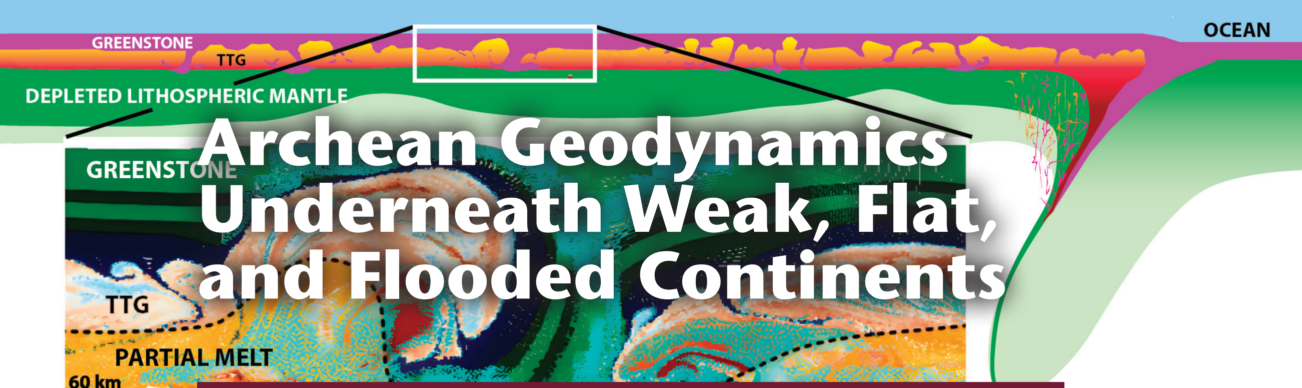
Archean Geodynamics Underneath Weak, Flat, and Flooded Continents



During the Archean dual-mode geodynamics, gravitationally unstable greenstone belts drove intracrustal tectonics, whereas spreading protocontinents forced the subduction of adjacent lithospheres.

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Although a significant volume of crust was extracted from the mantle early in Earth's history, the contribution of felsic rocks to the sedimentary record was minimal until ~3.0 Ga. On a hotter Earth, this conundrum dissipates if we consider that the felsic crust was buried under thick basaltic covers, continents were flooded by a near-global ocean, and the crust was too weak to sustain high mountains, making it largely unavailable to erosion. Gravitational forces destabilized basaltic covers within these weak, flat, and flooded continents, driving intra-crustal tectonics and forcing episodic subduction at the edges of continents. Through secular cooling, this dual-mode geodynamics progressively transitioned to plate tectonics.

Keywords: Archean continents; geological record; geodynamics

INTRODUCTION

Continents exist for hundreds of millions of years and hold the memory of past tectonic and geodynamic processes. The continental crust itself is often presented as the cumulative outcome of plate tectonics because modern crust is typically created, reworked, and recycled at subduction zones. Some proxies for mantle depletion indicate that 50% to 75% of the continental crust had formed by 3.0 Ga (Campbell et al. 2003), which is consistent with the geologic evidence of the existence of pre-3.0 Ga felsic crust. Although this record extends back to 4.0 Ga (Bowring et al. 1989), other proxies indicate that recycling of the oceanic crust only started at 3.8–3.6 Ga (e.g., Kemp et al. 2010; Bauer et al. 2020). How much of the early crust was the product of plate tectonics is therefore uncertain. To advance this debate, we address here two important paradoxes. The first paradox stems from the surprisingly tenuous contribution early felsic sources had to the pre-2.5 Ga sedimentary record (Taylor and McLennan 1995). The second paradox arises because the recycling of oceanic crust left little trace in proxies sensitive to the recycling of sedimentary and weathered rocks until ~2.5 Ga (Valley et al. 2015).

To address the first paradox, we review crustal growth and corresponding mantle depletion, and present a model to explain the late appearance of felsic crust in the sedimentary record. To address the second paradox, we discuss key attributes of Archean (4.0 to 2.5 Ga) crust in the light of numerical experiments. These point to a dual-mode of geodynamics in the Archean. One tectonic mode involved episodic sinking of basalt covers into and through the felsic crust, a process called sagduction (Mareschal and West 1980), that could explain the second paradox. The other tectonic mode involved the gravitational spreading of weak continents, thereby forcing episodic subduction at their margins. While the first mode progressively disappeared because of the secular cooling of the Earth, the second mode evolved into modern plate tectonics.

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PRODUCTION AND EVOLUTION OF THE CONTINENTAL CRUST

Crustal Perspective on Continental Growth

Fine-grained sediments provide global estimates of the average composition of Earth's emerged surface. The composition of Archean fine-grained sedimentary rocks suggests that the emerged crust was dominated by mafic-ultramafic rocks during most of the Archean and transitioned to more felsic compositions no earlier than ~3.0 Ga (Taylor and McLennan 1995). This is supported by the $^{87}Sr/^{86}Sr$ of marine carbonates, which identifies fluxes of weathered felsic crust into the oceans. Marine carbonates show a departure at ~3.0 Ga from the mantle trend they inherited from basalts, towards a more radiogenic trend recording an increased contribution of felsic lithologies to continental runoff (e.g., Flament et al. 2013). This is consistent with the lack of prominent pulses of crust production before ~3.0 Ga (e.g., Condie et al. 2017) recorded by the distribution of ages in detrital zircon, an igneous mineral resistant to erosion and alteration that can be dated using the U-Pb chronometer. Overall, the sedimentary record suggests no significant volumes of felsic crust at the surface before ~3.0 Ga, despite the well-documented presence of pre-3.0 Ga felsic crust in most cratons. In what follows, we discuss what we know about the origin of old felsic crust.

