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Modern-style continental tectonics since the early Archean

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ABSTRACT

We infer the nature of continental tectonics in the Precambrian by combining our understanding of the principles that govern the present-day behaviour of the continents, which over large parts of the Earth's surface does not resemble plate tectonics, with the available observations from the Precambrian rock record. We show that the hotter mantle potential temperature had little effect on continental rheology and behaviour. In contrast, higher rates of radiogenic heating played a key role in crustal melting and the subsequent rheological evolution of the continents, although feedbacks between heating, melting and dehydration would have resulted in Precambrian continental rheology and deformation rates similar to the present day. By at least the early Archean, and possibly earlier, deformation fabrics preserved in large swathes of crust and the pressure- temperature histories of metamorphic rocks imply mountain-building by shearing and thrust-stacking above rigid underthrusting foreland. This style of mountain-building requires the presence of rigid crustal blocks, which are the anhydrous residue of prior melt extraction following thickening of the crust and lithospheric mantle. The rheology of the forelands of mountain belts controls the crustal thickness contrasts that can be supported, and along with the rate of radiogenic heating that crustal thickness controls the temperatures attained. The interplay between decreasing radiogenic heating over Earth history and the development of strong crustal blocks that can support thick mountain ranges leads to optimal conditions for the generation of tonalite–trondhjemite–granodiorite (TTG) melts during hot Archean continental mountain building events, explaining both their Archean prevalence and subsequent decline. However, the overall style of continental tectonics has remained unchanged since at least the early Archean, and TTG petrogenesis simply represents this style of tectonics superimposed upon a higher rate of radiogenic heating. We suggest that this intrinsically modern-style deformation of the continents was preceded by a phase of (early?) Hadean tectonics resembling the deformation of skin on custard. Hydrous fault zones are too weak to form a 'stagnant lid' on Earth-like planets, and density-driven vertical motions (e.g. 'sagduction') occur at insignificant rates, further emphasising the uniformitarian nature of continental tectonics from the early Archean onwards.

1. Introduction

In this paper we examine the tectonics of the continents in the Precambrian, with a particular focus on the Archean. As such, our focus is different from a number of recent reviews of Precambrian tectonics that have specifically addressed the timing of onset of plate tectonics (e.g. Korenaga, 2013; Brown and Johnson, 2018; Cawood et al., 2018; Hawkesworth et al., 2020; Palin et al., 2020; Palin and Santosh, 2021), which, as we discuss below, is a distinct question from the nature and evolution of continental tectonics. We specifically focus on the continents for two reasons. The first is that the vast majority of oceanic lithosphere is doomed to be lost from the geological record by recycling into the mantle, with only patchy fragments remaining near the surface as ophiolites (e.g. Furnes and Dilek, 2022). In contrast, the buoyancy of the continents presents us with a much more voluminous record to examine. The second reason for focusing on the continents is that the past two decades have seen dramatic progress in our understanding of the structure, material properties, and behaviour of the continental lithosphere at the present day, as described below. These advances have paved the way to re-examine Precambrian continental tectonics through the lens of these new insights, which is the purpose of this paper. Our approach does not mean we are assuming that ancient continental tectonics resembles that at the present day; we are instead relying on the underlying physical and chemical principles being unchanged, and asking how these principles would have interacted with the characteristics of the Precambrian Earth that were different from the present day

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(e.g. the rates of radiogenic heating, Jaupart et al., 2016, and the temperature of the convecting mantle, Herzberg et al., 2010). We take a mostly geophysical approach, combined with some thermodynamic modelling, and as such this paper complements recently published investigations of continental deformation that have more of a structural, geochemical, or petrological focus (e.g. Windley et al., 2021; Cawood et al., 2022). As with recent reviews of present-day faulting and continental tectonics (e.g. Copley, 2018; Jackson et al., 2021), our approach will be based upon combining observations with conceptual and numerical models that contain small numbers of free parameters. This approach is more computationally straightforward than some of the sophisticated numerical schemes that have been developed to study the geodynamics of the present day and the early Earth. The benefits of this approach include the tractability and stability of the necessary calculations, and the ability to directly relate model behaviour to the motivating observations and the values of small numbers of model parameter combinations. However, the resulting models do not produce simulations that recreate the small-scale details of observations.

Because of our focus on the continents, we are therefore not going to address the question of when plate tectonics 'began'. In this paper we use the term 'plate tectonics' in the sense that it was defined (McKenzie and Parker, 1967; Le Pichon, 1968): as a purely kinematic concept, which does not concern the forces acting on the plates. The theory of plate tectonics states that the motions of large parts of the Earth's surface can be described by rigid-body rotations about axes that pass through the centre of the planet (McKenzie and Parker, 1967; Le Pichon, 1968). This definition immediately highlights the difficulty in studying plate tectonics in the distant past, beyond the $<$ 200 Myr record of oceanic magnetic anomalies. The rigid motion of a plate will leave little geological signature except in the comparatively small deformation belts, syn- to post-tectonic sedimentary accumulations, and magmatic arcs along its edges (Fig. 1). Additionally, as we discuss below, much of present-day and ancient continental deformation does not conform to the geometrical rules of plate tectonics, being widely distributed and not explainable as the motions of large rigid blocks about poles of rotation (Fig. 1, and as was recognised in the immediate aftermath of the application of plate tectonics to the oceans; e.g. McKenzie, 1972; Molnar and Tapponnier, 1975). Therefore, the study of the record of continental deformation has little to say about the onset of plate-like behaviour in a global sense, although the presence of local rigid blocks can be inferred on a more regional scale.

We will begin by describing the current understanding of the principles that govern continental deformation at the present day, before providing an overview of the conceptual models that have been previously proposed for continental deformation in the Precambrian. We then use the available observations and models to infer the likely nature of continental tectonics throughout the Precambrian, and discuss the current ambiguities and topics for future study.

Throughout this paper we use the thermal definition of the lithosphere, being the part of the Earth that transfers heat dominantly by conduction, rather than convection (e.g. Parsons and Sclater, 1978; McKenzie, 2006). The distinction between continental and oceanic lithosphere becomes ambiguous in the distant past, where the nature of 'oceanic-style' lithosphere is uncertain (e.g. Cawood et al., 2022), although some fragments of preserved Eoarchean stratigraphy have similarities with some modern oceanic environments (e.g. Nutman et al., 2017). For the purposes of this paper, we define any lithosphere as continental that has an average density that is low enough to prevent widespread recycling of that material into the mantle by it sinking under the influence of gravity, by either modern-style subduction or any other density-driven process. This definition therefore includes regions of thick mafic crust that have a density too low to be recycled into the mantle.

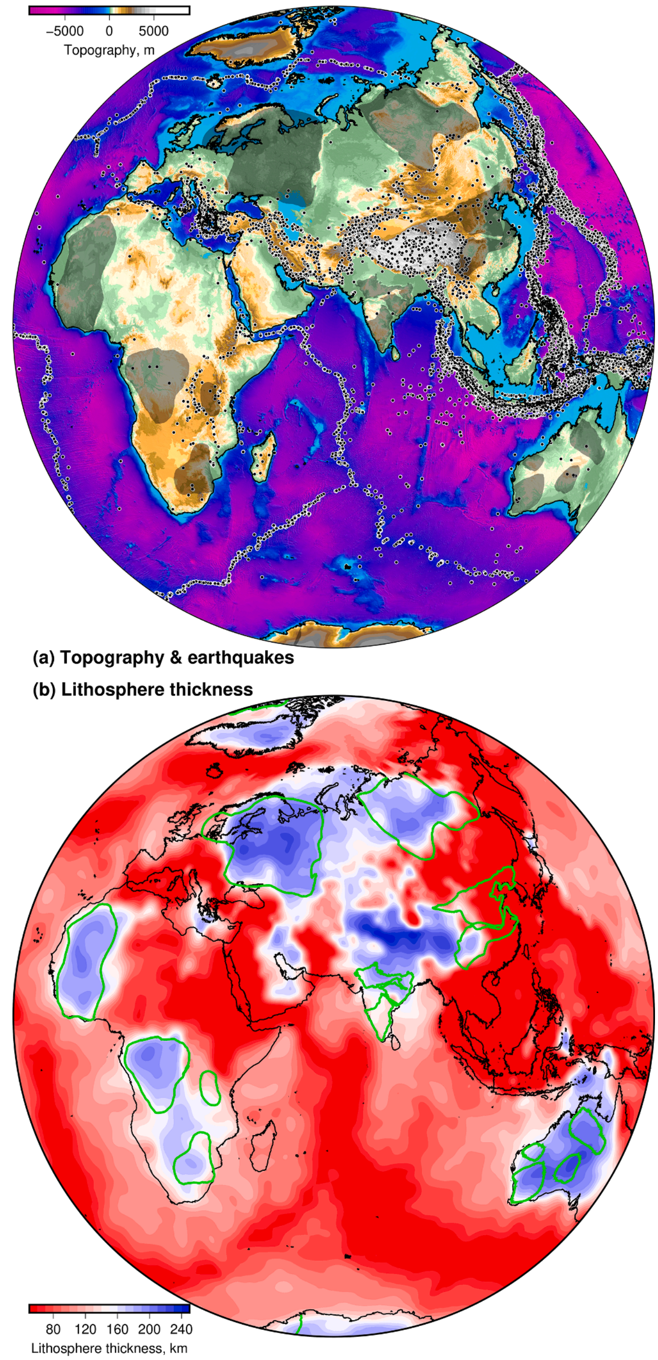

Fig. 1. (a) Topography, earthquakes (dots; magnitude $\geq$ 5.5 in the ISC-EHB bulletin from 1964 to 2020; Engdahl et al., 1998; International Seismological Centre, 2023), and locations of Archean cratons (shaded). (b) Lithosphere thickness from Priestley et al. (2018), overlain with craton outlines (green lines). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

2. Underlying principles revealed by active tectonics

2.1. Rheology, kinematics, and dynamics

Given our approach outlined above, we first present a summary of the current state of knowledge of the active tectonics of the continents. A thorough discussion of this issue is provided in Appendix A, with the aim of providing an overview and summary of relevant recent literature to aid others in relating the geological record to active deformation. We here summarise the main principles that will be used in the remainder of this paper, and note that more explanation can be found in Appendix A.

Fig. 2 illustrates the main forces acting on the lithosphere at the present day (e.g. Forsyth and Uyeda, 1975). 'Slab pull' and 'ridge push' probably have generally similar magnitudes at the present day (e.g. Copley et al., 2010), but uncertainties regarding the composition, structure, rheology, and mode of recycling into the mantle of oceanic-style lithosphere in the Precambrian makes the sizes of these forces in the past uncertain (e.g. Korenaga, 2006; Weller et al., 2019). Crustal thickness contrasts, even when isostatically compensated, result in lateral gradients in Gravitational Potential Energy (GPE), and so a buoyancy force being exerted between mountains and their adjacent lowlands (e.g. Artyushkov, 1973; England and McKenzie, 1982; Molnar and Lyon-Caen, 1988). This force can be the largest exerted on the lithosphere in situations where large mountain belts are present. However, convergence can continue because the force-balance on plates is intrinsically three-dimensional (i.e. smaller forces acting around the entire circumference and base of a plate can balance the large buoyancy force exerted by mountain belts), highlighting the importance of thinking three-dimensionally, rather than along cross-sections through the Earth (e.g. Copley et al., 2010). Because the lithosphere has finite strength, there is therefore a limit to how large the buoyancy force can be, and so an upper limit on the GPE contrast between a mountain belt and the surrounding lowlands. Once this limiting GPE contrast has been reached, a given mountain belt can only expand laterally, along- and/or across-strike, rather than increase in thickness. In this sense, mountain range elevations can act as a 'pressure gauge' for the stresses that are able to be transmitted through the lithosphere (Molnar and Lyon-Caen, 1988).

Within the continents, large lateral variations in rheology lead to striking contrasts in motions and deformation (Fig. 1). Some regions behave as large rigid plates, following the geometrical rules of plate tectonics (McKenzie and Parker, 1967; Le Pichon, 1968). However, in significantly-sized regions of the continents, the geometrical principles of plate tectonics do not apply (e.g. McKenzie, 1972; Molnar and Tapponnier, 1975; England and Jackson, 1989). Regions thousands of kilometres in extent (comparable to the smaller oceanic plates), deform in a spatially-distributed manner. This feature is spectacularly seen in the wide belts of earthquakes, and the mountain belts and basins produced by recent faulting, that stretch through Europe and Asia, and along the western margin of the Americas (Fig. 1; e.g. England and Jackson, 1989). Appendix A describes the alternatives to plate tectonics that were proposed to study these regions of distributed continental deformation, the situations in which they apply, and the main principles of relevance when interpreting the geological record.

Dedicated studies of earthquake depth distributions and mechanisms, when combined with estimates of the temperature structure of the lithosphere and of the effective elastic thickness, have revealed a series of patterns (e.g. Copley et al., 2011a; Copley and McKenzie, 2007; Craig et al., 2012, 2014; Jackson et al., 2008; Maggi et al., 2000; McKenzie et al., 2005): (1) in the relatively undeforming, plate-like, regions of the continents, the sparse earthquakes that do occur have depths of up to 40–50 km, approximately corresponding to the depth of the 600 °C isotherm; (2) within these plate-like continental regions, the forces transmitted through the lithosphere are dominantly supported by the thick seismogenic layer, mostly within the crust, meaning that the thickness of that seismogenic layer is a proxy for the strength of the lithosphere; (3) in regions of pervasively distributed continental deformation (e.g. much of the Alpine-Himalayan belt; Fig. 1), earthquakes are confined to the upper 10–20 km of the crust, roughly corresponding to temperatures of $\leq$300–400 °C; and (4) within these pervasively deforming continental deformation belts, the thinner seismogenic layer and the tractions exerted upon the base of that layer by the deforming ductile mid- to lower-crust mean that the dynamics are governed by the ductile part of the crust, and possibly also the ductile lithospheric mantle. The plate-like continental regions with high strength and large seismogenic thickness also generally have thick lithosphere and are thought to have achieved these properties by past thickening resulting in partial melting of the crust, which removes volatiles and leaves a strong, anhydrous, residue (e.g. McKenzie and Priestley, 2016). The accompanying melt extraction from the lithospheric mantle can compositionally stabilise thick lithospheric roots for billions of years (e.g. Jordan, 1975; McKenzie and Priestley, 2016). The pattern of melt loss resulting in rigid plate-like continental lithosphere can also be seen in the compilation of Cawood et al. (2018), which shows that the Earth's major cratons became stable and undeforming after a period of crustal melting to generate potassium-rich granites, which would leave behind a strong, anhydrous, residue. In contrast, the interiors of many deformation belts are composed of accreted island arcs, or rocks that have been fluxed by water during subduction-enabled ocean basin closure before continent–continent collision. The hydrous nature of such terranes results in a thin seismogenic layer and a relatively low vertically-integrated strength, due to the effectiveness of creep deformation.

It has become clear that faults are considerably weaker than predicted by 'Byerlee's Law' (Byerlee, 1977) due to a some (possibly spatially-variable) combination of the frictional properties of fault zone minerals, such as weak phyllosilicates, and the magnitude of pore fluid pressure, which natural hydrofracture shows can be above lithostatic (e.g. Lachenbruch and Sass, 1980; Sibson, 2004; Lockner et al., 2011; Copley, 2018). Commonly-inferred fault strengths are on the order of megapascals to tens of megapascals (corresponding to effective coefficients of friction of $\leq$ 0.1–0.3), and such faults are only able to support the forces commonly acting upon the lithosphere without

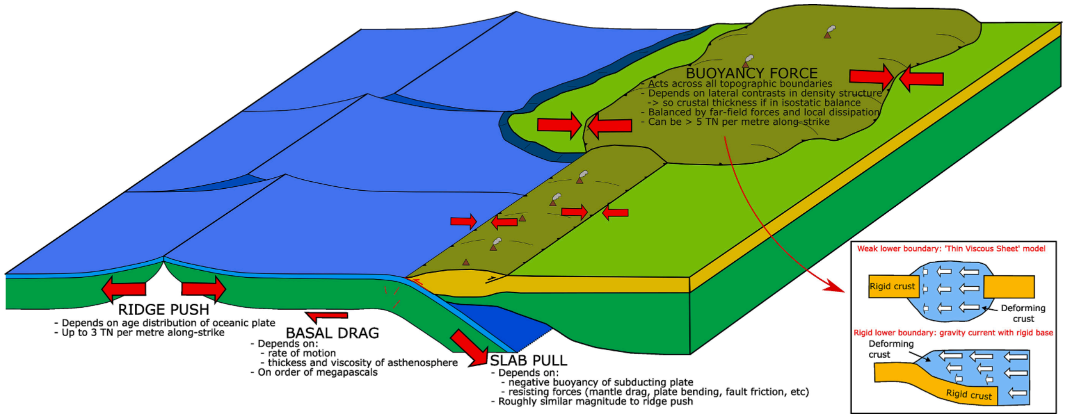

Fig. 2. Schematic diagram of the main forces being exerted on the lithosphere. Other commonly-discussed aspects of the force balance (e.g. the shear stresses on faults) are these forces resolved onto individual structures. For details of the force magnitudes, and the effects of rheology contrasts on the deformation of mountain belts, see Forsyth and Uyeda (1975); McKenzie et al. (2000); Copley et al. (2010).

deforming when the seismogenic layer is thick (≥20 km; Copley, 2018). Therefore, rigid behaviour in the continents occurs in regions where past dehydration by melt extraction has increased the thickness of the seismogenic layer to this value or higher.

