# Subaerial Weathering Drove Stabilization of Continents

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**Summary paragraph**

Earth's silica-rich continental crust is unique among the terrestrial planets and is critical for planetary habitability. Cratons represent the most imperishable continental fragments and form ~50% of the Earth's continental crust; yet, the mechanisms responsible for craton stabilization remain enigmatic[1]. Large tracts of strongly differentiated crust formed between 3 and 2.5 billion years ago, during the late Mesoarchean and Neoarchean time periods[2]. This crust contains abundant granitoid rocks with elevated concentrations of U, Th and K; the formation of these igneous rocks represents the final stage of stabilization of the continental crust[2,3]. Here we show that subaerial weathering, triggered by the emergence of continental landmasses above sea level, facilitated intracrustal melting and the generation of peraluminous granitoid magmas. This resulted in reorganization of the compositional architecture of continental crust in the Neoarchean. Subaerial weathering concentrated heat producing elements into terrigenous sediments that were incorporated into the deep crust where they drove crustal melting, and the chemical stratification required to stabilize cratonic lithosphere. The chain of causality between subaerial weathering and the final differentiation of the Earth's crust implies that craton stabilization was an inevitable consequence of continental emergence. Generation of sedimentary rocks enriched in heat producing elements at a time in Earth's history when the rate of radiogenic heat production was on average twice the present-day rate resolves a long-standing question of why many cratons were stabilized in the Neoarchean.

**Cratons and continent stability**

The most enduring blocks of continental crust, cratons, form refractory nuclei that have remained stable for billions of years and are some of the longest-lived and expansive geological features on Earth. Archean (>2.5 billion year old, Ga) cratons host the majority of the global gold and platinum inventories, and are important repositories of other critical mineral deposits, such as Li-bearing pegmatites. Cratons also contain key archives of ancient planetary environments including Earth's oldest preserved rocks[4], as well as records of ancient surface environments[5] and the climatic response to changes in the solid Earth system[6]. Cratons, defined here as blocks of >150 km thick stable lithosphere[1], are preserved in the Archean rock record as Eo- to Neoarchean (4.0-2.5 Ga) granitic intrusions and supracrustal rocks that comprise Archean 'granite-greenstone' belts. They have exceptional longevity – these packages of crust have remained stable and isolated from tectonic reworking for billions of years[7]. How these unique lithospheric domains were stabilized remains unresolved.

Critical to craton stability is the enrichment of heat producing elements (HPE) U, Th, and K in the upper crust relative to the lower crust. This serves to reduce temperatures in the deep crust and uppermost mantle, thereby strengthening the lithosphere to an extent that it becomes resistant to deformation[8,9]. Intracrustal transport of the HPEs occurs predominantly during melting, typified by post-orogenic magmatism, whereby partial melting concentrates HPE in the melt phase which ascends to depths of neutral buoyancy in the crustal column. Thus, from a crustal perspective, craton formation is marked by the timing of emplacement of post-orogenic granitoids, sometimes referred to as the Neoarchean granite bloom[7,10]. The precise timing of this key crustal differentiation event differs between cratons but consistently occurs between ~3.1 Ga and 2.5 Ga (Fig. 1b;[2,11]). The granite bloom was succeeded by tectonic quiescence over time scales of hundreds of millions to billions of years, indicating that this process represents the final stage of cratonization[11]. Importantly, these plutonic suites contain the first large-scale evidence for potassium-rich granites; Mesoarchean and older granitoids are dominated by distinctly different rock compositions[12,13]. Pre-granite bloom rocks contain mostly sodic ($K_2O/Na_2O < 0.7$) tonalite-trondhjemite-granodiorite-suite rocks (TTGs) and associated mafic rocks, forming Archean "grey gneiss" provinces[13]. These TTG-suite rocks are widely believed to be formed by partial melting of basaltic protocrust[13,14] and thus represent primary additions to the felsic continental crust, not the products of final stabilization.