The stable oxygen isotope ratio $^{18}O/^{16}O$, expressed as $\delta^{18}O$, and radiogenic hafnium isotope ratio $^{176}Hf/^{177}Hf$, expressed as εHf (see Toolkit), measured in zircon, are tracers of the source of the melt from which the zircon crystallized. Whereas $\delta^{18}O$ of mantle zircon ranges from 5.5‰ to 5.9‰ today, zircon from sources that interacted with water at low temperature (e.g., sediments and altered crust) have $\delta^{18}O$ values reaching 12‰. The $\delta^{18}O$ of old zircons shows a change from values that remained stable and ≤ 7.5‰ before 2.5 Ga, to increasingly higher values in the Proterozoic (2.5 Ga to 541 Ma) (see Toolkit). $\delta^{18}O$ values between 6.5‰ and 7.5‰ point to a steady-state contribution between surface and mantle material in the source of old zircons. Increasing $\delta^{18}O$ values above 7.5‰ after 2.5 Ga indicate a shift towards a growing contribution of supracrustal lithologies to the production of felsic crust (Valley et al. 2015). Basaltic crust derived directly from partial melting of the mantle inherits its εHf(t). Recurrent partial melting of this crust in a closed system leads to felsic rocks with younger zircons showing increasingly negative εHf(t) values. This trend is observed in the oldest known suite of zircons from the Jack Hills of the Yilgarn craton and from the Acasta gneiss of the Slave craton, i.e., it began well before 3.8 Ga (e.g., Kemp et al. 2010; Bauer et al. 2020). In contrast, the involvement of juvenile magmas (i.e., magmas derived from the mantle or remelting of a short-lived basaltic crust) leads to a shift towards higher εHf(t) values. Such a shift is well-documented and occurs at different times on different cratons, beginning between 3.8 and 3.6 Ga (e.g., Kemp et al. 2010; Bauer et al. 2020). The shift towards higher εHf(t) points to increasing recycling of juvenile basaltic crust, whereas the shift to higher $\delta^{18}O$ points to increasing recycling of surface-altered rocks from 2.5 Ga onwards. Although both shifts might be explained by subduction of altered oceanic crust, the difference in their timing requires an explanation. Because no $\delta^{18}O$ shift is observed at 3.8–3.6 Ga, it is difficult to invoke modern style subduction to explain the εHf(t) shift at that time.

Mantle Perspective on Continental Growth

Over geologic time, the extraction of the crust from the mantle must have resulted in a corresponding depletion of incompatible elements (see Toolkit) in the upper mantle, the region where partial melting occurs most often. For example, the Nb/U ratios of juvenile mafic rocks (e.g., mid-ocean ridge basalts) are characterized by a current value of 47. Assuming Earth's primitive mantle initially had a Nb/U of 30, the increase to an average of 47 must have resulted from partitioning of U into the continental crust (e.g., Hofmann et al. 1986). Basaltic rocks as old as 3.5 Ga have slightly lower Nb/U ratios (43) than those of modern oceanic basalts (47), suggesting ~75% of the crust was extracted from the mantle by 3.5 Ga (Campbell et al. 2003). Further back in time, the U-Pb and Hf isotopic compositions of magmatic zircon older than 3.8 Ga show limited signs of early mantle depletion (Fisher and Vervoort 2018). This is rather puzzling because the 4.4 Ga Jack Hills zircon (Wilde et al. 2001), the Acasta gneisses (4.03 to 3.94 Ga) in the Slave province (Bowring et al. 1989), and small $^{142}Nd/^{144}Nd$ excesses (see Toolkit) in 3.8 Ga mafic rocks of the Isua supracrustal belt (Greenland) suggest that continental crust extraction had begun by 4.3 Ga, before ^{146}Sm went extinct (see Toolkit; Boyet et al. 2003; O'Neil et al. 2024 this issue). This apparent paradox can be resolved if we consider that before 3.8 Ga, large mantle plumes and an intense bombardment of the surface by planetesimal impacts could have efficiently re-mixed the early crust back into the convecting mantle. A model that does not require the dramatic recycling of the early lithosphere (the outermost layer of the Earth above the convective mantle) was recently proposed by Guo and Korenaga (2023). They noted that the depleted mantle is unlikely to melt before being well mixed with the undepleted mantle, which they calculated could take ~700 million years. Hence, the signal of a depleted mantle after 3.8 Ga is not incompatible with crustal growth before 4.0 Ga. Nevertheless, if 50% to 75% of the crust was extracted by 3.5 Ga, one needs to explain why there is little discernible evidence for felsic sources in the pre-3.0 Ga sedimentary record.

RECONCILING THE CRUST AND MANTLE RECORDS: THE WEAK, FLAT, AND FLOODED CONTINENTS HYPOTHESIS

The increasing εHf(t) in zircon beginning at 3.8–3.6 Ga supports an enhanced contribution of basaltic crust to the production of felsic magmas (Laurent et al. 2024 this issue). To explain this shift, it is tempting to invoke the initiation of a plate tectonic regime involving subduction and melting of oceanic crust and sedimentary rocks (Fisher and Vervoort 2018; Bauer et al. 2020). However, this scenario is in apparent conflict not only with the lack of a corresponding $\delta^{18}O$ shift in zircon (i.e., the second paradox), but also with the secular evolution of Archean fine-grained sedimentary rocks, carbonates, and detrital zircon showing limited emergent felsic continental crust before ~3.0 Ga (i.e., the first paradox). To address the first paradox and reconcile early growth from 4.0 to 3.0 Ga with the quasi-absence of felsic sources in the sedimentary record, one needs to revisit the implicit assumption that the early felsic continental crust was above sea level, and therefore available to influence the composition of detrital sediments. In other words, we need to envision the isolation of the felsic reservoir from the surface until ~3.0 Ga.