The dynamics of lithosphere deformation depends upon whether rigid plate-like regions are present, and in particular whether such material underlies a deformation belt at depth, for example the anhydrous and rigid crust of an undeforming foreland being thrust beneath a mountain belt (e.g. McKenzie et al., 2000; Beaumont et al., 2010; Copley et al., 2011b). In that case, the dynamics of the flow of the deforming part of a mountain belt are governed by vertical planes deforming by simple shear, in a similar manner to a drop of honey spreading out over a table; the driving forces from GPE contrasts result in the shearing of the deforming crust above the rigid underthrusting base. The topography forms a flat top and steep front (e.g. Huppert, 1982), as seen on many range-fronts where rigid forelands underthrust mountain belts (e.g. Penney and Copley, 2021), and the surface motions are in the direction down the local topographic slope (e.g. Copley and McKenzie, 2007). Where no rigid material is present at depth, vertical planes deform by pure shear, more gentle topographic slopes are formed, and the dynamics are governed by a combination of: (1) any GPE contrasts present; (2) the far-field motions of the plates bounding the deformation belt; and (3) the distribution of rigid blocks on the edges of, and within, the deforming zone (e.g. England and McKenzie, 1982; England and Houseman, 1986; Flesch et al., 2001; Penney and Copley, 2021). These different styles of mountain-building lead to distinct pressure (P) and temperature (T) histories in associated metamorphic rocks. Analysis of the P -T conditions experienced in the commonly-observed 'Barrovian' metamorphic belts typical of mountain-building implies that for the majority of past mountain belts, thickening occurred in the presence of rigid underthrusting foreland material (e.g. England and Thompson, 1984; Copley and Weller, 2022), consistent with the deformation patterns observed at the present day (e.g. Argand, 1924; Barazangi and Ni, 1982; Molnar and Lyon-Caen, 1988; Lamb et al., 1997; Nissen et al., 2011; Craig et al., 2012).

2.2. The (re-)distribution of radiogenic heating

Before turning our attention to the Precambrian continents, we first summarise important recent results regarding the distribution of radiogenic heating within the continental crust. The incompatible nature of the most significant heat-generating elements (U and Th at the present day), means that they are much more prevalent in the crust than mantle (e.g. Jaupart et al., 2016). This incompatibility was previously thought to result in their inevitable concentration in the upper crust, due to upwards advection in melts. However, recent results have indicated that because these elements are often hosted in accessory minerals (e.g. monazite and zircon in metapelites, and apatite and allanite in metabasites), which can persist at supra-solidus temperatures, the overall rate of radiogenic heating in rocks undergoes no systematic changes when the wet solidus is crossed (e.g. Alessio et al., 2018; Yakymchuk and Brown, 2019; Weller et al., 2020). Such inferences are consistent with the lack of correlation between metamorphic grade and rate of radiogenic heating in global compilations (e.g. Hasterok et al., 2018).

However, the very high levels of radiogenic heating observed in some granites (e.g. Hasterok et al., 2018), and the low rates of heating in some lower crustal sections required by surface heatflow measurements (e.g. Jaupart and Mareschal, 1999) implies that there is indeed a redistribution of radiogenic heating under some circumstances. It is likely that this redistribution occurs through heating and melting, but that the wet solidus needs to be exceeded by large amounts before the breakdown of the relevant accessory minerals releases the majority of U and Th into the melt. Depending on composition, the breakdown of radiogenic accessory phases may occur at hundreds of degrees above the wet solidus (e.g. Rapp et al., 1987; Bea and Montero, 1999; Yakymchuk and Brown, 2014; Kinney et al., 2024). It therefore seems likely that melting within 100–200 °C of wet solidus temperatures may have a limited effect on the distribution of radiogenic heating, but that above this temperature the radiogenic isotopes may be removed into the melt and transported into upper crustal levels. We discuss below the implications of this concept for the evolution of continental tectonics.

3. Introduction to proposed models of Precambrian continental tectonics

Having summarised the principles governing the rheology and behaviour of the lithosphere at the present day, we will now briefly describe the main conceptual models that have been previously proposed for Precambrian continental tectonics, before interrogating these models based upon the above principles and the range of available observations. The models below have been proposed to operate at a range of different times in Earth history, with some authors suggesting that multiple tectonic styles operated sequentially, or simultaneously within restricted regions of the Earth (e.g. Palin and Santosh, 2021; Cawood et al., 2022).

In the stagnant lid model of tectonics (e.g. Solomatov and Moresi, 1997; Weller and Lenardic, 2018), the lithosphere does not participate in mantle circulation, and there is no significant strain within it. Rather, the entire planet is covered by a single undeforming lithospheric shell. This situation is equivalent to that thought to exist on Mars at the present day (e.g. Nimmo and Stevenson, 2000). Intermediate between this model and that of modern-style tectonics is a 'squishy lid' tectonic regime, in which there is some internal deformation of regions of the lithosphere (e.g. Lourenco et al., 2020; Cawood et al., 2022). A further variant involves planets experiencing periods of stagnant lid behaviour, punctuated by periodic 'mantle overturns' and associated shallow deformation, or periods of planet-wide deformation (e.g. Sleep, 2000; Bedard, 2018). The 'Heat Pipe' model (e.g. Moore and Webb, 2013) is a variation on the stagnant- or squishy-lid viewpoint, and also features a lithosphere undergoing little deformation. However, melt transport upwards through the lithosphere leads to plentiful volcanism and thickening of the crust. The dynamics of stagnant/squishy lid models are governed by the balance between the strength of the lithosphere and the forces being exerted upon it by convection in the underlying mantle. Stagnant lid models are generally proposed to be relevant to Hadean and Archean times (see references above), although some authors have suggested this mode of deformation into the mid-Proterozoic (e.g. Piper, 2013; Stern, 2020).

A class of model emphasises density-driven vertical motions within the lithosphere, particularly within the crust. This broad school of models is variously described as 'sagduction' (which is the term we use in the remainder of this paper), 'density-driven overturns', and 'partial convective overturns', amongst others. The common feature of these models is that more-dense rocks are present above less-dense rocks (for reasons of temperature, composition, or both), resulting in vertical motions driven by the density difference (e.g. Collins et al., 1998; Van Kranendonk et al., 2004; Lin et al., 2013; Francois et al., 2014). The behaviour of a model in which sagduction is important depends upon the magnitude of the density contrasts, the viscosity of the rocks involved, and the geometry of the sagducting bodies.

A further conceptual model views Precambrian continental tectonics as equivalent to that at the present-day. Although the volume and composition of the continental crust and mantle lithosphere will have changed through time, this viewpoint emphasises the equivalent nature of tectonic deformation to that at the present day. By definition, all views of the temporal evolution of continental deformation must eventually result in modern-style tectonics. The main difference between previous studies is when such a style of behaviour is thought to become dominant (e.g. Palin et al., 2020). The main features of present-day continental deformation can be thought of as the lateral contrast between rigid plate-like regions and wide belts of pervasive deformation (e.g. England and Jackson, 1989), and the effects of these rheology contrasts on the dynamics of the deformation (e.g. Copley et al., 2011b). As discussed above, the crustal thickness in large mountain belts is a proxy for the strength of the bounding forelands (Section 2.1; e.g. Molnar and Lyon-Caen, 1988).

4. Model discrimination based on observations and calculations

We will now discuss the observations and lines of logic that we have available to probe the nature of Precambrian continental tectonics, and use them to assess the viability of the above viewpoints.

4.1. Continental growth through time

The use of geochemical proxies to infer the history of growth of the continental crust has received much attention, with many different models being proposed. We refer the interested reader to recent works including Hawkesworth et al. (2010); Dhuime et al. (2017); Spencer et al. (2017); Cawood and Hawkesworth (2019). Debate has focused not only on the measurement and interpretation of the isotope patterns in the zircon record, but also on how representative that record is (Cawood et al., 2022). The preservation of un-reset detrital zircons requires (meta-)sedimentary rocks, derived from a relatively felsic source region, that have not undergone re-melting or had their zircons reset (e.g. Rubatto, 2017), and are available at the surface for sampling. The record is therefore skewed towards sedimentary rocks deposited on stable cratonic crust, or zircons that have been re-eroded from such settings and re-deposited elsewhere. In addition to this bias in tectonic setting, there is also the well-known periodicity to zircon production (and preservation) during the assembly of supercontinents, due to the large volumes of felsic magmas produced (see references above). A further concern when interpreting the ancient isotopic record relates to the timescales required for a signal of melt extraction to be circulated through the mantle, and re-sampled as a 'depleted' mantle source. This timescale is likely to be on the order of half a billion years on average (Rudge, 2006), but the inefficient stirring and mixing in the mantle means that the timescale will vary significantly between different packages of mantle, which is significant when our observational record is limited. There are therefore multiple timescales that need to be considered when interpreting the zircon record, including mantle convective timescales, crustal residence times, the timescales of surface processes, and the time intervals between repeated tectonic events in continental material of a given rheology.

Due to the above considerations, and the thorough and insightful previous work on the topic (e.g. Hawkesworth et al., 2010; Dhuime et al., 2017; Spencer et al., 2017; Cawood and Hawkesworth, 2019), we therefore do not consider in this paper the growth of continental volume through time. Instead, we focus on the behaviour of the continental material that was present at a given time. As discussed in the Introduction, we define 'continental' as crustal columns that are not recycled into the mantle because they are less dense than peridotite. Results from the Jack Hills zircons imply that at least some such material (i.e. felsic igneous rocks) was present from at least 4.3 Ga, and had a source that included a component of material altered in a (near-)surface hydrosphere (Mojzsis et al., 2001; Watson and Harrison, 2005). However, the original igneous rocks that hosted these (now-detrital) zircons have since been lost from the surface record, and some question the evidence for the involvement of supracrustal geochemical reservoirs (Whitehouse et al., 2017). We discuss below how this continental material could have behaved.

4.2. Mantle potential temperature and lithosphere strength

One of the most unambiguous signatures of temporal changes in the solid Earth is the rarity of komatiites in the second half of the Earth's history. The prevalence of such high-Mg magmas has been used to infer higher mantle potential temperatures in the early Earth than at the present day (e.g. Nisbet et al., 1993; Arndt, 2003; Grove and Parman, 2004). However, controversy surrounds how much hotter the convecting mantle may have been. Herzberg et al. (2010) used the Fe and Mg contents of non-subduction-related basalts to suggest melt fractions corresponding to potential temperatures of 1500–1600 °C in the Archean, a view that has gained significant traction. However, Ganne and Feng (2017) suggested that the range between the hottest and coldest potential temperatures was similar in the past to the present day (e.g. comparing the potential temperatures beneath mid-ocean ridges and in plumes) and that the absolute change through time was more modest (e.g. ~150 °C). McKenzie (2020) suggested that komatiites could be produced from mantle plumes only ~50 °C hotter than at the present day. Attempts to model the evolution of potential temperature in the past are often posed in terms of the Urey ratio (e.g. Korenaga, 2008). This quantity represents the ratio of the rate of heat production within the Earth to the rate of heat loss, which at the present day is dominated by extensional tectonics in the oceans. However, the rate of heat loss depends upon the details of the tectonic configuration and behaviour, which are unknown in the early Earth, and additionally there is no simple relationship between the temperature of the Earth and the rate of tectonic activity, so it is therefore likely that the Earth's Urey ratio has changed through time (e.g. Jaupart et al., 2007).

Our concern here is not to estimate the mantle potential temperature in the past, but rather to examine whether the controversial, but hotter, potential temperature would have significantly affected continental rheology and mechanics. We here focus on steady-state geotherms to investigate the feasibility of forming rigid plates in such a situation, and discuss below the evolution of temperature during deformation events. In this section we focus upon Archean-type values for the mantle potential temperature and rate of radiogenic heating, and note that the Proterozoic will represent a situation intermediate between the Archean and the present day.

Fig. 3 shows steady-state geotherms calculated using the method described in Copley and Weller (2022), which includes taking into account the temperature dependence of the thermal parameters. A full description of the construction of these geotherms can be found in the supplemental information. The black curve shows a generic model for a modern-style situation with a mantle potential temperature of 1315 °C and an average rate of radiogenic heating in the crust of 0.8 $\mu W/m^3$ (based upon the results of Rudnick and Gao (2014); Hacker et al. (2015); Jaupart et al. (2016); Copley and Weller (2022), and references therein). In light of the above discussion regarding the distribution and movement of radiogenic heating within crustal sections (Section 2.2), we show models with two different configurations: panel (a) shows a model in which this heating is spread evenly through the 35 km thick crust, and panel (b) shows a model where the rate of heating in the upper third of the crust is five times greater than that in the lower two thirds (e.g. Hacker et al., 2015), but with the same total radiogenic heat output in the crust as a whole. Solid geotherms correspond to lithosphere thicknesses of 200 km (as commonly observed in the plate-like regions of continents at the present day; Fig. 1), and dotted lines to thicknesses of 100 km (the approximate thickness to which the lithosphere can grow by cooling, without requiring tectonic thickening or compositional changes; e.g. Parsons and Sclater, 1978; Crosby et al., 2006; Holdt et al., 2022). As discussed in Section 2.1, within plate-like regions of the continents the forces are dominantly transmitted through the seismogenic layer, so the thickness of that layer provides a useful proxy for lithosphere strength in such regions. Fig. 3c therefore shows the seismogenic thickness calculated from the thermal models for temperatures corresponding to the approximate base of the seismogenic layer in hydrous (350 °C; open symbols) and anhydrous (600 °C; solid symbols) settings.

Red curves and symbols show models equivalent to the present-day ones shown in black, but with the mantle potential temperature increased to 1550 °C. The green and blue curves are equivalent to the black and red ones, but with the rate of radiogenic heating increased to

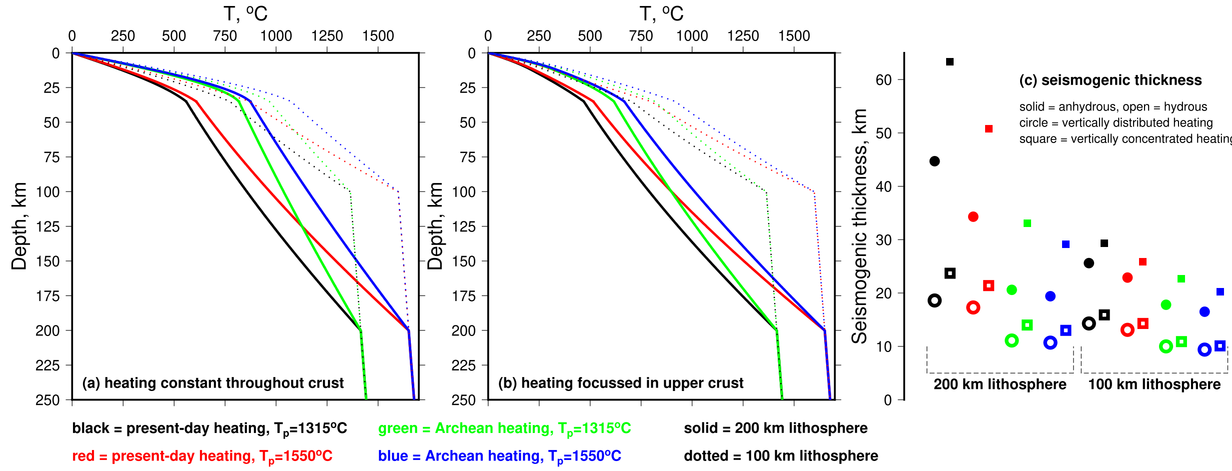

Fig. 3. (a and b) Steady-state geotherms for a range of lithosphere thicknesses, mantle potential temperatures (T_p), and volumes and distributions of radiogenic heating. These geotherms were constructed using the method of Copley and Weller (2022), and a detailed description of the methods can be found in the supplementary material. (c) Seismogenic thicknesses calculated from the geotherms in (a) and (b). See text for details.

simulate Archean conditions. The appropriate amount of increase is difficult to estimate, as this factor depends not only upon the rate of radioactive decay, which is well known, but also the spatial distribution of radiogenic isotopes in the past, which isn't. We here multiply all the rates of radiogenic heating in the various scenarios by a factor of two to simulate plausible Archean conditions, but note that multiplying by a factor of three (see supplemental information) does not significantly affect our conclusions. We retain a crustal thickness of 35 km in all models, due to the $P-T$ conditions recorded by the metamorphic rock record implying that crustal thicknesses may have been similar in the Archean to the present day (e.g. England and Bickle (1984), and discussed in more detail below). If the crustal thickness were thinner in the past, then the seismogenic thickness would have been larger due to the reduced amount of radiogenic heating (for a given crustal concentration), as indicated in the models shown in the supplement for a crustal thickness of 25 km. If the crustal thickness were larger, for example the ~45 km suggested by Tang et al. (2021) using Eu anomalies in zircons, then our conclusions described below still hold true (see supplemental material). The supplemental information also shows that our conclusions discussed below are unchanged if a higher mantle potential temperature of 1650 °C is used, beyond the upper bound suggested by Herzberg et al. (2010).