In contrast to the older TTG-suite rocks, Neoarchean granites are potassic ($K_2O > 2.0\%$, $K_2O/Na_2O > 0.7$; Fig. 1a) and can be peraluminous (herein defined as granites with molar $Al/[Ca - 1.78P + Na + K] > 1.1$) and their chemical characteristics are incompatible with partial melting of mafic protocrust. Instead, the genesis of potassic and peraluminous granites require melting of older, intermediate composition continental crust (i.e., TTG crust) and sedimentary protoliths, respectively[2,15,16]. Thus, Neoarchean granites represent the final stage of crustal differentiation in the formation of Earth's continents. Neoarchean granites are enriched in the heat producing elements U, Th, and K (Fig. 2c) such that their formation by partial melting substantially depleted the lower crust of these elements, serving to strengthen and stabilize lithospheric blocks.

Prevailing explanations for the petrogenesis of the Neoarchean granites invoke heating of the lithosphere to induce partial melting of pre-existing crust. This requirement is supported by Neoarchean metamorphic terranes that preserve a record of high-temperatures in the middle and lower crust over timescales consistent with radiogenic heating in thickened crust [17,18]. Previously proposed mechanisms include mantle plumes impinging on the base of the lithosphere[19], lithospheric inversion[17], convective removal of lithospheric mantle[16], conductive equilibration of thickened crust[2,20], and advection of magmatic heat[21,22]. Though several aspects of these models are appealing, no one model can successfully account for all the geological constraints necessary to explain formation of cratonic nuclei in the Neoarchean.

A major obstacle, at least for some well-studied cratons, to models that invoke asthenospheric heat transfer as a mechanism to induce crustal melting is that formation of thick mantle lithosphere preceded (i.e. >2.8 Ga) granitoid emplacement in many cratons[1,23]. Impinging mantle plumes, or convective instabilities would destroy such lithospheric roots[17]; the existence of older mantle roots thus renders asthenospheric heating an unlikely heat source for widespread crustal melting in some areas. In contrast, conductive equilibration of thickened lithosphere with radiogenic crust offers a satisfactory explanation for the attainment of suprasolidus crustal temperatures[24] while maintaining thick lithosphere. Indeed, such thickening has been invoked to account for the thermal evolution of the thickest tract of lithosphere on modern Earth beneath Tibet[20]. In particular, it is important to note that granitoids with comparable compositions to the Neoarchean granite bloom event can be produced by this mechanism[25]. However, the invocation of crustal thickening to explain widespread anatexis on modern Earth requires anomalous crustal enrichment of the HPEs relative to the composition of modern continental crust.

Here, we use the Archean rock record to quantify the heat production rates of Archean TTG terranes. Global compilations[26,27] show that these terranes have substantially lower heat production (1-2 $\mu W/m^3$ at 2.8 Ga, Fig. 2d) than modern crustal compositions would have had in the Mesoarchean (~3-4 $\mu W/m^3$). This finding implies that extrapolation of crustal thickening to the Mesoarchean based on modern crustal compositions[20] overlooks significant secular changes in the composition of crust[12] and is therefore not warranted. We build on previous efforts that have demonstrated the importance of radiogenic heating for the production of Archean cratonic lithosphere[20,28] by combining rock-specific calculations of Neoarchean heat production with thermal models to evaluate the potential for crustal thickening to drive crustal differentiation.

**Archean heat production**

To quantify the heat production rates of typical rock types found in Archean terranes, we compiled geochemical bulk-rock analyses and calculated heat production rates at 2.8 Ga, an average age for the onset of granite bloom events, (Fig. 2) using present day U, Th, K concentrations in each rock type. We used major element rock compositions to define two groups of Archean crustal rock types, a 'high-Si' group (> 60 wt.% $SiO_2$) that is dominated by TTG-like compositions and a 'low-Si' grouping (< 60 wt.% $SiO_2$) that corresponds to basaltic rocks. Sedimentary rock data were partitioned into siliciclastic, shale and mafic sediment groups; our Archean shale composition represents an estimate of the archetypal sediment composition. Details of the calculations are provided in the Methods.