This isolation is conceivable because the felsic crust was likely buried under thick layers of basalt (Arndt 1999). The ubiquitous presence of pillow lava in preserved ancient basalts suggests that many of these basaltic rocks were emplaced below sea level and onto flooded continents (Arndt 1999; Flament et al. 2008). On a hotter Earth, flooding can be explained by the reduced water storage capacity of a warmer mantle (Dong et al. 2021), and a shallower seafloor forcing oceans to overspill onto continents (Flament et al. 2008). Furthermore, as the strength of rocks is strongly temperature-dependent, the hotter continental lithosphere was much weaker and unable to sustain significant orogenic topography (Rey and Houseman 2006; Duclaux et al. 2007), making Archean landscapes flatter and with a lower freeboard (see Toolkit; Arndt 1999; Rey and Coltice 2008). Overall, no more than 3% to 4% of the Earth's surface (i.e., equivalent to the surface area of South America today) was above sea-level before the late Archean (Flament et al. 2008). The hypothesis of weak, flat, and flooded continents solves the first paradox by reconciling the early extraction of a significant volume of continental crust with its late appearance in the sedimentary record. Consequently, the measured shift in the geochemical signatures of detrital sedimentary rocks, detrital zircons, and carbonates does not trace crustal growth, but rather the exhumation of the felsic crust to the surface, its exposure to alteration and erosion, and ultimately its coupling to other geochemical reservoirs such as the atmosphere, oceans, biosphere, and mantle (Flament et al. 2013). To solve the second paradox, one needs a process that can explain the recycling of basaltic crust without leading to a significant increase in the $\delta^{18}O$ in crustal magmas.

ARCHEAN DUAL-MODE GEODYNAMICS: INTRA-CRUSTAL MASS REDISTRIBUTION, TRANSIENT SUBDUCTION, AND THE SECOND PARADOX

Petrology and Structure of Archean Cratons

The geologic record of Archean cratons is unique in many ways. Archean crust consists of an association of volcanic rocks with minor sedimentary rocks (greenstone belts), structurally in contact with underlying felsic rocks from the tonalite-trondhjemite-granodiorite (TTG) suite (Laurent et al. 2024 this issue). The structural architecture of the Archean crust is dominated by either TTG domes (up to 100 km in diameter) surrounded by narrow greenstone belts and/or prominent ductile strike-slip faults juxtaposing TTG gneiss and greenstone belts (e.g., Rey and Houseman 2006; Duclaux et al. 2007). The geochemistry of TTG suggests that many were formed by melting of hydrated mafic sources at pressures compatible with the stability of amphibole or garnet (Moyen and Martin 2012). Although earlier models of TTG formation invoked melting of subducting oceanic crust, melting of a hydrated mafic crust at 25 to 50 km depths is generally consistent with TTG geochemistry (e.g., Laurent et al. 2024 this issue). Greenstone belts are typically 5 to 25 km thick and accumulated over tens of millions of years. They are mostly composed of high-Mg rocks ranging from komatiites (MgO > 18%) to komatiitic basalts (12% < MgO < 18%) and abundant tholeiitic basalts (6% < MgO < 12%). Polybaric decompression melting of the mantle at depths ≤ 150 km can explain the volume of basalts in greenstone belts, the large dispersion of their FeO/MgO ratio, and the formation of the melt-depleted lithospheric mantle (Griffin et al. 2009), which forms the thick (~200 km) and buoyant roots of Archean cratons. Geochemical data from mantle xenoliths and seismic tomography suggest a stratification of the lithospheric mantle with, from top to bottom, strongly depleted, moderately depleted, and undepleted mantle (cf. Griffin et al. 2009; Rey et al. 2014 and reference therein).

Interpretations of the geology of cratons are debated. Greenstone belts are often presented as either ancient oceanic crust formed in back-arc basins above a subduction zone, and later accreted onto continents, or as stacks of lava flows formed via decompression melting in mantle plume heads. There are difficulties with both propositions. On one hand, it is curious that so many back-arc basins have survived erosion at convergent margins. On the other hand, the short duration of plume head volcanism, typically less than 5 million years, is hardly compatible with the duration of greenstone volcanism, which often extended over tens of millions of years. The interpretation of the crustal architecture is also strongly debated, and often reduced to an opposition between "horizontal tectonics," a misnomer for plate tectonics, and "vertical tectonics," a misnomer for diapirism. While some interpretations of TTG domes invoke polyphase contractional tectonics, or post-collisional extension, others invoke intra-crustal gravitational tectonics involving the sinking of greenstone belts into a hot and weak felsic crust, a process named sagduction (e.g., Mareschal and West 1980; Thébaud and Rey 2013). Similarly, prominent Archean strike-slip faults are interpreted either as accommodating the syn-collisional lateral escape of rigid blocks, or in terms of strain partitioning in a hot crustal environment under convergence (e.g., Duclaux et al. 2007).

Making Sense of the Geological Record

Numerical modeling can help explain first-order petrological and structural attributes of Archean cratons. The internal and boundary conditions of numerical experiments are informed by the thermal and mechanical properties of rocks and by first-order geological observations. Their temporal evolution is constrained by the physics of heat and mass transfer. Tracking the temperature and pressure evolution of particles makes it possible to consider processes such as partial melting and phase changes, and, thus, to integrate magmatic and geochemical observations with geodynamic and tectonic processes (e.g., Rey et al. 2014). Hence, modeling outputs can deliver physically