Comparing on Fig. 3c the black symbols with the red ones, and the green ones with the blue ones, shows that if the other parameters remain constant then increasing the mantle potential temperature to 1550 °C has minimal effect on the seismogenic thickness. By extension, it would therefore have minimal effect on the strength of continental lithosphere that is in thermal steady-state. Changing the total amount of radiogenic heating (i.e. comparing black with green, and red with blue), has a much more significant effect, as does changing the vertical distribution of that heating (comparing circles of a given colour with squares of the same colour), and changing the lithosphere thickness (comparing the left half of Fig. 3c with the right half). We note that changing the vertical distribution of radiogenic heating, from a state where it is evenly distributed to one where it is concentrated at shallow depths, may represent an evolution through time due to radiogenic isotopes being transported by melts (Section 2.2).

A higher mantle potential temperature in the past is therefore likely to have had little direct effect on the strength of the continental lithosphere in steady-state 'intraplate' settings. A more important consideration is the thickness of that lithosphere, the hydration state of the crust, and the amount and distribution of radiogenic heating within it. The available information regarding the thickness of the continental lithosphere on the early Earth, and the composition and history of the ancient crust, is patchy. However, some observations that are available from multiple different locations can provide some insights. Re-Os ages imply the production of melt-depleted lithospheric roots beneath what are now cratons by at least ~3 Ga (Pearson et al., 2021), and plentiful 'cratonic' diamonds were being formed by at least 3.5 Ga (e.g. Helmstaedt et al., 2010), implying that in places the lithospheric mantle had become thickened by that date. High-grade melt-depleted crustal metamorphic rocks were present from at least 3.65 Ga (Horie et al., 2010). These dates all imply that by the early Archean the seismogenic thickness in at least some parts of the continental crust in thermal steady-state could have been at least 20–30 km, and possibly higher (i.e. solid symbols on Fig. 3c representing anhydrous crustal rocks, potentially underlain by thick lithosphere), regardless of the mantle potential temperature. The present-day continents show us that such regions are able to support the forces exerted upon them by large-scale plate driving forces, and GPE contrasts with adjacent mountain belts, without appreciably deforming (e.g. Copley, 2018; Knight et al., 2021; Wimpenny, 2022). Therefore, we can infer that rigid block-like regions of the continents are likely to have been in existence from at least the early Archean – at least parts of the Archean continents probably weren't 'squishy'. Additionally, based upon the discussion above regarding the limit that foreland strength plays on mountain range elevation (e.g. Molnar and Lyon-Caen, 1988), these rigid plate-like regions would have been able to support significantly-sized mountain belts on their edges. These conclusions regarding Archean continental behaviour also apply to the Proterozoic, as the decreasing rates of radiogenic heating would result in increasing values of the seismogenic thickness compared to the Archean.

4.3. Fault strength and stagnant lids

Having established that parts of the Precambrian continental lithosphere were likely to have been strong enough to behave in a rigid manner by at least the early Archean, we now address the other extreme of whether the lithosphere could have been so strong as to form a single, undeforming, planet-encircling plate (i.e. 'stagnant lid' tectonics; Solomatov and Moresi, 1997; Weller and Lenardic, 2018). The dynamics of this type of tectonic system have been investigated in detail by Weller and Lenardic (2018), who describe how the transition between stagnant lids, 'mobile lids' (i.e. deforming lithosphere), and states that fluctuate between these end-members, depends upon the balance of the forces being exerted on the base of the lithosphere and the vertically-integrated strength of the lithosphere. Due to the nature of their upper boundary condition, the driving forces in their model result from internal density contrasts, as topography is not present on the upper boundary. The results show that for yield stresses in the upper layer of less than ~50 MPa, only 'mobile lid' solutions exist. Above this stress level, the lithosphere is strong enough to behave as a 'stagnant lid', or to oscillate between these two states.

Understanding whether the early Earth could support stagnant lid tectonics therefore requires understanding the vertically-integrated strength of the lithosphere. At the present day, we know that some regions of the continents are strong enough to support vertically-integrated stresses that would be compatible with stagnant lid models, as indicated by the high differential stresses required to support the observed gravity anomalies and GPE contrasts (i.e. tens to hundreds of MPa; e.g. Copley, 2018). However, to produce a planet-encircling stagnant lid requires all of the lithosphere to be able to support such stresses, as if weak zones are present then they will deform, and the tectonics will resemble rigid regions separated by deformation belts, much like at the present day.

As discussed in Section 2.1, the strong and plate-like parts of the continents at the present day generally have a characteristic geological history, being anhydrous and with thick lithosphere due to a prior history of tectonic thickening and melt extraction. However, the limit on whether the Earth can support a single 'stagnant lid' will depend on the strength of other areas of the continents that have not yet undergone this process. In the pervasively deforming regions of the continents, earthquake stress-drops (which are likely to represent the great majority of the pre-earthquake shear stresses that can be supported by faults; e.g. Copley, 2018) are generally on the order of 1–10 MPa (e.g. Kanamori and Anderson, 1975; Allmann and Shearer, 2009). This estimate also provides an upper bound on the level of differential stress in the ductile part of the crust and mantle, as if the differential stresses were higher than can generate seismic failure then the rocks would deform by earthquakes, rather than flow. Therefore, in such regions the vertically-integrated strength is likely to be an order of magnitude less than required for a 'stagnant lid' tectonic regime. As discussed in Section 2.1, the weakness of active faults in comparison to 'Byerlee's Law' is likely to be due to a combination of weak phyllosilicates being present in fault zones, and high pore fluid pressures. Both of these situations are dependent on there being available hydrous fluids, to fill pores and to form hydrous phylloslicates. The implication from the oxygen isotope record of a surface or shallow hydrosphere being present by at least the time of formation of the Jack Hills zircons (i.e. before 4 Ga; Mojzsis et al., 2001), implies that there is little reason why fault strength should have been uniformally higher than it is at the present day at any time from the mid-Hadean onwards. The weakness of faults therefore indicates that the early Earth would not have experienced a stagnant lid tectonic regime.

4.4. Density contrasts and vertical motions

The physical viability of density-driven overturns of the crust ('sagduction') depends upon the geometry of the lithologies, the density contrasts between them as a function of pressure and temperature, and the rheology. The viability of this style of tectonics can be readily addressed using a simple model of 'Stokes settling' of a denser lithology through less-dense surroundings (e.g. Batchelor, 1967; Miocevich et al., 2022). Configurations in which the deformation is as 'drips', or similar, still mechanically connected to a near-surface layer, result in slower motions as work is done in internally deforming the 'drip'. For the case of Stokes settling, the rheology of the sinking lithology is relatively unimportant, as the sinking velocity of a rigid sphere and an inviscid one only differ by a factor of 1.5; the dynamics are instead governed by the rheology of the surrounding material, which must deform to accommodate the sinking of the denser body (e.g. Batchelor, 1967).

The viability of density-driven vertical motions in the crust was recently investigated by Miocevich et al. (2022). An important component of this work was constructing bespoke phase diagrams ('pseudosections') of a range of common lithologies, to establish their mineralogy as a function of pressure and temperature, and so the density contrasts between them. Fig. 4a shows the results of Miocevich et al. (2022) displayed in a different format to that paper. For a range of viscosities and sizes of a sinking body, the figure shows the rate at which the more-dense lithology will sink, in mm/yr (note the logarithmic scale). The example shown is for a mafic lithology sinking through a felsic one (e.g. meta-basalt through meta-granitoid, with a density contrast of 440 kg/m^3). These estimates are upper bounds of the possible velocities, as they are calculated for an inviscid body that is unconnected to any overlying layers (see Miocevich et al. (2022) for further details).

For bodies of dimensions of 1–10 km, as are commonly inferred to be emplaced at depth by vertical motions, viscosities of less than 10^{19}–10^{21} Pa s would be required for the rocks to sink at a geologically significant rate of 1 mm/yr (thick contour on Fig. 4a). The most viscous rocks encountered by a 'sagducting' body will be immediately beneath the seismogenic layer, at temperatures of ~350 °C in hydrous rocks or ~600 °C in anhydrous ones. Fig. 4b shows effective viscosity as a function of temperature calculated from ductile flow-laws for olivine, clinopyroxene, anorthite, and quartz (Rybacki and Dresen, 2000; Mei and Kohlstedt, 2000; Bystricky and Mackwell, 2001; Hirth and Kohlstedt, 2003; Rutter and Brodie, 2004; Hier-Majumder et al., 2005; Rybacki et al., 2006). We have used a grain size of 5 mm, a water fugacity of 1 GPa, and a strain rate of 10^{-14} /s, similar to that in the shearing mid-crust of the Tibetan Plateau at the present day. At temperatures corresponding to the brittle-ductile transition, none of these minerals have a low enough viscosity to allow the mafic body to sink at a significant rate. Close to the wet solidus, some of these flow-laws become viable. However, rocks that have been above the solidus and had melt extracted from them will become anyhydrous, in which case the 'dry' versions of the flow-laws become relevant. These flow-laws are all incompatible with a significant rate of sinking of mafic material through a felsic crust over almost the entire range of crustal temperatures. Taken together, these results therefore imply that sagduction is not an important process in continental tectonics (Miocevich et al., 2022). Such a result is unsurprising, given that present-day observations show us that mafic and ultramafic ophiolite sequences can be emplaced within the continental crust at high temperatures, and remain in the upper crust without 'sagducting' to depth. Indeed, the preservation of Precambrian mafic and ultramafic lithologies at high crustal levels (e.g. in the greenschist facies; Anhaeusser, 2014) demonstrates the ineffectiveness of density-driven tectonics within the continents: if the mechanism were effective, mafic rocks would not have overlain granitic rocks for billions of years.

The possible exception to the above conclusion is where plentiful melt is present to reduce the effective viscosity of the surroundings, in which case density-driven vertical motions could be more effective. Such a mechanism is likely to aid the proposed density-driven segregation that occurs beneath island arcs during the formation of continental crust (e.g. Hacker et al., 2015). However, it is unlikely that significant volumes of melt would have been present at the greenschist-facies conditions at which many granite-greenstone belts are preserved (e.g. Anhaeusser, 2014), and which are commonly taken to represent the structural imprint of sagduction (e.g. Van Kranendonk et al., 2004; Francois et al., 2014). To have a significant effect on the rheology, and so effectiveness of density-driven vertical motions, such melt would need to be present in the liquid (i.e. rather than solidified) form over significant thicknesses and lateral extents. Although melt has intruded upper-crustal successions in multiple greenstone belts (e.g. de Wit et al., 1987; Anhaeusser, 2014), we are not aware of any observations that imply the simultaneous presence of liquid melt over the vertical and lateral extents that would be required to make sagduction viable. Alternative models for the formation of the structures seen in granite-greenstone belts will be discussed below.

4.5. Radiogenic heating

Amongst the most well-constrained and dramatic changes in the

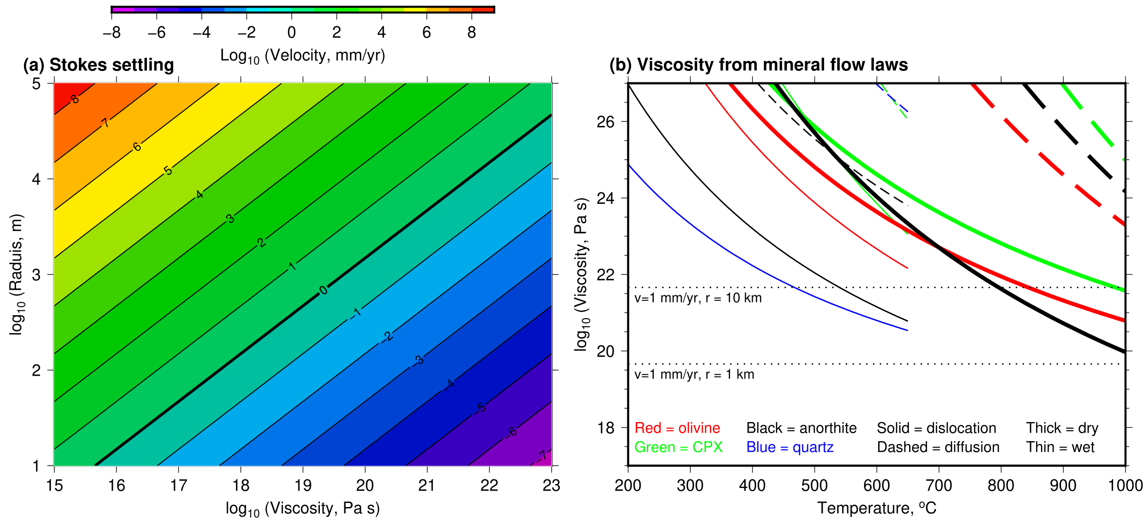

Fig. 4. (a) Sinking velocity (note logarithmic scale) for meta-basalt through meta-granitoid, as a function of the radius of the sinking body and the viscosity of the granitoid. (b) Effective viscosities calculated from mineral flow-laws using the parameters described in the text. 'Wet' flow laws are only shown up to a generic temperature for the wet solidus (650 °C). CPX = clinopyroxene. v = velocity.

properties of the Earth since it formed is the decrease in the rate of radiogenic heating. The most important radiogenic isotopes at the present day, ^{232}Th and ^{238}U, were slightly more abundant in the early Earth, but were supplemented by large contributions from ^{40}K and ^{235}U, which have short enough half-lives that they are now mostly extinct (e.g. O'Neill et al., 2020b). For example, the total rates of radiogenic heating in the entire Earth at 2, 3 and 4 Ga would have been higher than the present day by factors of approximately 1.7, 2.4, and 3.7. However, as noted above, our incomplete knowledge of the formation and differentiation of the crust means that the geological distribution of these radiogenic isotopes is not well known (e.g. O'Neill et al., 2020b).

Fig. 3 shows an example of how higher rates of radiogenic heating could have affected the thermal structure of the continents in steady-state, undeforming settings. We suggested above that some parts of the Archean (and younger) continental lithosphere would have acted in a strong plate-like manner, after prior melt extraction and the production of lithospheric mantle roots. In addition to these results we also need to consider how the tectonics of deformation belts could have been affected by the higher rates of radiogenic heating in the early Earth.

Fig. 5 shows calculations of the thermal evolution of the lithosphere during a generic mountain-building event, equivalent to those presented in Copley and Weller (2022). In panel (a), the rate of radiogenic heating is set at a value often obtained by studies of present-day continental crust (1 μW/m^3). Panel (b) shows an equivalent model, with the only difference being that the rate of radiogenic heating is doubled in the upper half of the crust. Panels (c) and (d) show the effects of doubling the rate in the entire crust when compared to (a), or tripling the value in the upper half of the crust. In all of these models, where the temperature exceeds 900 °C the radiogenic heating is reduced to a background value of 0.7 μW/m^3, and the remainder is redistributed through the crust at depths shallower than the 900 °C isotherm, to simulate the effects of heat-producing isotopes being transported upwards by high-temperature melts (see Section 2.2).

