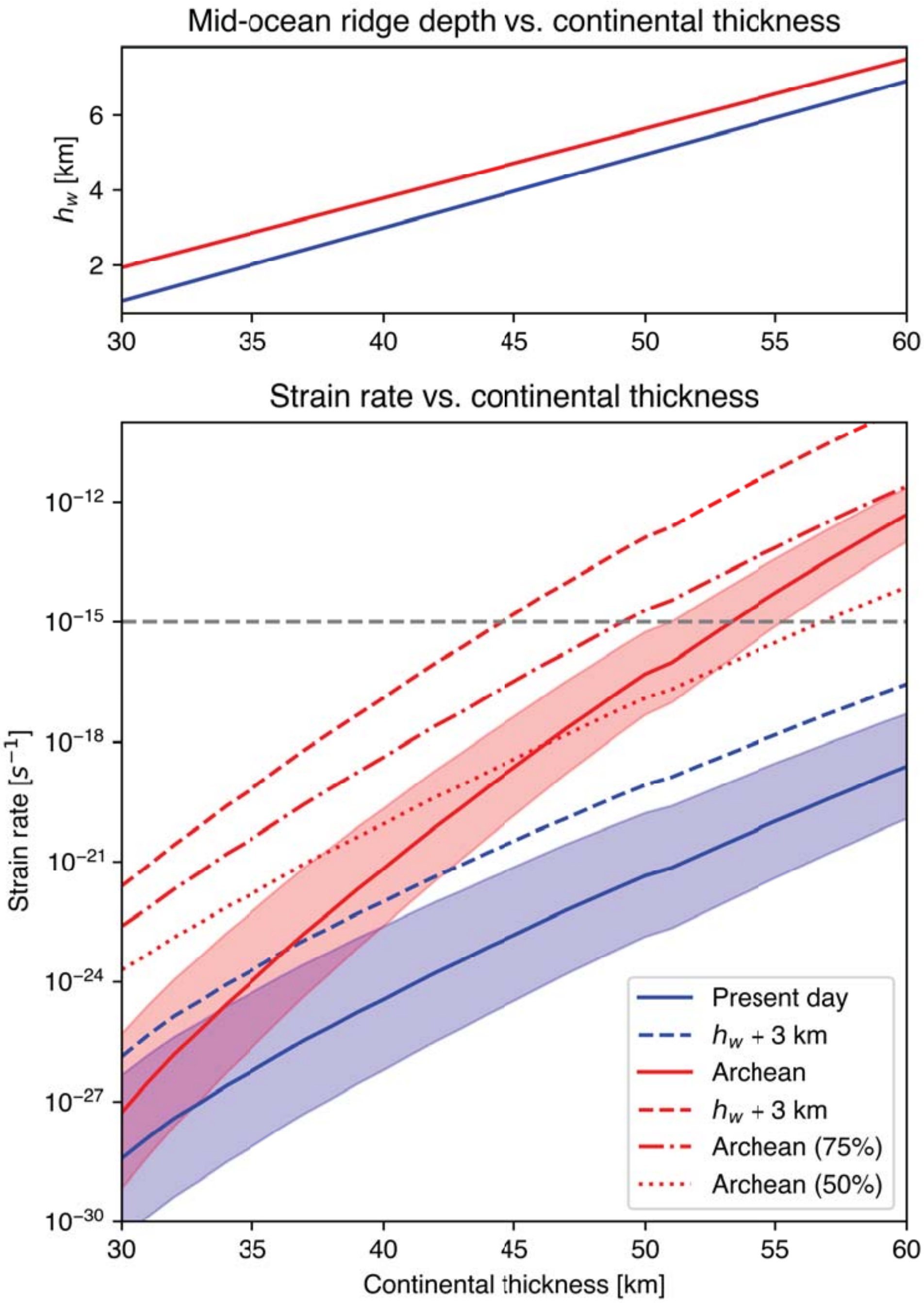

Continental thickness = 40.0 km
A
Lithostatic pressures (present)
Oceanic
Continental
Difference
Depth [km]
Pressure [MPa]
B
Lithostatic pressures (Archean)
Oceanic
Continental
Difference
Depth [km]
Pressure [MPa]
C
Continental Geotherm
Present day
Archean
Archean 75%
Archean 50%
Depth [km]
Temperature [C°]
D
Strain rate vs. depth
Present day
Archean
GT95 (Archean)
H01 (Archean)
Depth [km]
Strain rate [s^{-1}]

Figure 3



Mid-ocean ridge depth vs. continental thickness
h_w [km]
2
4
6
30
35
40
45
50
55
60
Strain rate vs. continental thickness
Strain rate [s^{-1}]
10^{-12}
10^{-15}
10^{-18}
10^{-21}
10^{-24}
10^{-27}
10^{-30}
Present day
h_w + 3 km
Archean
h_w + 3 km
Archean (75%)
Archean (50%)
30
35
40
45
50
55
60
Continental thickness [km]