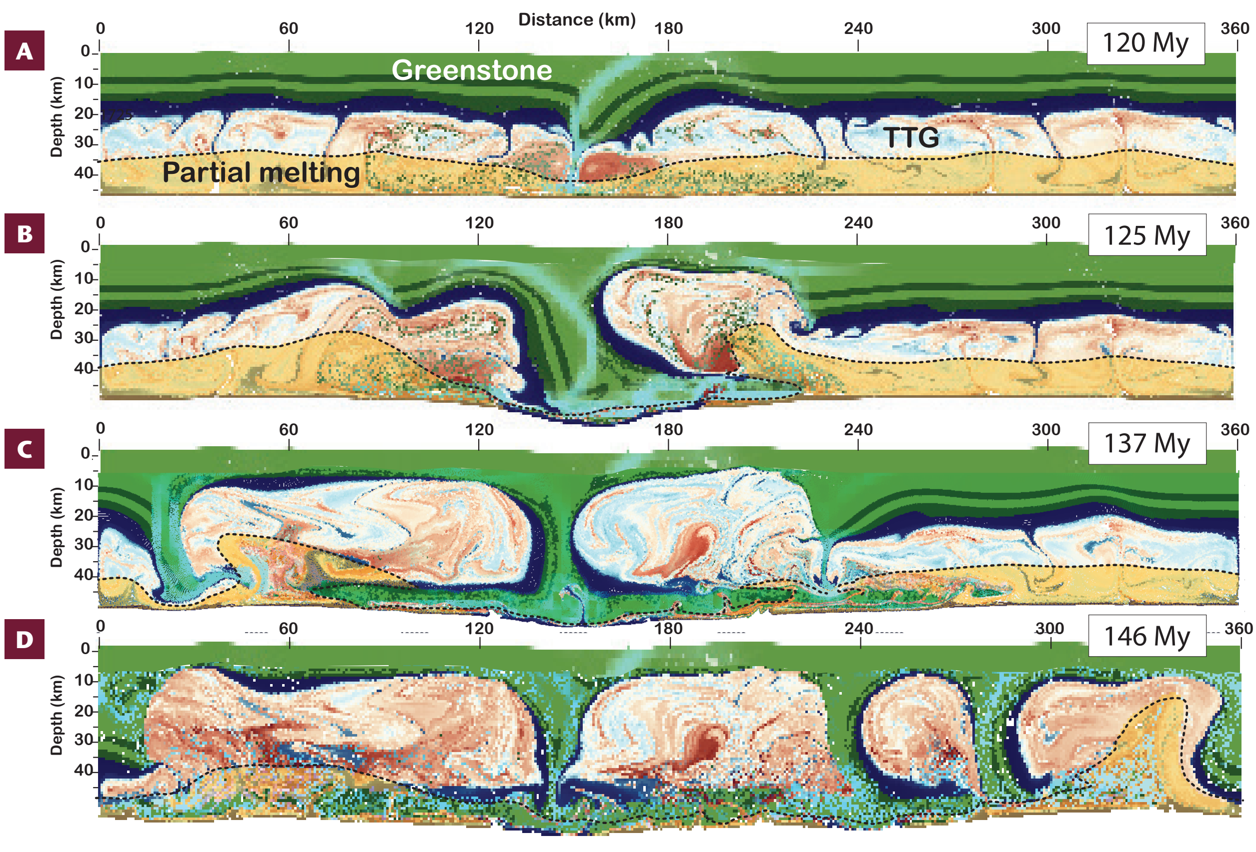

FIGURE 1 Accumulation of volcanic rocks (green to dark blue) leads to partial melting and TTG magmatism (A). The density inversion between greenstones (2840 kg·m^{-3}) and the TTGs (2720 kg·m^{-3}) drives sagduction of the cover and exhumation of TTG into domes. The slow build-up phase lasts ~120 million years (A) and precedes the instability phase during which a single dome can develop in a few million years (B), and multiple domes over ~100 million years. For modeling details see François et al. (2014).