In all of the models on Fig. 5, increasing the rate of radiogenic heating dramatically increases the temperatures attained during the modelled configuration of crustal thickening. Clearly there will be feedbacks in operation, with the temperatures influencing the rheology and so the dynamics of mountain belts. Additionally, the figure shows only one of a wide range of possible mountain-building configurations. However, the simple models in Fig. 5 reveal the likely hundreds-of-degrees of additional heating that will occur for a given mountain-building event in the presence of plausible values for the distribution of radiogenic heating in the Archean and Proterozoic. Two major and inter-linked implications of these high temperatures are on the rheology (and therefore deformation) of mountain belts, and on the production of crustal melts. We now discuss each of these in turn. However, we first note that the models shown in Fig. 5 support prior suggestions that the widespread production of high-grade and melt-depleted metamorphic rocks in the Precambrian was related to the high rates of radiogenic heating that were present (e.g. McKenzie and Priestley, 2008; Jaupart and Mareschal, 2015).

4.5.1. *The dynamics of hot mountain belts*

We will first discuss the case where rheology depends only upon temperature, and then extend our discussion to include the effects of composition. Higher temperatures within a mountain belt will result in a decrease in the thickness of the seismogenic layer, and a reduction in the average viscosity of the ductile part of the lithosphere. The effect of these rheological changes on the dynamics of a mountain belt depends upon the geometry of the deformation. For the simple case of homogeneous thickening between two converging rigid blocks (Fig. 6a), the rate at which the range can shorten, and so the rate of thickening, depends on the depth-integrated strength of the mountain belt (e.g. Sonder and England, 1986). This quantity is dominated by the effects of the strong upper layers of the crust, within the seismogenic layer and coolest parts of the ductile crust. The solid line on Fig. 6a shows the results of a calculation for the depth-integrated strength of the crust (using the method described in the supplementary information), as a function of the Moho temperature, with the upper part deforming by brittle failure and the lower part by ductile flow with a viscosity given by the hydrous anorthite dislocation creep flow-law of Rybacki and Dresen (2004). Changing the temperature structure over the considered range affects the depth-integrated strength by a factor of 4, so for a given distribution of driving forces the rate of convergence and thickening in a mountain belt would only differ by a factor of ~4 if the Moho temperature is increased from 600 °C to 1000 °C.

As discussed above, the case of homogeneous thickening is not prevalent at the present day, and the more common configuration is of a mountain belt growing laterally over rigid underthrusting foreland material (Fig. 6b). In this case, the rate of motion is dominated by the weakest layers within the over-riding deforming crust. This configuration is illustrated in Fig. 6b, which also shows the velocity as a function of the temperature at the decollement between the underthrusting rigid crust (orange) and the overlying deformation belt (blue). Changing the

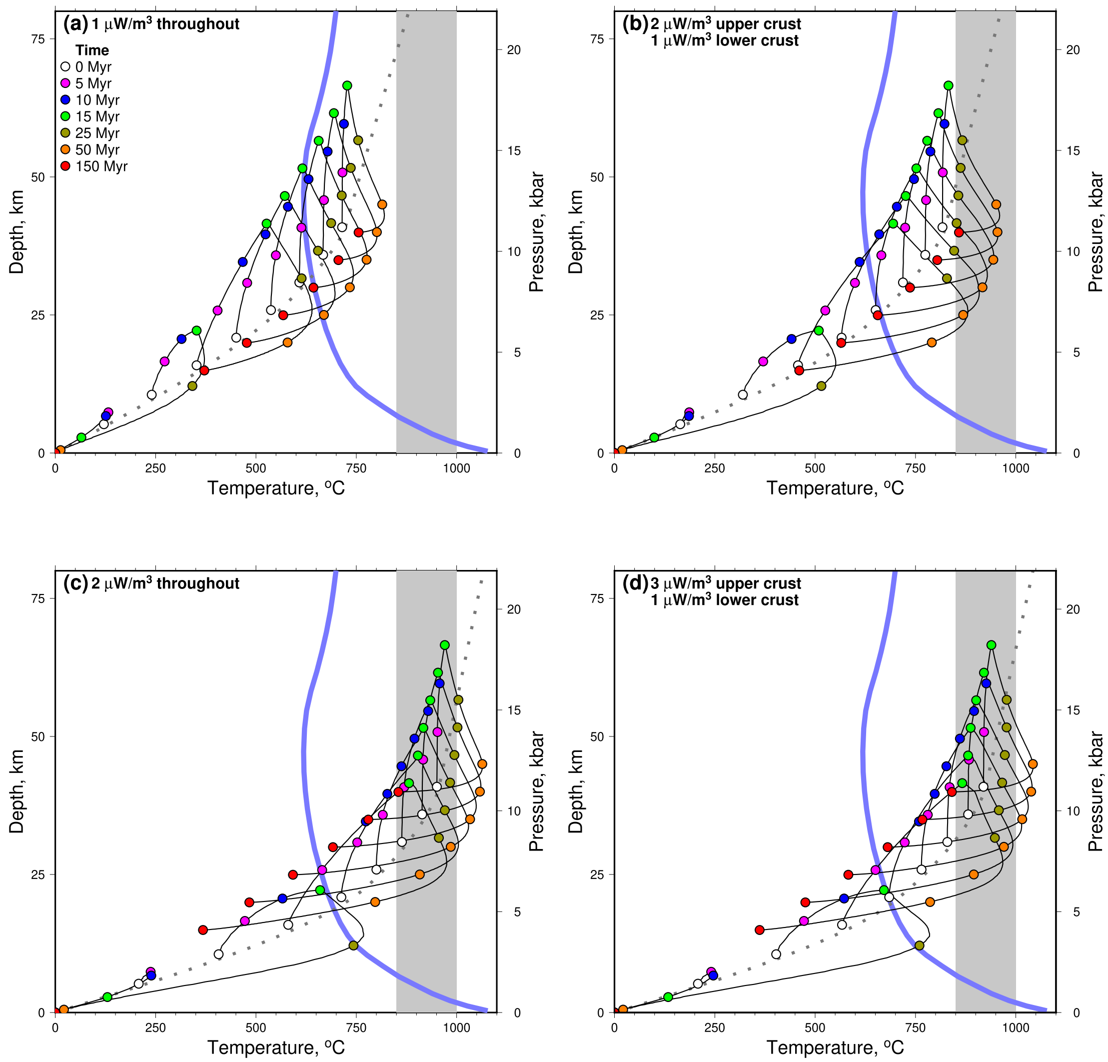

Fig. 5. Calculations for the thermal evolution of a generic mountain belt. (a) is taken from Copley and Weller (2022). The blue line shows the wet basalt solidus of Lambert and Wyllie (1972), and the shaded band the suggested temperature range for melting to produce TTG melts (Section 4.5.2). The model involves thickening of initially 40 km thick crust that includes rigid lower crust with an initial top depth of 15 km, at a rate of 2.5 mm/yr. The initial lithosphere thickness is 150 km, and thickening occurs for a duration of 15 Myr. The rate of radiogenic heating is 1.0 μW/m^3 throughout the crust, and the erosion coefficient is 1.5/s. See supplemental material for details of the calculations. The dotted line shows the initial geotherm. The lines track rocks initially located at 5 km depth intervals from 5 to 40 km, with the deepest point representing the Moho, and the coloured circles indicate times since the start of the model run. The top apex in the depth–temperature–time paths indicates the end of thickening at 15 Myr, at which point erosion then dominates the vertical motions, as the thickened crust continues to increase in temperature. The other panels show the results of identical models, but compared to (a), in (b) the rate of radiogenic heating is doubled in the upper half of the crust, in (c) it is doubled throughout the crust, and in (d) it is tripled in the upper half of the crust. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

decollement temperature from 600 °C to 1000 °C results in almost four orders of magnitude increase in deformation rates, due to the exponential dependence of viscosity on temperature. Once Precambrian mountain belts had thickened sufficiently to begin flowing over an adjacent rigid foreland, if the temperature effect were the only control on the rheology we might expect the rates of deformation to be much higher than is common at the present day, due to the higher temperatures resulting from the greater rates of radiogenic heating. Comparing the results from the two geometries considered here highlights the important role of the deformation geometry, controlled by the tectonic boundary conditions, and gives rise to the concept that the 'effective strength' of a mountain belt can very laterally due to the geometry of deformation, even if the profile of strength with depth in the deforming part of the crust is everywhere the same (as explained for the case of different parts of the Tibetan Plateau by Copley and McKenzie, 2007).

However, to the above considerations we need to add the effects of melting, which are two-fold. The presence of melt within the crust is known to have a dramatic weakening effect, although the size of that effect depends on how interconnected the melt is, and its geometry with respect to the deformation field (e.g. Cavalcante et al., 2016). Once the melt is removed, the remaining anhydrous residue will be much stronger than the pre-melt protolith (e.g. Diener and Fagereng, 2014). The rates and dynamics of melt transport are much more poorly known for crustal rocks than they are in the mantle (e.g. Brown, 2013). However, the preservation of macroscopic melt pockets frozen in place within their source rocks show that in-situ melt fractions can be much higher than in mantle rocks (e.g. >10 % compared to <1 %; Sawyer, 2008; Katz et al., 2022), which is likely to be due to a combination of the difference in dihedral angle between the crust and mantle (which affects pore connectivity and so melt drainage), and differences in the viscosities of mafic and felsic melts. The dashed lines on Fig. 6 show the model results for the case of rocks with temperatures above 750 °C switching from the hydrous to the anhydrous versions of the anorthite flow laws. As the temperature is increased, progressively larger portions of the crust

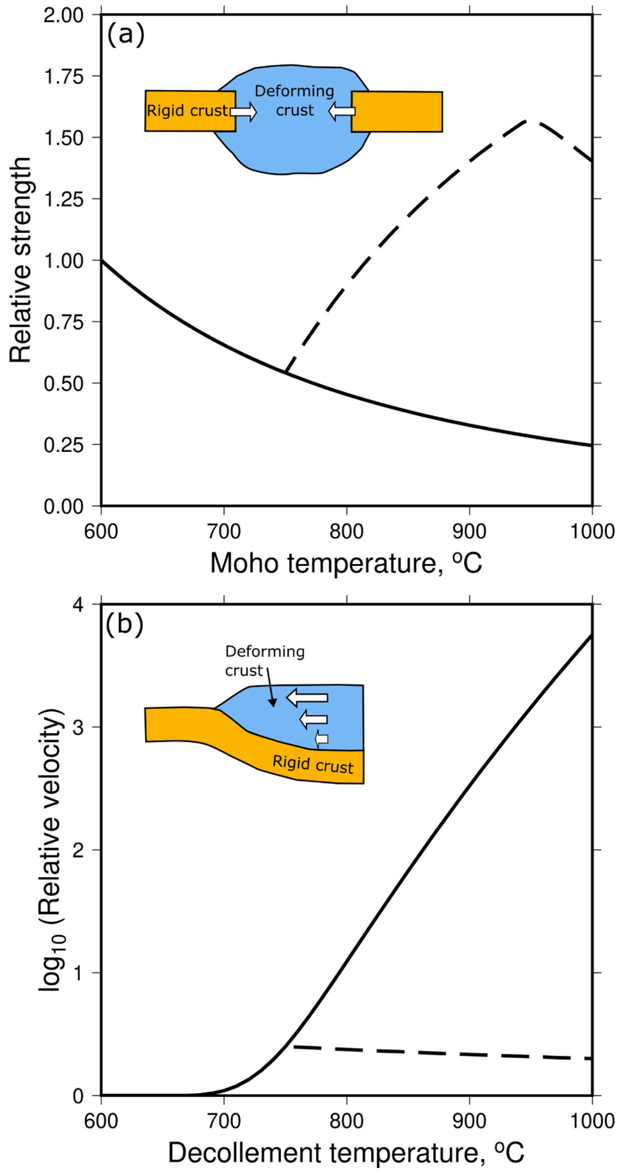

Fig. 6. (a) Vertically-integrated strength, normalised to that when the Moho temperature is 600 °C, calculated using the method described in the supplement. As a normalised measure of strength, the value is dimensionless. Solid line uses a hydrous anorthite flow-law throughout (Rybacki et al., 2006), and the dashed line switches to the anhydrous equivalent at temperatures above 750 °C. (b) Surface velocity for the case of shear over rigid underthrusting material, calculated using the method described in the supplemental material, normalised to the value when the decollement temperature is 600 °C. Note the logarithmic scale. Solid and dashed lines have the same meaning as in (a).

switch to the anhydrous flow laws. For the case of vertically-homogeneous deformation (Fig. 6a), the depth-integrated strength of the crust increases above the cooler and hydrous version. Only at Moho temperatures close to 1000 °C does the strength begin to decrease due to the temperature effects overwhelming the compositional effects. For the case of flow over rigid underthrusting foreland material (Fig. 6b), the rates of motion stabilise, and slightly decrease, above the temperature at which the anhydrous flow laws begin to be used. This effect occurs because the hottest part of the crust becomes strong enough to not appreciably deform, and the weakest layer is continually that experiencing temperatures just below the transition to the anhydrous flow law. The overall rate of motion slightly decreases because of the reduction in the thickness of the hydrous layer that is able to appreciably deform.

When combined, these effects imply that hot, ancient mountain belts may have had relatively similar rates of deformation to present-day ranges, the effects of increased temperatures being balanced by the effects of increased melt-induced dehydration. In this sense, melting and melt removal can act to maintain the strength of the crust (e.g. Diener and Fagereng, 2014). The difficulty in estimating rates of tectonic motion in ancient mountain belts means that we are not able to observationally test this suggestion, although we note that the available palaeomagnetic data implies that the overall rates of plate motion varied within limited bounds between the Mesoarchean (when the estimates begin) and the present (Pesonen et al., 2021). Such a similarity implies similar rates of overall convergence in mountain belts on the margins of these plates, as would be expected if the resistance to deformation in those belts had remained similar through time.

4.5.2. Melting hot mountain belts: A mechanism for TTG petrogenesis

We now address the compositions of melt that could be produced within hot mountain belts, and suggest that this setting provides an explanation for the generation of voluminous tonalite-trondhjemite-granodiorite (TTG) intrusions that are characteristic of the Archean. In this paper, we use the categorisation of granitoids suggested by Moyen (2020a). Striking features of TTG-dominated 'grey gneiss terranes' are that they are large (e.g. 100's km in extent; Neil et al., 2023), dominated by sodic granitoids (containing only rare potassium feldspar), were common in the Archean, and were progressively replaced by biotite (higher-K) granites from the late Archean onwards (Smithies, 2000; Moyen, 2020a; Cawood et al., 2018; Nebel et al., 2018). Trace element signatures have been used to propose a range of possible melting depths (inferred from the stability of plagioclase, rutile, and/or garnet in the source region; Foley et al., 2002; Moyen, 2011). However, recent studies have suggested that these signatures may instead be generated by fractionation of the magmas (Laurent et al., 2020; Rollinson, 2021), or be produced by melting over a protracted temperature range as the trace element signatures are sensitive to temperature as well as pressure (Palin et al., 2016b). The production of magmas able to crystallise TTG suites has long been thought to be by the partial melting of hydrated basalt (Rapp et al., 1991). However, although a range of geodynamic scenarios have been proposed, what is currently lacking is an explanation for the voluminous nature of TTG suites, and a reduction in TTG formation through time.

Melting of thick piles of basaltic rocks, with possibly an influence from plume activity, has been proposed as a mechanism (Smithies, 2000; Kamber, 2015; Johnson et al., 2017). Whilst some TTG-type intrusions are found in this context in Iceland (Reimink et al., 2014), there is no indication that this setting is able to produce the volumes of TTG preserved in ancient terranes, despite the mantle underlying Iceland having a potential temperature of ~1500 °C (Matthews et al., 2016). Shallow melting of underthrusting mafic material, either in a subduction-type setting or as underthusting unattached to a downgoing slab, also has the potential to produce TTG suites (Hastie et al., 2015). However, this mechanism does not provide an explanation for the widespread TTG occurrences in the past, or their relative rarity at the present day (Smithies, 2000; Moyen, 2020a).