In the Mesoarchean, radiogenic heat production was around double the modern rate, but this depends upon the specific concentrations of each radioactive isotope (due to the dramatic differences in the half lives of $^{40}K$, $^{232}Th$, $^{235}U$ and $^{238}U$). Common to all Archean cratons is the predominance of TTG suites in poly-metamorphosed basement gneiss complexes[13]. Our calculations show that by the Neoarchean, felsic igneous crust--dominated by sodium-rich, intermediate to felsic TTG granitoids--had heat production rates <2 $\mu W/m^3$ (the 'high Si' group: median = 1.28, s.d. = 1.76, 75th percentile = 2.2, n = 2433; Fig. 2a). This defines the upper limit of the heat production capacity of pre-granite-bloom Archean crust as Archean terranes contain a mix of TTG and mafic gneisses[7,29] – the addition of mafic rocks to this package will reduce the internal heat production capabilities of pre-granite-bloom Archean crust (the 'low Si' group: median = 1.13 $\mu W/m^3$, s.d. = 2.6, 75$^{th}$ percentile = 1.16, n = 3945). In stark contrast, Archean sedimentary rocks from felsic sources (data from ref[30]), have significantly elevated heat production values that range from 1 to 6 $\mu W/m^3$ (median = 2.54 $\mu W/m^3$, s.d. = 1.5, 75$^{th}$ percentile = 3.66, n = 193; Fig. 2b) with an asymmetric distribution to higher values. Of these sediments, shales have the highest heat production rates (3-5 $\mu W/m^3$) as they are enriched in U compared to non-shales. Heat producing elements are also concentrated in both potassic (1-5 $\mu W/m^3$) and peraluminous (2-8 $\mu W/m^3$) Neoarchean granites (data from refs[2,15,31]). Peraluminous granites necessitate sedimentary source rocks, while many potassic granites in the Neoarchean have isotopic signatures that implicate involvement of older continental crust[2]. Further, as many melt/bulk rock partition coefficients for the HPE are <1[32,33], this implies that these peraluminous granites formed from HPE-enriched sedimentary protoliths. The potential for fractional crystallization to cause HPE-enrichment requires further evaluation; small grain-sizes of U- and Th-bearing accessory phases makes their physical separation from the melt implausible until they become included by a crystallizing major mineral.

**Thermal evolution of thickened crust**

Our assessment of Archean heat productivity based on actual rock compositions allows us to investigate the role of radiogenic heat production in cratonization. We use these rock-based heat production values to constrain one-dimensional thermal models of the cratonization process in which the thermal structure of thickened crust evolves through the combined effects of conductive relaxation of isotherms and radiogenic heating following thickening. Details of the calculations are provided in the Supplementary Information. Critically, we examine the ability of various distributions of crustal heat production to cause partial melting and differentiation of the continental crust in the Neoarchean.

Our calculations show that thickening of Archean TTG crust for a range of geologically plausible parameters fails to result in significant partial melting. Figure 3a shows an example result for TTG crust (1.4 $\mu W/m^3$, evenly distributed) in which 30 km-thick crust is instantaneously thickened by a factor of two. While this configuration results in >400°C heating of the middle and lower crust over 50 Myr, the peak temperatures attained do not significantly exceed the TTG solidus (700-800°C over the crustal pressure range, calculated using average TTG composition, see Methods for details) and produce only low-volume melts (<2%) that are insufficient to account for the voluminous record of magmatism preserved in the Neoarchean granitoid rock record[2,15,31]. This finding suggests that stacking of even pure TTG crust prior to the Neoarchean granite bloom could not produce the hallmark geological signatures of cratonization[20] and that an alternative heat source is required to produce the voluminous granitoids that mark the stabilization of Archean continental nuclei globally.

The common occurrence of metasedimentary rocks in Meso-Neoarchean granulite terranes[34,35] implies that heat production within thickened Neoarchean crust exceeded that of pure-TTG crust. Our calculations show that the incorporation of sediments into thickened crust has the potential to induce significant melting of both the sediment and adjacent TTG crust. Figure 3b shows the effect of sediment incorporation on the thermal evolution of Mesoarchean crust; for example, a 10 km-thick layer of sedimentary rock with an average heat production of 3 $\mu W/m^3$ incorporated at 30 km depth results in granulite-facies metamorphism and the generation of granitic melt fractions exceeding 40% after 50 Myr (green lines, Fig. 3b). Elevated rates of heat flow through the upper crust mean that the potential for sediments to undergo partial melting increases with depth of burial such that the distributed incorporation of low-HPE sediments throughout the deep crust (2 $\mu W/m^3$ between 30 and 60 km) is sufficient to yield melt volumes that locally exceed >25 vol.% (e.g. light blue line, Fig. 3b). The heat generated from buried sediments also stimulates partial melting of adjacent, non-metasedimentary crustal lithologies (Fig. S-9 and S-10), producing cogenetic suites of metaluminous and peraluminous granitoids. A further consequence of deep sediment burial and focused radiogenic heating is the downward conduction of heat from the lower crust into the underlying mantle lithosphere. Provided the negative heat flux is sufficient to cause melting of fusible rocks in the uppermost mantle lithosphere, this mechanism could account for the occurrence of minor volumes of mantle-derived mafic rocks exposed in some cratons during this interval.