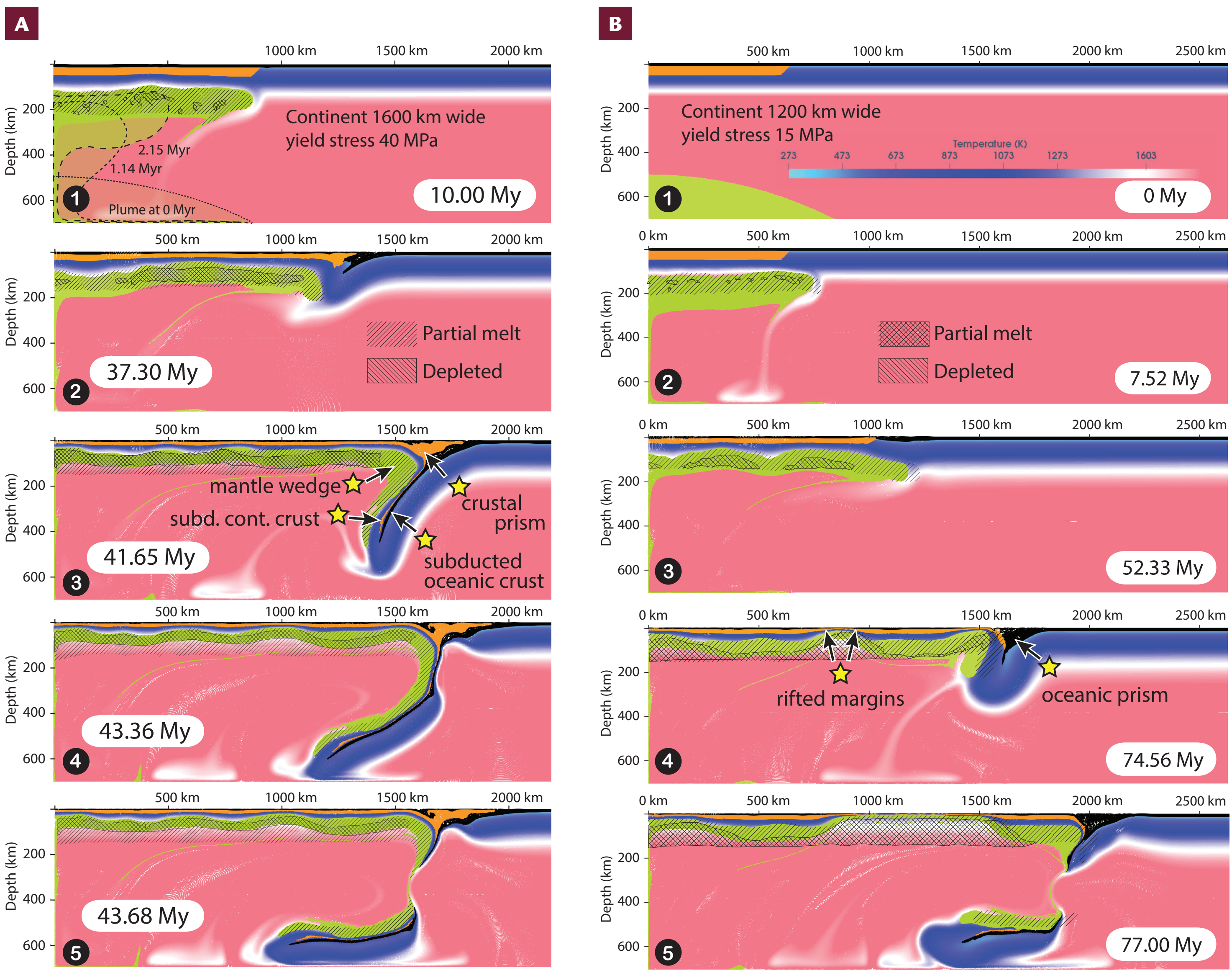

FIGURE 2 A numerical model with a mantle 200 °C hotter than present delivers a stagnant lid (blue) above a convective mantle (pink). Temperature is laterally averaged to remove convective stresses. A continental crust (orange) and a basaltic crust (black) are embedded into the lid. Plume material (pale green) is added at the bottom of the convective mantle (**A1**, **B1**). In (**A**), the protocontinent is wider and stronger than that in (**B**). Panels **A1** and **B1-2** show that the plume rises through the mantle in 2–3 million years and spreads laterally, forming a protocontinent in less than 10 million years. Gravitational spreading of these protocontinents forces subduction of the adjacent lid (**A2-5**, **B4-5**). Subduction drags portions of the protocontinent into the mantle (**A3-5**, **B4-5**). Thinning and rifting of the protocontinents lead to decompression melting of the underlying mantle. For modeling details see Rey et al. (2014).

plausible thermal, tectonic, and magmatic predictions that promote a better understanding of the composition and architecture of Archean cratons. A major advantage of this approach is that it mitigates the cognitive bias from our familiar imagery of the modern Earth, which prevents us from considering different, but plausible, scenarios.

Archean intra-crustal tectonics. Simple mathematical and analogue models have suggested that small lateral changes in the thickness of greenstone belts can drive their sagduction and the coeval exhumation of partially melted crust into gneiss domes (Mareschal and West 1980). In these pioneering experiments, sagduction led to diapirs rising vertically between sinking greenstone belts and delivered strain fields with a radial symmetry. In contrast, thermomechanical experiments using realistic thermal and mechanical properties (FIG. 1) show that, contrary to expectation, sagduction can deliver horizontal shortening and asymmetric strain fields like those generated by modern plate tectonics. In this setting, the stacking of asymmetric folds, layer-parallel folds, thrusts, and nappes, are not the signature of horizontal plate movement or plate tectonics (Thébaud and Rey 2013). Rather, they imply mass redistribution in a finite volume, where the downward motion of greenstone belts is accommodated by horizontal convergent transport in the upper crust to replace sinking material, and horizontal divergent flow in the deeper crust to make space for sinking material (FIG. 1). Therefore, sagduction tectonics is no more vertical than it is horizontal. Furthermore, thermal gradients in sagducted greenstones can be comparable to those in the subduction regime (François et al. 2014). Hence, at first glance, sagduction-related strain and temperature fields can mimic those of plate tectonic processes. Nevertheless, a key difference between modern and ancient strain fields is that strain in Archean cratons is often partitioned into