We propose an alternative mechanism for TTG formation that can account for both their spatial extent, and decreasing prevalence through time, and which is based upon one of the few quantities that we know has definitely changed through Earth history: the rate of radiogenic heating. We suggest that prior attempts to understand the petrogenesis of TTG suites have not adequately considered the nature and characteristics of distributed continental deformation. Fig. 5b–d, discussed above, shows the evolution of temperature through time in a mountain-building event with Archean-style levels of radiogenic heating. Fig. 7 shows the melt volumes and composition that would be produced by melting basalt or diorite as a function of temperature and water content (see supplementary material for details), at the pressure of 12 kbar that Palin et al. (2016b) previously suggested as most compatible with the volume and trace element distribution of TTGs. We use an enriched

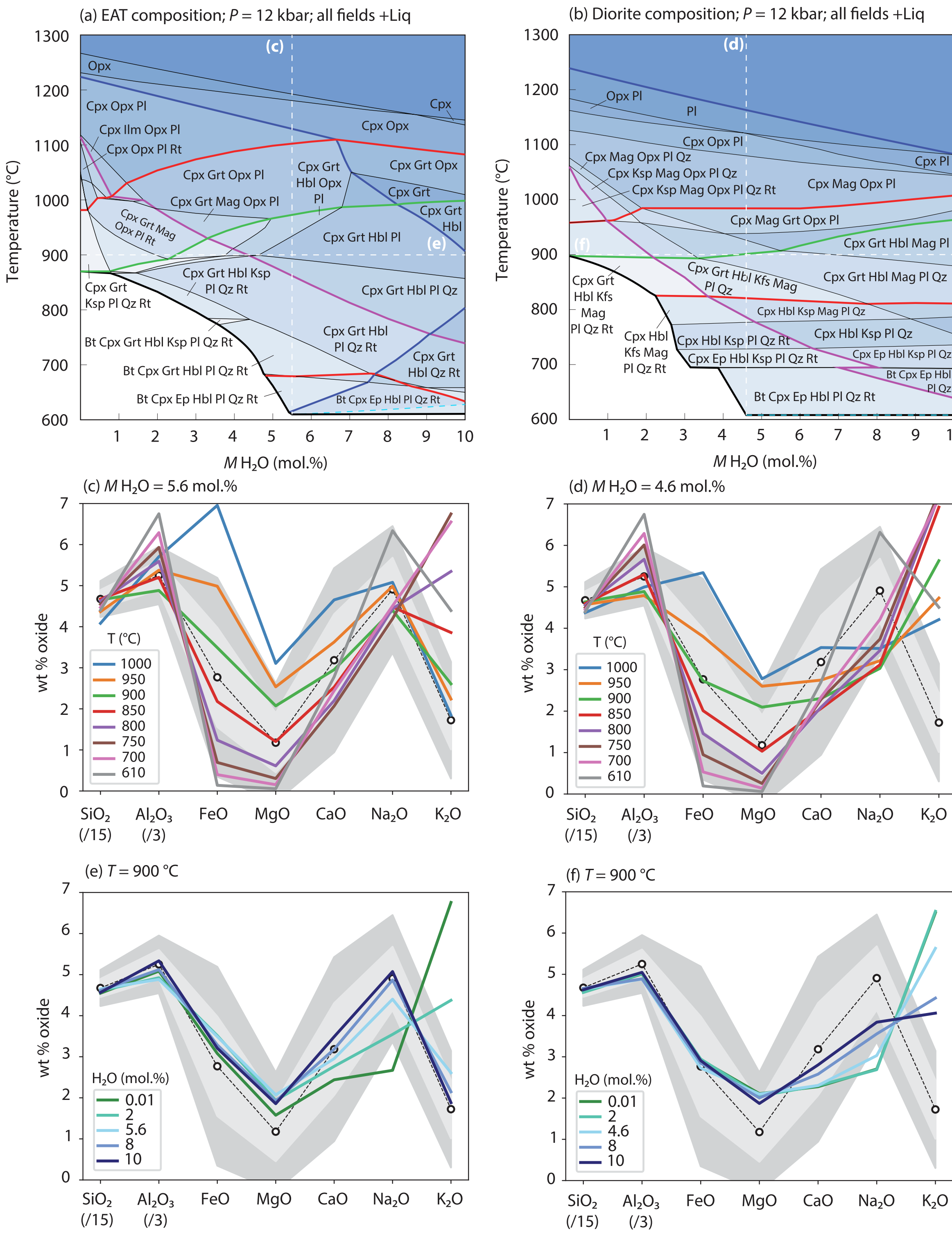

Fig. 7. (a,b) T -M_{H2O} supra-solidus bespoke phase diagrams ('pseudosections') at 12 kbar for (a) the enriched Archean tholeiite (EAT) composition of Condie (1981) and (b) the average diorite composition of Cox (1971). Mineral abbreviations follow Whitney and Evans (2010). Zero-mode isolines are highlighted for garnet (red), hornblende (green), rutile (purple), plagioclase (blue) and liquid (thick black). Dashed white lines correspond to the sections along which melt compositions are explored in (c–f). (c,d) Melt compositions (oxide wt%) as a function of selected temperatures for the fluid-saturated solidii of (c) the EAT composition (5.6 mol.% H_2O), and (d) the diorite composition (4.6 mol.% H_2O). The white dots in the background show the 'average sodic TTG' from Moyen and Martin (2012) (n = 1439), with the light and dark shading representing one and two standard deviations about the mean, respectively. (e,f) As (c,d) but showing melt compositions as a function of selected water contents at T = 900 °C for the (e) EAT composition, and (f) the diorite composition. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Archean tholeiite for the basalt composition, which has previously been suggested to be a suitable source for Archean TTGs (Bedard, 2006; Martin et al., 2014). We also model an average diorite composition (Cox, 1971) to simulate the possible source rocks later in Earth history when the continental crust had become more evolved than the dominantly-mafic Archean crust (e.g. Smit and Mezgar, 2017; Hawkesworth and Jaupart, 2021). These calculations demonstrate that the melting of basalt that is water-saturated at the solidus is able to produce TTG compositions at temperatures of ~900–950 °C (Fig. 7c & e, with an associated temperature uncertainty of ~50 °C; Powell and Holland, 2008; Palin et al., 2016a). The melting of diorite across a spectrum of temperature and hydration conditions always results in melts that are too enriched in potassium to crystallise TTGs (Fig. 7d,f).

Taken together, Figs. 1, 5, and 7 suggest the following model for TTG formation. Distributed deformation of the continents results in thickening, and therefore heating (delayed by tens of Myr from the onset of thickening; England and Thompson, 1984; McKenzie and Priestley, 2016; Copley and Weller, 2022), over regions hundreds to thousands of kilometres in size (Fig. 1). For Archean levels of radiogenic heating, these regions can easily attain temperatures of ≥ 900 °C for rocks that were initially below the wet basalt solidus, at pressures consistent with trace element constraints, even for relatively modestly-sized mountain-building events (Fig. 5). At these temperatures, if it was hydrated the predominantly basaltic Archean crust (Smithies et al., 2005; Chen et al., 2020; Hawkesworth and Jaupart, 2021) would then melt to produce magmas that can crystallise to form TTGs (Fig. 7). Although we are suggesting nothing new in terms of the protolith or P –T conditions for TTG formation, we suggest that spatially-distributed continental mountain-building events represent an important and previously-overlooked tectonic setting for this process.

In this model, as geological time advances, four factors will reduce the likelihood of TTG production. The first is that as the rate of radiogenic heating decreases, so will the temperatures (and so melt fractions) resulting from a given style of collisional event. These lower melt fractions result in higher potassium contents (Fig. 7), and the production of potassium feldspar- and biotite-bearing granitoids, which is a transition described by Moyen (2020b). Second, as time advances it is more likely that a given batch of crust will have sediments deposited on top of it, which will increase the potassium content of any melts resulting from melting a crust of mixed sedimentary and igneous lithologies, if the sediments can be transported to melt regions at depth (Mojzsis et al., 2001). Third, as the continental crust becomes increasingly differentiated by melting, the source region becomes progressively more likely to contain intermediate-composition lithologies, which Fig. 7d & f shows are unable to generate TTG-composition magmas, but are able to produce the more K-rich melts typical of the late Archean onwards. Finally, once a crustal section has experienced a high-temperature melting event, the resulting refractory residue will produce limited additional melt. The only source of melt will then be the TTGs themselves, which can re-melt, produce the high-K magmas often seen before craton stabilisation (Cawood et al., 2018), and then become a rigid crustal blocks themselves.

An implication of the above model is that we would expect the lower crust to be partially composed of the residue from which the TTGs were extracted. The limited exposures of Archean lower crust means that this implication is difficult to test. However, the patchy evidence provided by xenoliths implies that the lower crust is likely to be more mafic than the overlying material, and less radiogenically active (e.g. Rudnick and Gao, 2014; Hacker et al., 2015). Both characteristics would be as expected for material that melted to form TTGs, and achieved high enough temperatures that the accessory minerals that carry radiogenic isotopes would be likely to have melted too, releasing the radiogenic elements into the melt. A further implication of this model is that, in the atypical cases where greenstone belts formed contemporaneously with nearby TTGs (e.g. Hawkesworth and Kemp, 2021), they would have formed within, or adjacent to, mountain belts. However, the generally unknown amount of tectonic motion that has occurred during the juxtaposition of TTGs and greenstone belts means that the actual distance represented by 'adjacent' is poorly constrained. It is therefore not clear whether stratigraphic observations from greenstone belts can be used to test our model of TTG petrogenesis, given that the volcano-sedimentary protoliths could represent either of: (1) crust formed in any tectonic environment that exists in the surroundings of a mountain belt, juxtaposed against the TTGs by tectonic motion; or (2) plausibly represent mafic melts emplaced at shallow levels within a mountain belt due to melting of the underlying mantle (e.g. Yakovlev et al., 2019), presumably at higher volumes than occurs at the present day because of the greater temperatures achieved in the Archean, with the sedimentary components representing deposition in intra-montane basins or on the floor of a range-covering ocean.

A remaining question related to TTG petrogeneis, relevant to our proposed model and previous suggestions, is the requirement to be generating TTG melts for extended timescales of over 100 Myr (e.g. see the compilation of Cawood et al., 2018). From the Proterozoic onwards, even the largest mountain-building events are characterised by high-grade metamorphism and melt production over timescales of < 100 Myr (e.g. Weller et al., 2021). Although we lack the observations to suggest a definitive resolution to the issue of the longevity of TTG production, a possibility is provided by the processes that are thought to limit the lifespan of Proterozoic-to-recent mountain belts. Weller et al. (2021) suggested that the longevity of a mountain-building episode is limited by the timescale required for underthrusting rigid lithosphere on the margins of mountain belts to meet at depth, and resist further shortening (e.g. as is currently beginning to occur in western Tibet; Craig et al., 2012). If this suggestion is correct, then the sparser distribution of rigid crustal blocks in the Archean would result in longer timescales before they collide at depth beneath a mountain belt (as this timescale is set by their initial separation and the convergence rate), possibly resulting in mountain belts that last longer, with a more protracted period of underthrusting, thickening, and melting to produce TTGs. This suggestion can be tested by additional research into the longevity and configuration of Archean mountain belts.

We therefore suggest that TTG petrogenesis can be achieved in the uniformitarian setting of the interiors of spatially-extensive continental mountain belts, in the presence of the dominantly-basaltic Archean crust and the elevated levels of radiogenic heating relevant to such times. However, we note that not all TTGs necessarily need to have been formed in exactly the same manner, given the geochemical variability that is present (e.g. Johnson et al., 2019; Moyen, 2020a). Therefore, other (uniformitarian or otherwise) mechanisms may also be relevant, such as those described above including the melting of thick basaltic piles in settings broadly equivalent to oceanic plateaus (with the balance between igneous addition and tectonic thickening debated, e.g. Smithies, 2000; Bedard et al., 2013; Kamber, 2015; Palin et al., 2016b; Johnson et al., 2017), or during underthrusting of oceanic lithosphere with or without an attached subducting slab (e.g. Arndt, 2013; Palin et al., 2016b; Hastie and Fitton, 2019). However, these other mechanisms do not provide as clear an explanation for the spatial and temporal distribution of TTG suites as our suggested method of petrogenesis.

4.6. Deformation fabrics and lower crustal rheology

A key observation for understanding ancient continental tectonics is the nature of structures and fabrics in the preserved rock record. The most striking observation is the range in geometries and scales of deformation fabrics (e.g. Choukroune et al., 1997). Just as at the present day, such a diversity implies a tectonic configuration with a range of different types of boundary (i.e. extensional, compressional, strike-slip, and combinations of them). An obvious question to ask is whether any fabrics or structures are prevalent in Precambrian terranes that are not forming at the present day, to reveal whether there has been a significant evolution in tectonic style.

The 'dome and keel' structures of a subset of Archean terranes have previously been proposed to have a non-uniformitarian origin (e.g. Van Kranendonk et al., 2004; Francois et al., 2014). A detailed structural analysis of these features is beyond the scope of this paper. However, we direct readers towards publications that have demonstrated that equivalent structures form in the more recent geological past, by a combination of diapirism, folding, and/or core complex formation (e.g. Marshak et al., 1992; Yao et al., 2021; Kusky et al., 2021). In the absence of any clear difference in deformation styles between the Precambrian and the present day, we instead turn our attention to the question of whether the similarity with the present day places any constraints upon Precambrian rheology or tectonics. We initially focus on the earliest available widespread record, from the Archean, and discuss below whether there are any significant changes between that time and the present day.

One of the structural styles seen in multiple large (i.e. hundreds of kilometres in extent) tracts of Archean terranes is pervasive flat-lying fabrics. This crustal-scale structure is interpreted to represent thickening by simple shear and the stacking of thrust slices, and is best seen in seismic reflection images. Examples include: the Eoarchean to Palaeoarchean Narryer Terrane of Australia (Sellars et al., 2022), which underwent granulite to upper-amphibolite facies metamorphism and partial melting in the Eoarchean, before subsequent deformation and metamorphism throughout the Archean (Kinny and Nutman, 1996); the Palaeoarchean eastern Kaapval craton (de Wit and Tinker, 2004), metamorphosed and deformed in the Palaeoarchean (Wang et al., 2020); the Meso- to Neo-Archean Yilgarn craton core (Drummond et al., 2000), deformed and metamorphosed in the Neoarchean (Goscombe et al., 2019); the Meso- to Neo-Archean Baltic (Karelia) craton (Mints et al., 2009), metamorphosed and deformed in the Neoarchean (Holtta et al., 2016); and the range of Archean and Proterozoic components of the Canadian Shield (Clowes et al., 1996), with metamorphic and deformation ages reaching from the Eoarchean onwards (e.g. David et al., 2009; Iizuka et al., 2007). These observations place constraints on the nature of Archean continental tectonics because of the situations in which such fabrics can be produced (which are found throughout Earth history, up to the present day). Although many tectonic settings can locally form flat-lying fabrics, to do so over horizontal dimensions of hundreds of kilometres is only thought to be possible in two settings. The first is during large amounts of horizontal extension and vertical thinning (e.g. Dewey et al., 1993). However, this process is unlikely to explain the flat-lying fabrics seen in many of the Archean continental interiors, as these fabrics were produced coincident with thickening (as indicated by the pressures experienced by the metamorphic rocks; e.g. Harley, 1989), and the preserved fabrics contain structures associated with horizontal convergence (e.g. folding and thrust faulting). The only known compressional setting in which flat-lying fabrics are formed over large horizontal lengthscales is during the underthrusting of rigid foreland crust beneath a mountain belt (Figs. 2 & 6b; Section 2.1 and Appendix A). In this situation, vertical planes deform by simple shear above the rigid underthrusting foreland material (Section 2.3; e.g. McKenzie et al., 2000; Copley et al., 2011b). The presence of such fabrics therefore places constraints on the timing of formation of large 'plate-like' regions of rigid continental lithosphere, which are able to provide such a rigid lower crust within a mountain belt; where absent, vertical planes would instead thicken by pure shear. As discussed in Section 2.1, crust that is able to remain strong at the temperatures experienced within the deep crust of mountain belts is likely to be anhydrous. The examples given above imply that some such rigid and anhydrous regions were present by at least the early Archean, consistent with the results presented in Section 4.2. Given the difficulty of relating the age of metamorphic samples exposed at the surface to the evolution of an entire crustal section, and the difficulties of seeing through subsequent metamorphic events, not all flat-lying fabrics within the preserved crustal sections necessarily formed that early, especially given that not all rocks will be involved in mountain-building events soon after they form. However, a pattern emerges that from the early Archean onwards there was plentiful deformation of vertical planes by simple shear, implying the presence of strong underthrusting regions of crust. This logic is consistent with the curvature of *P*–*T* loops from Archean belts that imply thickening above rigid lower crust (e.g. Copley and Weller, 2022). We note that this line of logic only applies to regions of flat-lying fabrics on lengthscales of tens to hundreds of kilometres, rather than to individual components of more local structures.