**Geology of Archean cratons**

Our calculations imply that the onset of terrigenous sedimentation in the Mesoarchean, caused by the rise of continental freeboard[5], resulted in a step change in the ability of the crust to undergo internal chemical differentiation. Weathering of TTG crust concentrated HPE's into sedimentary lithologies whose burial provided the heat source required to internally differentiate the continental crust and produce cratons.

Large volumes of sedimentary rocks have not been present on Earth's surface throughout the rock record. In fact, mature sedimentary packages are only preserved on Earth since the Mesoarchean (Fig. 1b)[5,36]. The Archean felsic TTG crust has long been known to have been submerged beneath sea level – evidenced in large part by submarine basaltic packages that overlie most Mesoarchean basement gneiss assemblages (e.g. ref[7]). Although sedimentary packages are known to occur back to 3.8 Ga, prior to ~3.0 Ga, sedimentary packages tend to consist of thin, immature, 'cover-group' sequences that have restricted catchments and likely volcanogenic origins[37]. During the Mesoarchean, large sedimentary basins, including fine-grained sediments, were deposited on pre-existing continental crust. For instance, the Witwatersrand basin, deposited ~2.7 Ga, contains up to 6 km of interbedded clastic sequences[38]; Meso-Neoarchean sedimentary basins indicative of exposed continental crust exist in many other cratons, including the Slave (up to 5 km of turbidites and shales at 2.67-2.63 Ga[39]), Pilbara (>2.9 Ga, including the De Gray Superbasin[40]), and Kaapvaal craton (the 3.26 Ga Fig Tree Group containing up to 1 km of shale[41]). Importantly, large volumes of HPE-enriched sedimentary rocks are not found on Earth prior to ~3.0 Ga (Fig. 1b) despite the fact that there are large tracts of pre-3.0 Ga crust found in Archean 'grey gneiss' terranes – such terranes are dominated by orthogneisses of igneous origin.

Our proposed mechanism makes specific predictions for the relative timings of sedimentation, metamorphism and magmatism in the Neoarchean. Sedimentation is required to be antecedent to metamorphism and crustal melting, and we logically expect that both Neoarchean K-rich granitoids and exhumed granulite-facies metasedimentary rocks are consanguineous. Furthermore, granulite-facies metamorphism in the middle and lower crust should precede or be contemporaneous with the timing of plutonism in a specific cratonic region. There is ample geochronological and petrological evidence from specific regions that supports these predictions (Fig 1b; SI). For instance, the Neoarchean granite bloom event, comprising both peraluminous and potassic granitoids (see Supplemental Information) of the Slave craton occurred between 2.62-2.58 Ga[21], but is predated by clastic sediment deposition (2.69-2.66 Ga[7]), while metasedimentary xenoliths preserve a record of granulite-facies metamorphic conditions at ~2.62-2.59 Ga[42], correlating with granitoid emplacement. These granulite-facies Archean metasedimentary rocks are depleted in HPE and underwent partial melting at lower crustal conditions[34]. Lower crust in the Kaapvaal craton, exposed in the Vredefort impact event, contains metasedimentary rocks that underwent granulite facies metamorphism at ~3.0 Ga[43], succeeded by voluminous granitoid plutonism[2]. Moreover, combined zircon O and Hf isotope systematics from Kaapvaal peraluminous granites imply a negligible influence of mantle-derived melts during sediment melting under lower-crustal conditions, consistent with granite formation in response to radiogenic heating from HPE-enriched sediments[44]. Granite formation was coeval with metamorphism in the deep crust of the Kaapvaal[35] and resulted in the redistribution of HPE's to the near-surface and net strengthening of the crust[3].