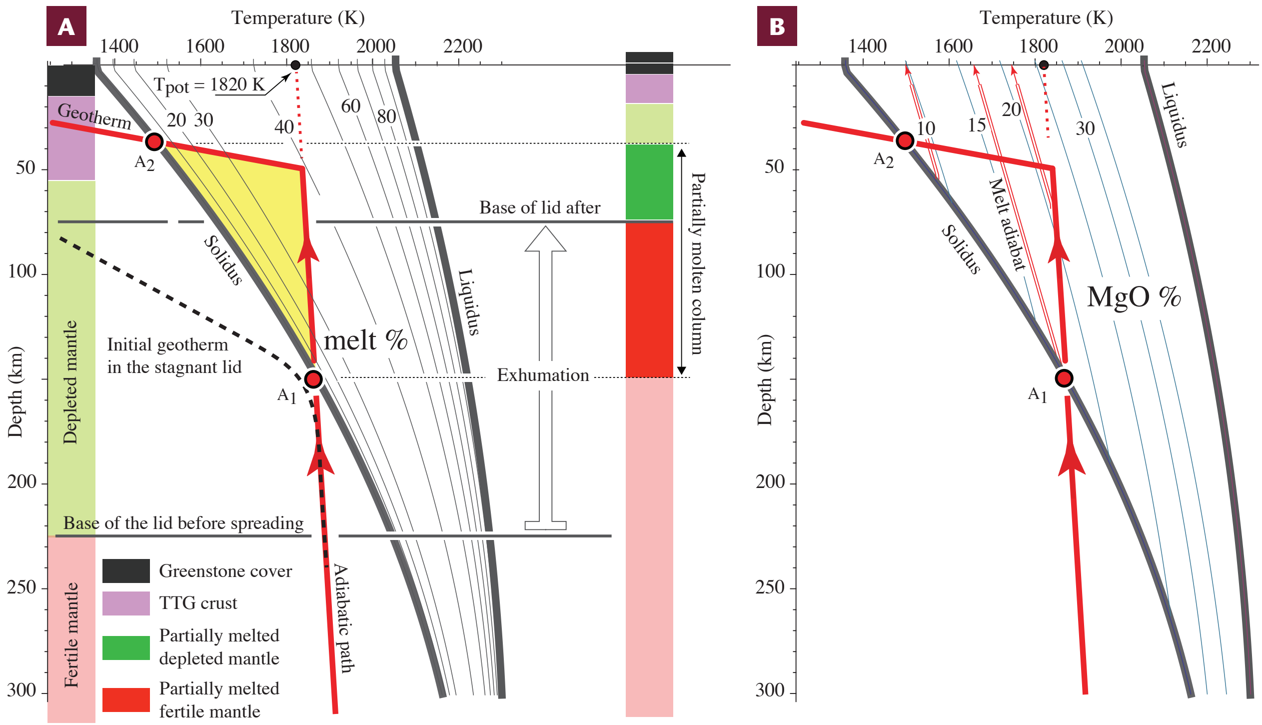

FIGURE 3 (**A**) Phase diagram showing how the spreading of a protocontinent triggers decompression melting of depleted and undepleted mantle. The columns on the right and left show the protocontinent structure before and after spreading. After spreading, the geotherm (in red) intersects the solidus at the base (A_1) and the top (A_2) of a partially molten layer. Whilst the base remains at 150 km depth, the top moves up (polybaric melting), leading to a basaltic layer whose thickness is given by the integration of the melt fraction in the yellow region (here ~20 km). (**B**) The MgO content of the melt extracted at various depths ranges from 9% to 21% (see Rey et al. 2014 for details).

adjacent domains of coeval extensional and contractional deformation (e.g., contractional greenstone belts adjacent to extensional domes; see Toolkit).

Because it involved the recycling of juvenile mafic rocks into TTG magmas, sagduction could explain the shift towards higher εHf(t) in TTG zircon from 3.8–3.6 Ga onwards. Sagduction could also explain the lack of a corresponding shift towards higher δ^{18}O. Compared with modern subduction zones, the volume of buried felsic sediments during sagduction was much more limited. In addition, whereas cold oceanic crust is buried in subduction zones, sagduction buried greenstones that evolved over a protracted period (>100 million years) under higher geothermal regimes (FIG. 1A), powering high-temperature (>230 °C) fluid–rock interactions (Thébaud and Rey 2013), which in turn may have lowered the δ^{18}O of rocks.

Although sagduction can explain Archean crust dominated by TTG domes and greenstone belts, Archean crust dominated by strike-slip tectonics suggests development under conditions of bulk shortening, which requires the existence of horizontal convergence. In what follows, we discuss the concept of mobile-lid geodynamics as a precursor to modern plate tectonics.

Archean Geodynamics

Secular cooling of the Earth and the transition from stagnant lid to mobile lid to plate tectonics. On a slowly cooling Earth, the geodynamics of the lithosphere–convective mantle system is expected to go through several regimes due to the strong dependence of rock viscosity with temperature. According to experiments and observations, hot rocks flow under low stresses, while cold rocks remain rigid, only breaking when a stress threshold (the yield stress) is reached. A strong temperature and stress dependence of viscosity with a stress threshold is required to achieve mantle convection coupled with plate tectonics (e.g., Coltice et al. 2017 for a review). Geochemical and petrological information on komatiites suggest that, during Archean times, the mantle was from 100 to 250 °C hotter than today. In numerical experiments that consider a mantle >150 °C hotter, the mantle vigorously convects underneath a cooler, rigid, and stable "stagnant lid" (Moresi and Solomatov 1998). The stagnant-lid geodynamic regime corresponds to the case where the lid and the convective mantle are mechanically decoupled. With secular cooling, the lid becomes progressively coupled to the convective mantle, and the stagnant-lid regime transitions to a "mobile-lid" regime, as convective stresses locally overcome the lid yield stress. Under a weak mechanical coupling, numerical experiments of mobile-lid systems produce only diffuse surface deformation, and at most, transient episodes of subduction (e.g., Coltice et al. 2017). The transition to a sustained plate tectonic regime, a particular mode of mobile-lid tectonics where deformation is strongly localized at plate boundaries, may have been achieved through further secular cooling and strengthening of the lid, as well as enhanced gravitational forces acting at continent–ocean boundaries in response to the deepening of the seafloor (Flament et al. 2008).

The role of protocontinents in Archean mobile-lid geodynamics. In the Archean, the lithospheric lid was compositionally distinct and hosted protocontinents that evolved into present-day cratons. Applying a simple decompression melting model to the hotter Archean mantle predicts that the oceanic crust would have been about three times thicker (see Flament et al. 2008 for a review) and the residual mantle more buoyant than today (Griffin et al. 2009). Geodynamic models developed to explore the stability of such a lithosphere show intermittent subduction with recurrent slab detachment (van Hunen and van den Berg 2008). However, in these models, subduction is imposed at the onset by embedding a slab ~250 km long. The initiation of subduction and the transition from a stagnant lid to a mobile lid can be achieved via a plume head 200 km in diameter piercing the entire lid (Gerya et al. 2015). However, such a plume head should begin to melt at a depth of 150 km and produce, in a few million years, a 50–60 km thick basaltic plateau. Neither of these predictions seem to fit the geological record as greenstone belts are at most 25 km thick and accumulated over many tens of millions of years. The presence of buoyant protocontinents embedded in the lid is a potential source of important horizontal gravitational stress that needs to be taken into consideration.