Such a style of deformation also allows us to infer that previous deformation events had affected that crust, during which it become hot enough to partially melt and become anhydrous and strong. The nature of these previous events is not well constrained, although the presence of sub-vertical seismic fast axes in the deep underlying lithospheric mantle at the present day (Priestley et al., 2020) imply that the original thickening may have involved the pure shear of vertical planes, as would be expected to occur before large rheological heterogeneities had developed.

When combined, the available observations therefore imply that pure shear thickening may have been the dominant mode of shortening in the very earliest part of Earth's history (Priestley et al., 2020), before significant partial melting of the crust had occurred to generate significant rheology contrasts. Once those contrasts had been established, the available record (from the early Archean onwards) implies mountain range behaviour equivalent to the present day, characterised by the widespread underthrusting of rigid foreland material on the margins of mountain belts and the resulting stacking of thrust sheets on top of this crust, much like in present-day and recent mountain belts (e.g. Ni and Barazangi, 1984; Lamb et al., 1997; DeCelles et al., 2001; Craig et al., 2012; Searle et al., 2019). It is not known when this style of mountain-building first became dominant, but the evidence for low-temperature hydrous melts preserved by the Hadean Jack Hills zircons (Watson and Harrison, 2005) implies that the initial production of the required rheology contrasts may have occurred very early in Earth history.

4.7. Ancient GPE contrasts

A further important line of evidence regarding the nature of continental tectonics in the past relates to the observed distribution of metamorphic *P*–*T* assemblages. England and Bickle (1984) proposed an elegant line of logic to estimate the relative strength of ancient and present-day continental lithosphere. They noted that the maximum pressures estimated from Archean metamorphic rocks are similar to those experienced at the roots of mountain belts in the recent past. This agreement implies that mountain belt thicknesses were similar in the Archean to now. As discussed in Section 2.1, the limit on the crustal thickness that can be achieved within a mountain belt is provided by the strength of the bounding lowland lithosphere (e.g. Molnar and Lyon-Caen, 1988). Therefore, England and Bickle (1984) proposed that the vertically-integrated lithospheric strength was similar in the Archean to that at the present day.

The database of Kuang et al. (2023) shows metamorphic pressure estimates back to the early/mid Archean, which supplement the compilation of Brown and Johnson (2018). The pressure estimates from 'Barrovian' metamorphic assemblages (Barrow, 1912) typical of regional metamorphism within mountain belts, implies no significant difference in metamorphic pressure compared to present-day mountain belts (Fig. 8f, and discussed in more detail below), albeit with relatively few samples available. This finding in turn implies no major differences between Archean and recent crustal thicknesses within mountain belts. The only significant difference in metamorphic pressure estimates compared to the present day relates to the record of blueschist- and eclogite-facies rocks, which will be discussed below. The argument of England and Bickle (1984) has therefore stood the test of time, and implies minimal changes in lithosphere strength since at least the mid Archean. Fig. 3 and Section 4.2 provide an explanation for this result, based upon other concepts that were not available to England and Bickle

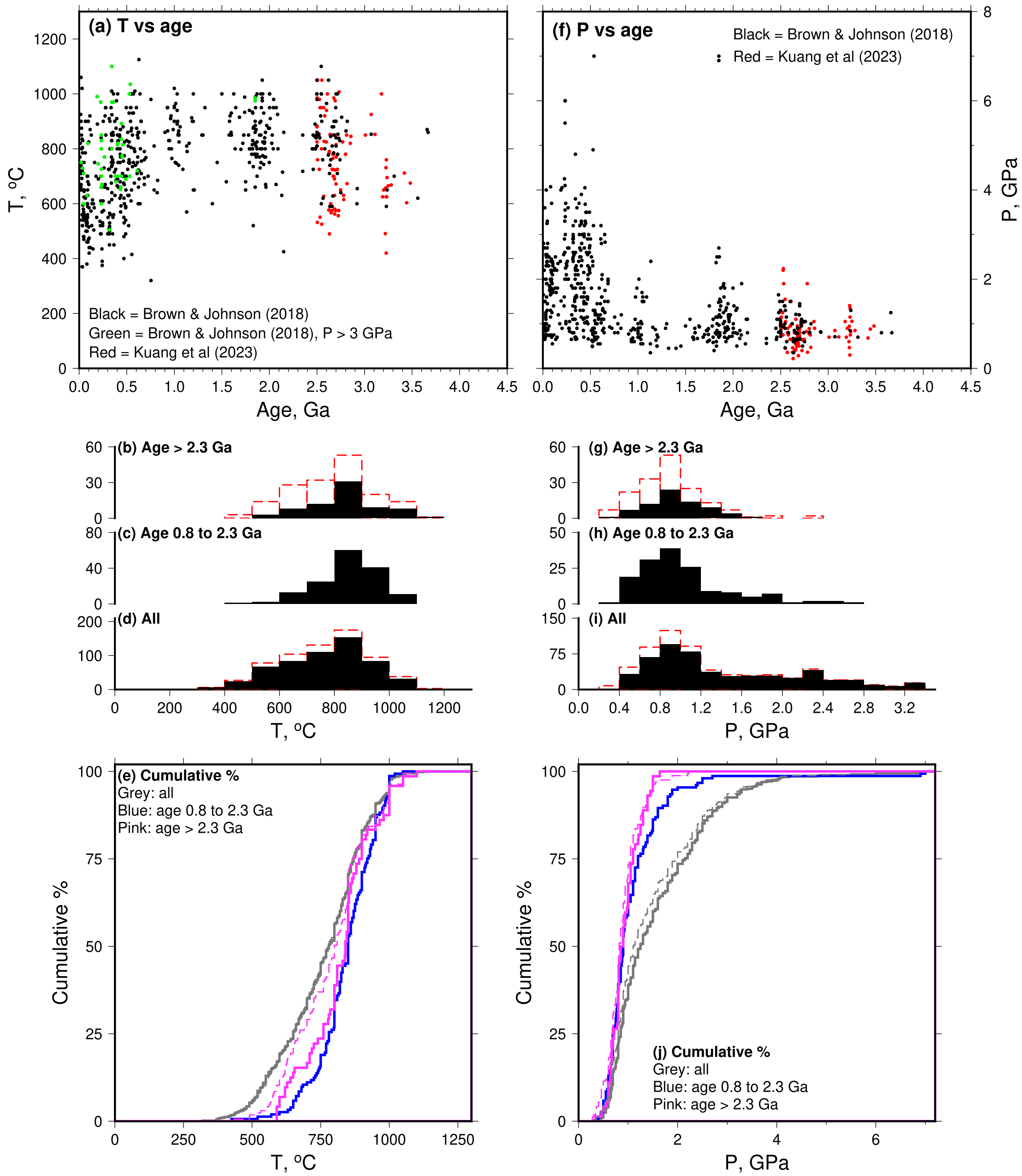

Fig. 8. The distribution of metamorphic pressure (P) and temperature (T) estimates as a function of timing of metamorphism. The left column (a–e) displays T, and the right column (f–j) shows P. On (a) and (f) the black dots are from the compilation of Brown and Johnson (2018), which we retrieved from the supplemental information of Sobolev and Brown (2019), and the red dots are from Kuang et al. (2023) (not including the entries labelled as being from Brown and Johnson, 2018). On (a), the green points are those with pressure estimates greater than 3 GPa. The centre panels (b–d and g–i) show histograms split into different age bins. Solid black bars show values from the Brown and Johnson (2018) dataset, and dashed red lines show the effects of adding the Archean compilation of Kuang et al. (2023). The lower panels (e and j) show empirical distribution functions, divided according to age. Solid lines show the catalogue of Brown and Johnson (2018), and dashed lines also include Kuang et al. (2023) (on (e) the dashed grey line almost entirely overlaps the solid one). The age divisions at 0.8 Ga and 2.3 Ga isolate the effects of preserved high-P rocks in recent times, and separate the Archean and oldest Proterozoic record from the remainder of the dataset at a time with few data points. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(1984) at the time they were writing, specifically related to the controls on the strength of the continental lithosphere.

5. Synthesis and outlook

As far back as the early Archean (i.e. Eo- to Paleo-Archean), the available indicators of deformation fabrics and P–T conditions, when combined with our modern-day understanding of continental rheology and dynamics, implies that the preserved regions of the continents were undergoing continental tectonics in a manner equivalent to that at the present day. Two linked changes through time, the rate of radiogenic heating and the production of voluminous TTG intrusions, do not require continental tectonics to have been distinct from the present day; as discussed above, we instead suggest that TTG petrogenesis represents the operation of modern-style continental tectonics in the presence of elevated radiogenic heating. The calculations presented above demonstrate that the interplay of melting and rheology limits the degree to which elevated heating would have changed the rheology or deformation of the continents compared to the present day.

The 'squishy lid' deformation suggested by Cawood et al. (2022) for the early Archean is in some ways the equivalent of what we refer to as modern-style distributed continental deformation. However, we instead suggest that 'modern-style' continental tectonics is a more appropriate term, which captures the key features discussed above of: the contrasts between plate-like rigid regions and those undergoing spatially-distributed deformation (Sections 4.2, 4.5, & 4.6); the similar thicknesses, rheologies, and metamorphic conditions in mountain belts (Sections 4.5 & 4.7); and the important dynamic consequences of rigid forelands underthrusting mountain belts (Sections 2.1 & 4.6). Alongside the mechanical and rheological arguments presented above, implying the unlikelihood of stagnant lid or sagduction-style behaviour (Sections 4.3 & 4.4), the available information points towards Archean continental tectonics resembling that at the present day.

However, the issue of the representativeness of the available information is an important one. With only localised fragments of information, it is important to ask whether modern-style continental tectonics was also localised within certain regions. It is unlikely that a chance sampling of the geology would result in only the preservation of an atypical style of behaviour, but there is the possibility that preservation potential is governed by tectonic style, i.e. that the preserved regions may have been preserved because they were acting in an atypical manner. Our arguments regarding the unlikelihood of stagnant lid or sagduction-style behaviour are unaffected by the patchiness of the record, as they only rely upon the presence of hydrous rocks, and our ability to estimate densities and rheologies (Sections 4.3 & 4.4). There is no reason why mountain belts generated by pure-shear thickening would be any less likely to be preserved than those generated by simple shear, and the available information regarding mountain belt deformation style (Section 4.6) and thickness (Section 4.7) is dominantly provided by high-grade metamorphic rocks from deep structural levels, which are most likely to be preserved. We therefore suggest that the available information is representative of continental tectonics in general. However, we note that future discoveries of a different style of behaviour in hitherto unrecognised early Archean rocks, or new constraints on the geometries or dynamics of continental deformation belts from the geochemical record, could require a re-evaluation of whether the behaviour we have suggested applied to all continental regions present at that time. We also emphasise that we have solely dealt with the continental record, rather than the (in our view) more difficult question of the timing of onset of modern-style subduction, and the extent to which that was a global or local phenomenon (e.g. Shirey and Richardson, 2011; Cawood et al., 2018; Palin et al., 2020).

5.1. The Hadean and Archean

It is natural to question what came before the Archean tectonics discussed above. The lack of large-scale exposures of pre-Archean rocks limits our ability to use the methods described above to probe continental tectonics in the Hadean. Although we have suggested that the oldest preserved rocks indicate the operation of modern-style continental tectonics, clearly the Earth must have been behaving differently immediately following accretion and impacts. However, some conclusions can be drawn by combining the patchy available observations with physical models. The existence of a (near-)surface hydrosphere in the Hadean (e.g. Mojzsis et al., 2001) makes the operation of a 'stagnant lid' tectonic configuration unlikely, because of the weakening role that water has on fault zones, both directly through pore fluid pressures and indirectly through the production of weak phyllosilicates (Copley, 2018). In this sense, it is possible to view the weakness and deformability of the continents as the natural result of the conversion of a wide range of minerals to phyllosilicates in the presence of water.

A likely trajectory for the tectonic evolution of the continental lithosphere on the early Earth therefore involves a relatively brief initial period in which the lithosphere was too weak to contain rigid continental blocks. At this time, the hot temperatures and resulting weakness of the crust would be likely to produce a tectonic configuration similar to skin on custard, pervasively deforming in response to any thickness contrasts that are created by thickening or intrusion, and to flow in the underlying Earth (Fig. 9a). The weakness of the crust would limit the size of topography and crustal thickness contrasts that could be generated and supported. However, once partial melting has begun, resulting in the generation of anhydrous rocks, then the mosaic of rigid crustal blocks and intervening weak regions would have resulted in continental deformation by the early Archean resembling that at the present day (Fig. 9b & c). The lack of direct geological record from non-continental regions limits our ability to make general statements about the tectonics of the entire planet, which is why we specifically note that it is only on the continents that we infer modern-style tectonics has operated since the early Archean, and possibly the Hadean.

5.2. The Archean onwards

Although the style of continental tectonics was set by at least the early Archean, the geological products have varied through Earth history. Initially, the paucity of rocks that have undergone a cycle of melt depletion and cooling to form a strong anhydrous residue would have left much of the continents too weak to have supported significant elevation contrasts (Section 2.1; Fig. 9a), limiting the spatial extent of further melt production and the formation of strong continental cores. However, as melt extraction gradually strengthened regions of the continents, mountain belts with thicker crust could be produced (Fig. 9b & c). Crustal thickness is one of the dominant parameters that controls the temperatures attained during mountain-building (England and Thompson, 1984; Copley and Weller, 2022), so stronger forelands would lead to thicker mountain belts, hotter temperatures, and more melting. The pace of craton formation, and TTG petrogensis, is therefore likely to have increased (Fig. 9d). The eventual brake would have been applied by the decreasing rate of radiogenic heating leading to a reduction in the temperatures experienced within mountain belts, and in the volume of material undergoing partial melting. The present day represents the continued evolution of this situation, where widespread crustal melting in mountain belts generally only occurs in the largest ranges (e.g. with crustal thicknesses $\geq$ 60 km and lifespans of tens of millions of years; Weller et al., 2021). Fig. 6 shows that it is unlikely that the rates of deformation in the Precambrian were vastly different to those at the present day, and for a given configuration of forces are likely to have been within a factor of $\sim$2 of present-day values (which themselves vary by over an order of magnitude). However, caution must be exercised when inferring these rates from strength distributions alone, as the uncertain nature of the non-continental parts of the Earth means that the forces driving the deformation of the lithosphere are not well known.