The model shown in FIGURE 2A1 considers that protocontinents developed through the accumulation of basaltic rocks extracted from a mantle plume, and from the accretion of the melt-depleted plume head at the base of the lid supporting the thick basaltic crust. Numerical experiments show that these hot, weak, and buoyant protocontinents could have spread under their own weight, forcing the initiation of subduction and the destabilization of the adjacent lid (Rey et al. 2014). This model predicts that the gravitational spreading and thinning of the protocontinent induces decompression melting of the underlying mantle (FIGS. 2 and 3). The parameters that control the dynamics of the lid are, on the one hand, its stress threshold and its viscous resistance to mechanical stresses, and, on the other hand, the buoyancy and size of the protocontinents. By varying these parameters within realistic limits, a range of mechanical behaviors is obtained from stable spreading of continents in a stable lid regime, through continental rifting accommodated by transient episodes of subduction (FIG. 2). These models suggest that protocontinents could have kick-started transient episodes of subduction, until plate tectonics became self-sustained.

Making sense of the geochemical and geophysical record. In the models presented in FIGURE 2, deep mantle melting occurs in the garnet stability field, and melting becomes progressively shallower (FIG. 3A), producing a range of mafic rocks (FIG. 3B) including small volumes of komatiites, komatiitic basalts, and abundant tholeiitic basalts. The model also predicts the formation of a second layer of moderately residual mantle under spreading protocontinents via partial melting of the undepleted mantle (red layer in FIG. 3A). The transient subduction zones that initiate at their edges could explain the co-occurrence of arc volcanism and high-Mg basalts in some greenstone belts. The timing, volume, and compositional range of these basalts are consistent with first-order observations in Archean greenstone belts. In addition, the model explains the presence of a strongly depleted and buoyant layer of mantle in the roots of Archean cratons (pale and darker green in FIG. 3A) above a layer of moderately depleted mantle, inferred from petrological and geochemical studies of mantle xenoliths (Griffin et al. 2009). The strengthening through cooling of these depleted roots would have helped preserve the cratons.

CONCLUSION: DUAL-MODE ARCHEAN GEODYNAMICS

We envision an early Earth with dual-mode geodynamics where intra-crustal tectonics, driven by sagduction, was independent from evolving geodynamics of the global lithosphere–mantle system. In this framework, the diachronous onset of sagduction and modern subduction is constrained by positive shifts of εHf(t) and $\delta^{18}O$ in zircon, respectively. The extraction of a primordial mafic crust, before ~4.3 Ga, left a melt-depleted, Mg-rich, buoyant upper mantle less prone to partial melting. Before 3.8 Ga, the primordial mafic crust evolved into TTG via recurrent partial melting in a largely closed system. The progressive mixing of the depleted mantle with the undepleted mantle, via impacts, convection, and mantle plumes, enabled the resumption of volcanism and the accumulation of greenstone basalts from ~3.8 Ga onwards. The shift towards higher εHf(t) of zircon beginning at 3.8–3.6 Ga would record the onset of sagduction. Spreading protocontinents could have triggered polybaric decompression melting, explaining the compositional range of greenstone basalts and the vertical structure of Archean cratons. Until ~3.0 Ga, the felsic crust was largely isolated from the surface by ongoing volcanism and a near global ocean. From 3.8 to 2.5 Ga, transient subduction at the edges of protocontinents did not affect the $\delta^{18}O$ of zircon in subduction-driven magmas. This could be explained by a combination of limited sediment supply at the margins of flat and flooded protocontinents, short-lived subductions restricting the amount of subducted material, and fluid–rock interaction at temperatures > 200 °C in hotter subduction zones on a hotter Earth. Protocontinents would have slowly emerged by 3.0 Ga due to secular cooling and the resulting deepening of the seafloor. The strengthening of the lithosphere would have enabled the formation of high mountains, enhancing erosion and sediment supply, and allowed for longer episodes of subduction, shifting the $\delta^{18}O$ in some arc magmas. Weathering and erosion would have exhumed the felsic crust from underneath the greenstones, and geochemically coupled the newly exposed TTG crust to the Earth's surface. Eventually, secular cooling strengthened the crust and inhibited the development of greenstone belts, and sagduction was eliminated from Earth's geodynamic repertoire.

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REFERENCES

Arndt N (1999) Why was flood volcanism on submerged continental platforms so common in the Precambrian? Precambrian Research 97: 155-164, doi: 10.1016/S0301-9268(99)00030-3

Bauer AM and 5 coauthors (2020) Hafnium isotopes in zircons document the gradual onset of mobile-lid tectonics. Geochemical Perspectives Letters 14: 1-6, doi: 10.7185/geochemlet.2015

Boyet M and 5 coauthors (2003) ^{142}Nd evidence for early Earth differentiation. Earth and Planetary Science Letters 214: 427-442, doi: 10.1016/S0012-821X(03)00423-0

Bowring SA, Williams IS, Compston W (1989) 3.96 Ga gneisses from the Slave province, Northwest Territories, Canada. Geology 17: 971-975, doi: 10.1130/0091-7613(1989)017%3C0971:GGFTSP%3E2.3.CO;2

Campbell IH (2003) Constraints on continental growth models from Nb/U ratios in the 3.5 Ga Barberton and other Archaean basalt-komatiite suites. American Journal of Science 303: 319-351, doi: 10.2475/ajs.303.4.319

Coltice N, Gérault M, Ulvrová M (2017) A mantle convection perspective on global tectonics. Earth-Science Reviews 165: 120-150, doi: 10.1016/j.earscirev.2016.11.006

Condie KC, Arndt N, Davaille A, Puetz SJ (2017) Zircon age peaks: production or preservation of continental crust? Geosphere 13: 227-234, doi: 10.1130/GES01361.1

Dong J, Fischer RA, Stixrude LP, Lithgow-Bertelloni CR (2021) Constraining the volume of Earth's early oceans with a temperature-dependent mantle water storage capacity model. AGU Advances 2: e2020AV000323, doi: 10.1029/2020AV000323

Duclaux G, Rey P, Guillot S, Ménot R-P (2007) Orogen parallel flow during continental convergence: numerical experiments and Archean field examples. Geology 35: 715-718, doi: 10.1130/G23540A.1

Dhuime B, Hawkesworth CJ, Cawood PA, Storey CD (2012) A change in the geodynamics of continental growth 3 billion years ago. Science 335: 1334-1336, doi: 10.1126/science.1216066

Fisher CM, Vervoort JD (2018) Using the magmatic record to constrain the growth of continental crust—the Eoarchean zircon Hf record of Greenland. Earth Planetary and Science Letters 488: 79-91, doi: 10.1016/j.epsl.2018.01.031

Flament N, Coltice N, Rey PF (2008) A case for late-Archaean continental emergence from thermal evolution models and hypsometry. Earth Planetary and Science Letters 275: 326-336, doi: 10.1016/j.epsl.2008.08.029

Flament N, Coltice N, Rey PF (2013) The evolution of the $^{87}Sr/^{86}Sr$ of marine carbonates does not constrain continental growth. Precambrian Research 229: 177-188, doi: 10.1016/j.precamres.2011.10.009

François C, Philippot P, Rey P, Rubatto D (2014) Burial and exhumation during Archean sagduction in the east Pilbara granite-greenstone terrane. Earth and Planetary Science Letters 396: 235-251, doi: 10.1016/j.epsl.2014.04.025

Gerya TV, Stern RJ, Baes M, Sobolev SV, Whattam SA (2015) Plate tectonics on the Earth triggered by plume-induced subduction initiation. Nature 527: 221-225, doi: 10.1038/nature15752

Griffin WL, O'Reilly SY, Afonso JC, Begg GC (2009) The composition and evolution of lithospheric mantle: a re-evaluation and its tectonic implications. Journal of Petrology 50: 1185-1204, doi: 10.1093/petrology/egn033

Guo M, Korenaga J (2023) The combined Hf and Nd isotope evolution of the depleted mantle requires Hadean continental formation. Science Advances 9: eade2711, doi: 10.1126/sciadv.ade2711

Hofmann AW, Jochum KP, Seufert M, White WM (1986) Nb and Pb in oceanic basalts: new constraints on mantle evolution. Earth and Planetary Science Letters 79: 33-45, doi: 10.1016/0012-821X(86)90038-5

Kemp AIS and 7 coauthors (2010) Hadean crustal evolution revisited: new constraints from Pb–Hf isotope systematics of the Jack Hills zircons. Earth and Planetary Science Letters 296: 45-56, doi: 10.1016/j.epsl.2010.04.043

Laurent O, Guitreau M, Bruand E, Moyen J-F (2023) At the dawn of continents: Archean tonalite-trondhjemite-granodiorite suites. Elements 20: 174-179

Mareschal J-C, West GF (1980) A model for Archean tectonism. Part 2. Numerical models of vertical tectonism in greenstone belts. Canadian Journal of Earth Sciences 17: 60-71, doi: 10.1139/e80-006

Moresi L, Solomatov V (1998) Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus. Geophysical Journal International 133: 669-682, doi: 10.1046/j.1365-246X.1998.00521.x

Moyen J-F, Martin H (2012) Forty years of TTG research. Lithos 148: 312-336, doi: 10.1016/j.lithos.2012.06.010

O'Neil J, Rizo H, Reimink J, Garçon M, Carlson RW (2024) Earth's earliest crust. Elements 20: 168-173

Rey PF, Houseman G (2006) Lithospheric scale gravitational flow: the impact of body forces on orogenic processes from Archaean to Phanaerozoic. In: Buiter SJH, Schreurs G (eds) Analogue and Numerical Modelling of Crustal-Scale Processes. Geological Society, London, Special Publication 253: 153-167, doi: doi:10.1144/GSL.SP.2006.253.01.08

Rey PF, Coltice N (2008) Neoarchean lithospheric strengthening and the coupling of Earth's geochemical reservoirs. Geology 36: 635-638, doi: 10.1130/G25031A.1;

Rey PF, Coltice N, Flament N (2014) Spreading continents kick-started plate tectonics. Nature 513: 405-408, doi: 10.1038/nature13728

Taylor SR, McLennan SM (1995) The geochemical evolution of the continental crust. Reviews of Geophysics 33: 241-265, doi: 10.1029/95RG00262

Thébaud N, Rey PF (2013) Archean gravity-driven tectonics on hot and flooded continents: controls on long-lived mineralised hydrothermal systems away from continental margins. Precambrian Research 229: 93-104, doi: 10.1016/j.precamres.2012.03.001

Valley JW and 8 coauthors (2015) Nano- and micro-geochronology in Hadean and Archean zircons by atom-probe tomography and SIMS: new tools for old minerals. American Mineralogist 100: 1355-1377, doi: 10.2138/am-2015-5134

van Hunen J, van den Berg AP (2008) Plate tectonics on the early Earth: limitations imposed by strength and buoyancy of subducted lithosphere. Lithos 103: 217-235, doi: 10.1016/j.lithos.2007.09.016

Wilde SA, Valley JW, Peck WH, Graham CM (2001) Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. Nature 409: 175-178, doi: 10.1038/35051550 ■