Significant attention has been devoted to the evolving geological and geochemical record from the Archean onwards, and what it implies about the evolution of the solid Earth, atmosphere, and hydrosphere. Given our suggestion above, regarding the modern-style behaviour of continental tectonics throughout this time interval, it is worth examining the previously-suggested changes through time. We have discussed above the transition from the production of TTG-dominated granitoid suites to more potassium-rich compositions from the Archean onwards (e.g. Moyen, 2020b). This transition was accompanied by the general evolution of element ratios in sedimentary rocks that have been interpreted to represent derivation from dominantly mafic rocks in the early- to mid-Archean, followed by more intermediate sources (e.g. Smit and Mezgar, 2017; Chen et al., 2020). Such a pattern is consistent with the erosion of a progressively-differentiating continental crust, in the manner discussed above, or with the progressive decrease in mantle-

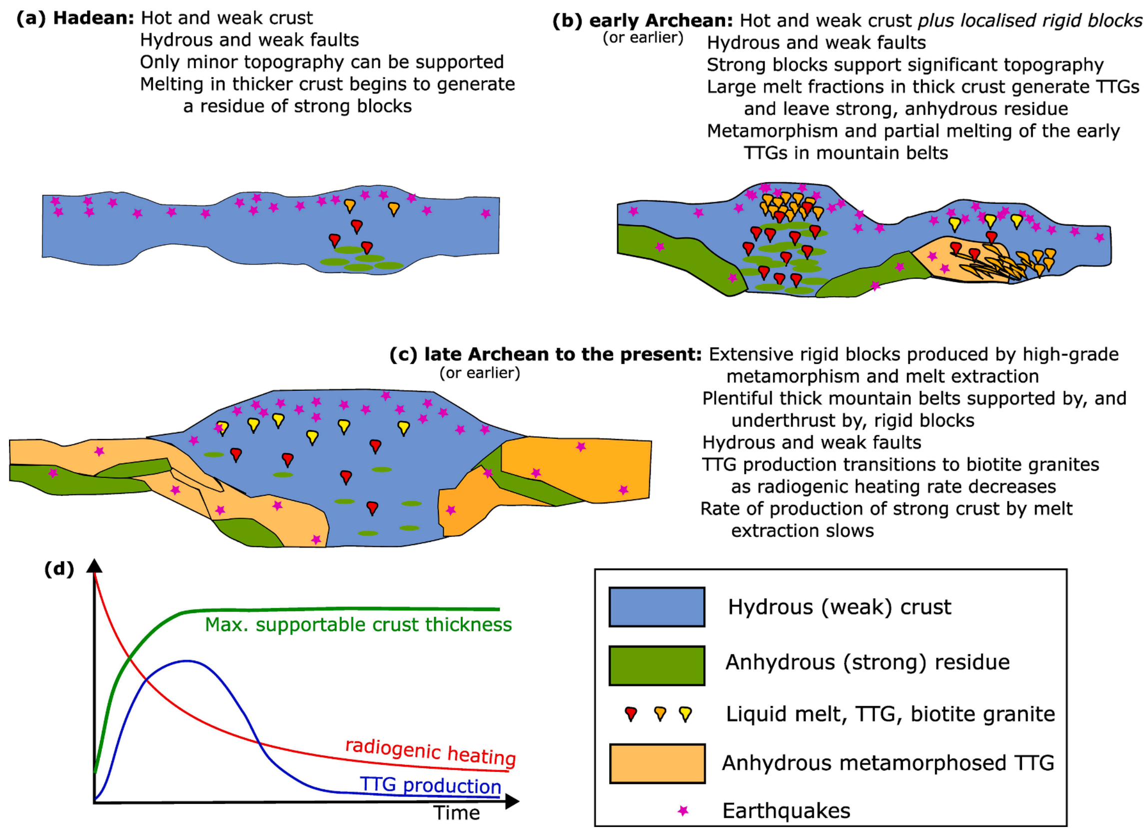

Fig. 9. Schematic diagram of the evolution of Precambrian continental deformation and TTG petrogenesis.

derived basaltic melts as the mantle potential temperature wanes, without requiring a transition in the underlying nature of the tectonics. As noted by Keller and Schoene (2018), secular changes in the compositions of mafic rocks are also consistent with decreases in mantle melt fractions due to a declining potential temperature, without requiring changes in the underlying tectonic processes.

A major research enterprise has involved the interpretation of the temporal evolution of metamorphic *P*–*T* estimates from the perspective of changes in underlying tectonic regimes (e.g. Brown and Johnson, 2018; Holder et al., 2019; Palin et al., 2020; Brown et al., 2022; Kuang et al., 2023). Fig. 8 highlights a number of features relevant to this topic, and shows the currently-available datasets divided into three time periods. The age divisions on Fig. 8 at 0.8 Ga and 2.3 Ga were chosen to isolate the preserved high-*P* rocks in recent times, and separate the Archean and oldest Proterozoic record from the remainder of the dataset at a time with few data points. We display *P* and *T* estimates separately because of the non-linearity of geotherms with depth making the interpretation of average gradients unclear (e.g. England and Thompson, 1984; Copley and Weller, 2022). We avoid splitting the dataset into different classes of temperature gradients because although these may have a petrological basis they do not uniquely correspond to tectonic regimes (except at the very lowest thermal gradients, but below the threshold commonly applied to distinguish 'low dT/dP' gradients). Our analysis is therefore distinct from previous studies that have focused on changes in apparent thermal gradients, with our approach instead focused on examining the distributions temperature and pressure through time to determine if there are any important features.

Temperature estimates from the period 0.8–2.3 Ga are effectively indistinguishable from those older than 2.3 Ga in the compilation of Brown and Johnson (2018) (Fig. 8a,b,c,e). Those younger than 0.8 Ga show more low-temperature estimates. If the compilation of Archean *P*–*T* estimates of Kuang et al. (2023) is included, the > 2.3 Ga distribution begins to resemble the entire dataset more closely (Fig. 8b,d,e).

Taken together, these features imply limited changes in the thermal regime of the continents through time, and indicate an under-sampling of low-temperature metamorphism at ages > 0.8 Ga in the dataset of Brown and Johnson (2018). The metamorphic pressure estimates show the well-known contrast in prevalence of $P > 3$ GPa metamorphic rocks from the periods before and after ~0.8 Ga (Fig. 8f,g,h,i), which the green points on Fig. 8a show is not responsible for the sparsity of low-T rocks in the > 0.8 Ga part of the database of Brown and Johnson (2018). The rarity of these ancient high-*P* rocks is a feature we interpret to be preservation bias, as a result of the typical position of emplacement of such rocks (e.g. eclogites) in orogens at high structural levels (which are prone to subsequent erosion), and due to the propensity of high-pressure assemblages to retrogress and/or behave metastably at high pressures (e.g. Peterman et al., 2009; Palin et al., 2017). If only the record older than 0.8 Ga is examined (Fig. 8g,h,j), there is no significant difference through time in metamorphic pressure estimates, with the only major feature in the dataset being the reduction in sample numbers further back in time. Finally, we note that the level of bias introduced into the dataset by choices of field area by metamorphic petrologists, and the choice of which datasets are used (e.g. whether Kuang et al. (2023) is used to supplement Brown and Johnson (2018), which adds more data but imposes an age-based difference in dataset methodology) cautions against the use of statistical tests, as the underlying distribution is not sampled in a statistically-robust manner. However, those who take the alternative viewpoint may note that the appropriate non-parametric statistical test (the Kolmogorov–Smirnov test) implies that on Fig. 8e the measurements that comprise the blue and pink curves can be produced by sampling from the same underlying distribution, as can the grey and dashed-pink curves, and on Fig. 8j the blue and pink curves.

Based upon the above logic, we therefore suggest that the patchy metamorphic record can be summarised and explained with three concepts: (1) that the maximum pressures of regional metamorphism in mountain belts show no significant changes through time, despite the patchier sampling of the Archean record due to the limited tracts of exposed rocks, implying consistent crustal thicknesses in mountain belts since the emergence of this record in the early- to-mid Archean ; (2) that the absence of eclogites older than 2.5 Ga (Loose and Schenk, 2018; Ning et al., 2022) and blueschists older than 0.8 Ga (Maruyama et al., 1996) is the result of preservation bias; and (3) that, although not analysed above due to the non-linearity of geotherms with depth, apparent changes in apparent thermal gradients through time (e.g. Brown and Johnson, 2018) are the result of the tectonically-imposed inconsistent sampling of ancient mountain belts (due to subsequent erosion and deformation; e.g. Weller et al., 2021).

The Archean and early Proterozoic sedimentary record, and the associated isotopic evolution of the constituent rocks, also presents an avenue for interrogating the prevailing tectonics when the rocks were deposited, including with reference to the degree of freeboard (i.e. the degree of emergence of landmasses from any oceans that were present). The preserved sedimentary record contains rocks deposited in a wide range of environmental settings (e.g. Eriksson et al., 2013; Cosgrove et al., 2024). That some of the sequences are marine and lacustrine is unsurprising given the concept of accommodation space, and the resulting prevalence of subaqueous sedimentary rocks in the better-constrained present day and recent past (e.g. Nyberg and Howell, 2015). The preservation of alluvial, aeolian, and glaciogenic sequences from the Palaeoarchean onwards (e.g. Cosgrove et al., 2024) shows that at least some subaerial landmasses were present by that time. The presence of coarse-grained siliciclatic rocks back to at least the Eoarchean (e.g. Nutman et al., 2017) implies the existence of at least local topography, resulting in high-energy transport and reworking. Whether the mountain belts we suggest were present from at least the early Archean were exposed subaerially, or covered by a thick ocean, depends upon the poorly-constrained total volume of water, and its partitioning between the Earth's surface and subsurface environments (e.g. Marty and Yokochi, 2006). The interpretation of the sedimentary Sr isotopic record is controversial, due to the competing influences of hydrothermal circulation and continental weathering, but it has been suggested to imply a high flux of weathered continental material to the oceans by the Palaeoarchean (e.g. Satkoski et al., 2016) or Eoarchean (e.g. Roerdink et al., 2022), which would require subaerial exposures of significant continental landmasses by that time. Thick and relatively undeformed sedimentary sequences are interpreted to mark the presence of a rigid substrate. They date to at least the Mesoarchean, and become generally more prevalent through time, although with the same periodicity as is seen in the pulsed 'supercontinent' cycle in the detrital zircon record (e.g. Bradley, 2008; Cawood et al., 2018). These sequences are interpreted as passive margins and intra-cratonic basins. Given that passive margins are, by their very nature of being continental margins, susceptible to deformation, metamorphism, and being lost from the geological record following closure of ocean basins, this record is likely to represent a latest possible time interval of the onset of passive margin sedimentation.

Taken together, the above information is consistent with our suggestion of the underlying nature of conti- nental tectonics being equivalent to the present day from at least the early Archean. Where identifiable changes occur (e.g. the onset of passive margin preservation, the evolving chemistry of igneous and metamorphic rocks), they are compatible with the changing potential for preservation and petrogenesis (relating to potential temperature and the rate of radiogenic heating) of the rocks involved, without there being any requirement for the underlying nature of continental tectonics to be different. This concept is supported by the above analysis of the deformation fabrics preserved by Archean and Proterozoic deformation belts. Such a view is consistent with the pulsed nature of the metamorphic record, the detrital zircon record, and the sedimentary basin record (all described above), as the very nature of plate motions involve waxing and waning rates of continental deformation, and an ever-changing balance between compression, extension, and transcurrent motions, depending on the evolving configuration of the global plate circuit (e.g. Merdith et al., 2021).

5.3. Future directions and alien worlds

Avenues for further progress in this field are presented by further interrogating the patchy geological record from the early Earth. The curvature of *P*–*T* loops has shown promise for interpreting the characteristics of the deformation event that caused metamorphism (Copley and Weller, 2022), so the application of increasingly sophisticated metamorphic techniques to ancient samples can be expected to provide additional constraints regarding the nature and tempo of ancient tectonics. However, caution should be exercised when interpreting the ancient record. Plentiful text has been devoted to the absence of perceived 'signatures' of various processes in the rock record, such as the eclogites commonly used to infer deep underthrusting (e.g. Brown, 2023). However, the geological record is far from complete, and importantly its nature and preservation depends upon tectonics – the very process for which we are aiming to interrogate it. The unsatisfactory nature of this situation is exemplified by eclogites. Their upper-crustal position in active orogens, and their susceptibility to retrogression, mean that is it highly unlikely that they will be found in deeply eroded ancient orogens (Weller and St-Onge, 2017; Weller et al., 2021), and their presence back to 2.5 Ga is itself a remarkable testament to detailed mapping and the ability to infer past events through a retrograde overprint (Loose and Schenk, 2018; Ning et al., 2022). It is much safer to interpret the observations we have available than to base a point of view on the absence of signatures that are easily destroyed by the process that they are taken to be a signature of. This procedure must itself take account of the spatial scales of tectonic deformation, and which signals are local, regional, or global.

A wider implication of considering the conditions required for stagnant lid tectonics, in tandem with the strength of active faults, lies in the burgeoning field of exoplanet science. The identification of planets from other solar systems has driven an increasingly large number of modelling studies (e.g. Guimond et al., 2022; O'Neill et al., 2020a; Stern et al., 2018). At their heart such studies require assumptions to be made regarding the tectonic configuration of those planets in order to address questions regarding the composition of, and coupling between, the solid components and the liquid or gas envelopes of the alien worlds. For other planets to posses a 'stagnant lid' would require one of two conditions. First, an absence of water at the surface and in the shallowest tens of kilometres of the crust could result in rocks with a brittle strength much higher than that of faults on Earth, and therefore able to sustain a single global plate. Second, the planet could be cool enough, either due to being small, or due to being deficient in radiogenic heating, that the seismogenic layer could be incredibly thick and the mantle convection very sluggish. Venus represents a possible example of the first situation (although the chasmata structures imply some tectonic activity is or was present; Nimmo and McKenzie, 1998), and Mars an example of the second. Any 'Earth-like' planet that has the possibility to sustain life, based upon liquid water at the surface, is therefore likely to be experiencing active lithosphere deformation. This concept poses problems for the modelling of such worlds, given the difficulties in modelling tectonic processes without plentiful observational constraints. The Earth will continue to be the prime laboratory for studying planetary behaviour.

6. Conclusions

The available observations, when combined with our physical and chemical understanding of continental dynamics, suggests that continental tectonics has resembled that at the present day since at least the early Archean. Although it is temping to interpret the patchy geological record from the early Earth from the perspective of exotic tectonic processes, uniformitarian behaviour (which, over large parts of the Earth's continental surface, does not resemble plate tectonics) is indicated by all available observations. The only discernible temporal changes relate to the petrogenesis of igneous rocks under slightly different ambient temperatures, primarily related to higher rates of radiogenic heating, rather than to fundamentally different tectonic styles. Although the nature of Hadean continental tectonics is made ambiguous by the lack of geological record, the presence of a hydrosphere at such times rules out a 'stagnant lid' mode of behaviour. It is more likely that an early weak crust, resembling pervasively deforming skin on custard, rapidly evolved into a modern-style configuration as melt extraction occurred.

CRediT authorship contribution statement

Alex Copley: Writing – review & editing, Writing – original draft, Software, Methodology, Formal analysis, Conceptualization. **Owen M. Weller:** Writing – review & editing, Writing – original draft, Methodology, Formal analysis.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

No data was used for the research described in the article.

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Appendix A. Present-day forces, rheology, and continental dynamics

This Appendix describes our present-day understanding of continental tectonics in a more detailed manner than in the main paper, with the aim of providing a useful summary to those interested in comparing the geological record to present-day deformation.

A.1. Forces

At the present day, gravity acting on lateral density contrasts in subduction zones and at mid-ocean ridges result in forces (often termed 'slab pull' and 'ridge push'; Fig. 2; e.g. Forsyth and Uyeda, 1975) that probably have roughly equal importance in driving the motion of the lithosphere, once local resisting forces in each of these settings are taken into account (e.g. Copley et al., 2010). The uncertain age, structure, behaviour, and mode of recycling into the mantle of the Precambrian equivalent of today's oceanic lithosphere limits our ability to consider how these forces may have evolved through time. Calculations imply that the negative buoyancy of oceanic lithosphere produced from hotter mantle may have been similar to that at the present day, because of the counterbalancing of thermal and compositional effects (Weller et al., 2019). However, the unknown thickness of the ancient oceanic mantle lithosphere presents major uncertainties (e.g. Solomatov, 1995; Korenaga, 2006; Hynes, 2013), even though the thickness of crust produced by adiabatic decompression melting reaching to the (near-)surface can be inferred to be in the range 7–25 km for mantle potential temperatures of 1300–1550 °C (e.g. McKenzie and Bickle, 1988; Weller et al., 2019).

The magnitude of one of the important forces driving present-day continental tectonics can be inferred beyond the age of preserved oceanic lithosphere. Crustal thickness contrasts, even when isostatically compensated, result in lateral gradients in gravitational potential energy (GPE), and so a 'buoyancy force' being exerted between mountains and their adjacent lowlands (Fig. 2; e.g. Artyushkov, 1973; England and McKenzie, 1982; Molnar and Lyon-Caen, 1988). This force can be colloquially thought of as that required to hold up the mountains, or that exerted by the mountains trying to spread out under their own weight. The magnitude of the buoyancy force exerted between mountains and the bounding lowlands depends upon the lateral contrasts in density structure in all material above the depth of isostatic compensation. At the present-day, for large mountain belts (e.g. the Tibetan Plateau and the Andes), the buoyancy force per unit length along-strike of the mountain belt is likely to be larger than any other force acting on the bounding plates, by possibly up to a factor of two (e.g. Dalmayrac and Molnar, 1981; Molnar et al., 1993; Copley et al., 2010; Wimpenny et al., 2018). However, convergence continues in these regions because the force balance in the lithosphere is intrinsically three-dimensional. In the case of southern Asia, the large force per unit length arising from the region of the Tibetan Plateau is balanced by the smaller forces per unit length being exerted on all other parts of the Indian and Eurasian plates (e.g. Copley et al., 2010). When considering the force balance on the lithosphere in the early Earth, this finding indicates that it is important to think three-dimensionally, rather than along cross-sections through the Earth. Following England and Bickle (1984), we discuss in the main paper the information that allows us to infer the sizes of crustal thickness contrasts in the Precambrian, and so the likely magnitude of the resulting buoyancy force.

A further important implication of buoyancy forces, which is discussed in the main paper, is related to the size of GPE contrasts that can be supported by the lithosphere. As discussed by Molnar and Lyon-Caen (1988), a mountain belt can only continue to grow in elevation and crustal thickness if the resulting GPE contrast with the surrounding lowlands can be supported by the strength of the lithosphere. Once a limiting GPE contrast has been reached, a given mountain belt can only expand laterally, along- and/or across-strike, rather than increase in thickness. In this sense, mountain range elevations can act as a 'pressure gauge' for the stresses that are able to be transmitted through the lithosphere (Molnar and Lyon-Caen, 1988). This link between the strength of the lowlands bordering a mountain belt, and the crustal thickness of the range itself, has important implications for understanding continental tectonics in the Precambrian. In particular, because the thickness of the crust is one of the dominant controls on the temperatures attained (e.g. England and Thompson, 1984; Copley and Weller, 2022), the relationship between the strength of the bounding lithosphere, mountain belt crustal thickness, and the petrogenesis of igneous and metamorphic rocks is an important one.

A.2. Lateral variations in behaviour

The behaviour of the lithosphere in response to the forces being exerted upon it depends on its rheology. As we will describe in this section, at the present day large lateral variations in rheology lead to striking contrasts in lithosphere motions and deformation. In the oceans, and large areas of the continents, the lithosphere acts as a small number of rigid plates separated by narrow deformation bands along their edges (Fig. 1), and in these regions the geometrical rules of plate tectonics apply (McKenzie and Parker, 1967; Le Pichon, 1968). However, an important concept arose in the immediate aftermath of the early publications about plate tectonics: in significantly-sized regions of the continents, the geometrical principles of plate tectonics do not apply (e.g. McKenzie, 1972; Molnar and Tapponnier, 1975; Dewey et al., 1986; England and Jackson, 1989). Rather than the motions being characterised by a small number of rigid plates with narrow deformation belts along their edges, regions with dimensions of thousands of kilometres, comparable to the smaller oceanic plates, deform in a spatially-distributed manner. For example, this feature is spectacularly seen in the wide belts of earthquakes, and the mountain belts and basins produced by recent faulting, that stretch through Europe and Asia, and along the western margin of the Americas (Fig. 1; e.g. England and Jackson, 1989).

An alternative to plate tectonics was required to study these zones of distributed continental deformation, and the subsequent debates gave rise two competing conceptual models: 'continuum' models, and 'microplate' or 'block' models. The continuum approach (e.g. Bird, 1978; England and McKenzie, 1982; Vilotte et al., 1982) emphasises the distributed nature of the deformation, and that some active deformation belts contain many faults, which are small and slow-moving when compared with the size of the deformation belt and the total deformation rates. The crust and lithospheric mantle are then modelled as a continuously-deforming medium, using the principles of fluid dynamics. The block modelling approach (e.g. Avouac and Tapponnier, 1993; McCaffrey, 2005; Meade and Hager, 2005; Reilinger et al., 2006) instead applies the concepts of plate tectonics, and separates deformation belts into rigid blocks, the motion of each of which can be described using a rotation pole. The debates between these approaches became increasingly polarised, with the models often being characterised as mutually exclusive. However, the analysis of earthquake locations and mechanisms, GPS velocities, and the records of past deformation preserved in the geological record, have made it clear that the most appropriate model depends on the location, date, timescale, and lengthscale being considered, and the scientific question that is being posed (e.g. England and Jackson, 1989; Thatcher, 2009). Both approaches have clear merits when applied correctly. For example, central Anatolia can be usefully thought of at the present day as a rigid block rotating about a rotation pole, with the motion relative to Eurasia mostly accommodated by the North Anatolian Fault: a strike-slip fault moving at a rate of ~2.5 cm/yr, and approximately following a small circle about the rotation pole (e.g. McKenzie, 1970; Reilinger et al., 2006). In contrast, in the past central Anatolia was pervasively deformed (e.g. Sengor and Yilmaz, 1981), and the deformation probably more closely resembled the present-day Turkish-Iranian Plateau to the east, characterised by many distributed and slow-moving faults accommodating a spatially smooth velocity field (e.g. Copley and Jackson, 2006; Reilinger et al., 2006), and therefore amenable to considering at the large scale as a continuously-deforming region, although this approach clearly doesn't preclude also examining the individual faults within that region.

When considering both approaches to studying regions of distributed deformation, in the contexts for which they are each most appropriate, some important concepts arise, including: (1) deformation belts contain many active faults, the locations, slip directions, and slip rates of which cannot be predicted from knowledge of the motions of the bounding plates; (2) deformation evolves in space and time within regions of distributed continental tectonics, including areas switching between extension, compression, and strike-slip deformation, or becoming (or ceasing to be), undeforming; (3) temporal and spatial variations in deformation are caused by the evolution of the driving forces (e.g. as crustal thickening or thinning change the magnitude and orientation of the GPE contrasts), of the rheology (e.g. due to temperature changes, melt extraction or addition, and hydration or dehydration), and of the geometry (e.g. due to the propagation of a mountain belt over a rigid bounding plate); (4) geological history plays an important role in present-day tectonics, from the perspective of imposing a template of pre-existing rheology contrasts upon a deformation belt; and (5) tectonic configurations can change on small scales, sometimes rapidly, and some deformation belts may not be easily identified today or leave an obvious geological signature (e.g. England and McKenzie, 1982; Sibson, 1985; Molnar and Lyon-Caen, 1988; England and Jackson, 1989; Lamb et al., 1997; Mattei et al., 2004; Allen et al., 2004; McKenzie et al., 2019; Whyte et al., 2021; Penney and Copley, 2021).

Important implications of the above findings for investigating ancient tectonics include an appreciation that distributed continental deformation can only be fully understood by making observations at the scale of hundreds to thousands of kilometres (i.e. the scale of entire deformation belts), that a snapshot of a specific time interval may not reveal all of the important processes that have affected the tectonic evolution of a suite of rocks (some of which may have been overprinted and lost from the record), and that the deformation in smaller regions may not be representative of a deformation belt as a whole. For example, normal-faulting in areas of thick crust can occur either due to the post-orogenic extension of a region (Gaudemer et al., 1988), or during continued convergence as a consequence of a curved geometry of the range edge (Armijo et al., 1986), a change in rheology of the bounding fold-thrust belt (Wimpenny et al., 2018), or changes in convergence rate or internal GPE contrasts (England and Houseman, 1989), and the cause would only become clear by examining the regions beyond the margins of the extending region. Similarly, strike-slip faults may accommodate the straightforward lateral translation of one crustal block relative to another, or may actually represent the extension or compression of a region, accommodated by a combination of faulting and rotations about vertical axes (e.g. Campbell et al., 2013). Put a different way, there are multiple different spatial configurations of deformation that can accommodate a given velocity field (e.g. England and Jackson, 1989), and that overall velocity field can only be inferred based upon observations at the scale of entire deformation belts, which can be difficult to achieve when looking at the fragmentary record of ancient deformation events.

A.3. Causes and consequences of spatial variations in rheology

A natural question that arises from the above discussion is what causes the lateral variations in continental rheology, and so behaviour, ranging from the regions behaving as large, rigid plates, to wide areas of pervasive deformation. Recent decades have seen considerable progress in relating present-day kinematics and dynamics to the thermal and compositional structure of the crust and upper mantle, and geological history, as described below. These advances point towards geological history as a dominant cause of lateral rheology contrasts, which in turn produce lateral variations in behaviour.

Modelling the *P* and *S* waveforms of earthquakes, and their surface reflections, can provide well-constrained estimates of earthquake mechanism and depth that improve upon those generated by global catalogues (e.g. Taymaz et al., 1991; Zwick et al., 1994; Craig et al., 2012). Although this process requires significant human input to select suitable waveforms, pick arrivals, and model the waveforms appropriately, the resulting image of the thickness of the earthquake-prone upper part of the lithosphere (the 'seismogenic thickness') has highlighted the importance of lateral variations in continental rheology. In the relatively undeforming interiors of many continents, the sparse earthquakes that do occur have depths of up to 40–50 km, into the lower crust and, in places, the uppermost lithospheric mantle (e.g. Maggi et al., 2000; Jackson et al., 2008). This depth approximately corresponds to that of the 600 °C isotherm in these regions, which is the same temperature to which earthquakes can occur in the oceanic lithospheric mantle (e.g. Abercrombie and Ekstrom, 2001; Craig et al., 2014; McKenzie et al., 2005). Estimates of the elastic thickness in the same regions, which is a measure of the thickness of the layer able to support long-term elastic stresses, imply that the elastic thickness varies in tandem with, but is lower than, the seismogenic thickness (e.g. Maggi et al., 2000; Jackson et al., 2008). This relation has been used to infer that in the relatively rigid and plate-like regions of the continents, the strength of the lithosphere is dominated by the seismogenic layer, which is also consistent with the stress release in earthquakes in such regions (Copley et al., 2011a). In this situation, the behaviour of the lithosphere is governed by the stresses that can be transmitted across faults, through the seismogenic layer. In contrast, within the Earth's major continental deformation belts, earthquakes are confined to the upper 10–20 km of the crust, roughly corresponding to temperatures of 300–400 °C (e.g. Maggi et al., 2000; Craig et al., 2012). In these regions, the thinner seismogenic layer and the tractions exerted upon the base of that layer by the deforming ductile mid- to lower-crust, mean that the dynamics are governed by the ductile part of the crust, and possibly also the ductile lithospheric mantle (e.g. Lamb, 1994; England and Molnar, 1997; Bourne et al., 1998; Copley and McKenzie, 2007).

The cause of the lateral variations in seismogenic thickness is related to geological history. The regions with a large seismogenic thickness including most or all of the crust are also generally characterised by mid- to high-grade Archean or Proterozoic metamorphic rocks at the surface, and thick lithosphere (which acts to insulate the crust from the convecting mantle; e.g. Jackson et al., 2008). This combination of characteristics results from past periods of mountain-building, during which the crust and mantle lithosphere were thickened, and the crust became hot enough to partially melt due to the amount of radiogenic heating in the thickened crustal column (e.g. McKenzie and Priestley, 2016). Following post-orogenic erosion and cooling, the result is a melt-depleted and (near-)anhydrous lower continental crust in isostatic equilibrium, with its surface close to sea level, underlain by thick mantle lithosphere (e.g. McKenzie and Priestley, 2016). It has been long established that hundreds of parts-per-million of water in nominally anhydrous minerals results in orders-of-magnitude reductions in their creep viscosity (e.g. Mackwell et al., 1998; Hirth and Kohlstedt, 2003). The resulting transition from deformation in earthquakes to that by ductile flow will therefore occur at hotter temperatures, and so deeper for a given geotherm, in rocks that contain a framework of nominally anhydrous minerals that are actually anhydrous on the hundreds of parts-per-million level, explaining the thick seismogenic layer in the cratonic continental interiors. In contrast, the interiors of many deformation belts are composed of accreted island arcs, or rocks that have been fluxed by water during subduction-enabled ocean basin closure before continent–continent collision. The hydrous nature of such terranes results in a thin seismogenic layer, due to the effectiveness of creep deformation. This rheology contrast between cratonic foreland and the hydrous interiors of deformation belts has been revealed not only through variations in seismogenic thickness and elastic thickness, but also through the spatial patterns of postseismic ductile flow at depth (e.g. Liu-Zeng et al., 2020; Wang et al., 2021).

Given that the thickness of the seismogenic layer varies laterally by large amounts, the strength of earthquake-generating faults is clearly an important parameter. It has become clear that faults are considerably weaker than values predicted by 'Byerlee's law' (Byerlee, 1977), due to some (possibly spatially variable) combination of the frictional properties of fault zone minerals (e.g. weak phyllosilicates) and the magnitude of pore fluid pressure (which natural hydrofracture shows can be above lithostatic) (e.g. Lachenbruch and Sass, 1980; Sibson, 2004; Lockner et al., 2011; Copley, 2018). Commonly-inferred fault strengths are on the order of megapascals to tens of megapascals (corresponding to effective coefficients of friction of $<$ 0.1–0.3), and such faults are only able to support the forces commonly acting upon the lithosphere without deforming when the seismogenic layer is thick ($\geq$20 km; Copley, 2018). Therefore, rigid behaviour occurs in regions where past dehydration by melt extraction has increased the thickness of the seismogenic layer. Observations from regions that have remained rigid since the Archean show that the rocks do not need to be completely dehydrated for this effect to occur. For example, the Archean granulite-facies assemblages of the Douglas Harbour Window in Canada remained undeformed through a subsequent Tibetan-scale mountain-building event (the Paleoproterozoic Trans-Hudson orogen) despite containing up to $\sim$20 vol% amphibole (Whyte et al., 2021). In that event, hydration and retrogression from an external fluid source (an overlying fold-thrust belt) caused a $\sim$3 km thick region to retrogress to an amphibolite-facies assemblage and deform sufficiently to develop a pervasive foliation, whereas the only strain affecting the underlying granulite-facies rocks was later shallow doming that left no visible deformation fabric in the rocks. It is possible that small proportions of hydrous phases may actually aid to strengthen the rock, by accommodating all of the available water, embedded within a rigid framework of nominally anhydrous minerals that are also actually anhydrous. The pattern of melt loss resulting in rigid plate-like continental lithosphere can also be seen in the compilation of Cawood et al. (2018), which shows that the Earth's major cratons became stable and undeforming after a period of crustal melting to generate potassium-rich granites, which would leave behind a strong, anhydrous, residue.

The geological implications of the production of strong and plate-like regions of the continents extends beyond the behaviour of those regions themselves. The dynamics of lithosphere deformation depends upon whether rigid material underlies a deformation belt at depth (e.g. McKenzie et al., 2000; Beaumont et al., 2010; Copley et al., 2011b). If rigid material underlies a deformation belt, for example anhydrous and rigid crust of the underthrusting foreland of a mountain belt, then the dynamics of the flow of the crust are dominated by vertical planes deforming by simple shear above this rigid lower boundary. Put more simply, the mountain belt will act in a similar manner to a drop of honey spreading out over a table; the driving forces from GPE contrasts result in the shearing of the deforming crust above the rigid underthrusting base. In this case, the topography forms a flat top and steep front (e.g. Huppert, 1982), as seen on many range-fronts where rigid forelands underthrust mountain belts (e.g. Penney and Copley, 2021), and the surface motions are in the direction down the local topographic slope (e.g. Copley and McKenzie, 2007). In contrast, where there is no rigid material present at depth, then the dynamics are governed by a combination of: (1) any GPE contrasts present; (2) the far-field motions of the plates bounding the deformation belt; and (3) the distribution of rigid blocks on the edges of, and within, the deforming zone (e.g. England and McKenzie, 1982; England and Houseman, 1986; Flesch et al., 2001). In this situation more gentle topographic slopes are formed than for the case where rigid material is present at depth (e.g. England and Houseman, 1986; Penney and Copley, 2021). These different styles of mountain building lead to distinct pressure (P) and temperature (T) histories in associated metamorphic rocks. Analysis of the P-T conditions experienced in the commonly-observed 'Barrovian' metamorphic belts typical of mountain-building implies that for the majority of past mountain belts, thickening occurred in the presence of rigid underthrusting foreland material (e.g. England and Thompson, 1984; Copley and Weller, 2022), consistent with the deformation patterns observed at the present day (e.g. Argand, 1924; Barazangi and Ni, 1982; Molnar and Lyon-Caen, 1988; Lamb et al., 1997; Nissen et al., 2011; Craig et al., 2012).

A further implication from recent work linking lithosphere structure and rheology, which will become relevant below, relates to the stability of the lower lithospheric mantle. Early models that considered the temperature-dependence of lithosphere density predicted the widespread convective instability of mantle lithospheric roots that became thickened by convergence, and their subsequent removal into the convecting mantle (e.g. Houseman et al., 1981; Molnar et al., 1993). However, seismic tomographic models have shown that thick mantle roots are present beneath active mountain belts (e.g. McKenzie, 2006). Such roots can be stabilised from removal by past melting events that have reduced their density due to compositional changes (e.g. Jordan, 1975; McKenzie and Priestley, 2016), an effect which can overwhelm the temperature dependence of the density. Thick lithospheric roots can therefore be stabilised for billions of years, such as those seen beneath cratons (e.g. Priestley et al., 2020). This view of the apparent stability of the lithospheric mantle is consistent with results that indicate the removal of the lower lithosphere is not necessary to produce the patterns of strain and topography seen in present-day mountain belts (e.g. Copley et al., 2011b).

Appendix B. Supplementary material

Supplementary material to this article can be found online at https://doi.org/10.1016/j.precamres.2024.107324.

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