At a global scale, our model predicts that granulite-facies metamorphism of sedimentary protoliths occurred during or prior to Neoarchean cratonization and that large tracts of melt-depleted metasedimentary crust currently exist at depth in Meso-Neoarchean orogenic crust. This requirement finds support from the observation that ~50 % of samples from exhumed Archean granulite terranes have peraluminous compositions[45], consistent with a metasedimentary origin and the incorporation of near-surface rocks into the Neoarchean lower crust. Furthermore, the conditions of metamorphism preserved by these rocks imply heating in middle and lower orogenic crust[46] (see PT data shown on Fig. 3). The few constraints that exist on the pressure-temperature-time evolution of such metamorphism indicate that granulite-facies metamorphism occurred within thickened crust[47] at a similar time to the Neoarchean granite bloom emplacement (Fig. 1b). Slow seismic wavespeeds and low $V_P/V_S$ ratios measured through some Archean cratonic regions (e.g. Kaapvaal craton[48]) provide additional support for the contention that portions of deep cratonic crust contain significant volumes of metamorphosed sedimentary rocks.

One aspect of the geological record that may seem to be at odds with our model is the observation that peraluminous granites are subordinate to potassic granitoids on some Neoarchean cratons[15]. The mechanism presented here predicts that melting was primarily driven by sedimentary heat production; however, it does not necessarily suggest that the Neoarchean magmatic signature should be dominated by pure sediment melts as the incorporation of sediments into the deep crust can stimulate the production of significant proportions of melt from pre-existing crust (see Methods). Generation of melts from pre-existing crust draws strong support from Nd and Hf isotope systematics of the global array of Neoarchean granites[2], notably including a large proportion of the western Slave craton[21]. Furthermore, post-Archean erosion may have removed upper crustal material containing such melts, as inferred from zircon oxygen isotope ratios[49].

**Tectonic style need not have changed**

Our proposed mechanism for cratonization, whereby weathering of emergent continents concentrated HPE into sediments that drove intracrustal differentiation, does not require a global reconfiguration in tectonic style in the Neoarchean to explain the observed increase in lithological diversity across this time period[12]. The mechanism is also consistent with the existence of cratonic mantle prior to Neoarchean granite formation, as likely required for continental emergence[1]. Several lines of evidence indicate that such lithospheric mantle was stabilized by imbrication[23], generating tracts of continental crust that could maintain freeboard and produce expansive sedimentary basins in the Neoarchean. These HPE-enriched sedimentary rocks were then incorporated into the deep crust by compressional tectonic activity that resulted in widespread melting and plutonism and was associated with the genesis of high-K, 'sanukitoid' magmatism produced by mantle wedge melting[50]. Incorporation of sedimentary rocks into the deep crust caused widespread crustal melting and plutonism; melt migration redistributed HPE from the middle- and lower- to the upper crust that, in turn, drove cooling and strengthening of cratonic crust. While Mesoarchean TTG-dominated crust was incapable of undergoing such differentiation, the appearance of rocks that concentrate heat producing elements within a plate tectonic regime is sufficient to cause intracrustal differentiation.

We emphasize that this mechanism has no specific requirement for the physical process by which sediments are incorporated into the deep crust. This may have occurred by relamination[51], tectonic underplating[52] or burial[53]; such processes are active on modern Earth and likely occurred in the Neoarchean[2]. Contrary to previous interpretations, we infer that tectonic regime need not have changed from the Meso- to Neoarchean in order to account for the geological evolution of Archean cratons. Instead, the mechanism proposed implies that craton stabilization was activated by continental emergence that, in turn, was driven by one of several viable processes unrelated to tectonic transitions[54–58]. The geological record can then be cast in terms of a pre-emergence (TTG dominated) and post-emergence (granite-dominated) planet. These findings demonstrate the importance of exogenic processes for the geodynamic evolution of planetary interiors and the generation and sustenance of habitable conditions. Ultimately, the onset of planetary-scale subaerial weathering led to not only dramatic atmospheric change, but also drove final distillation and stabilization of continents.

**Main figure legends**

**Figure 1**: Geological evolution of cratons at the end of the Archean. Panel **a** shows a stacked kernel density estimator for the composition of felsic (>62 wt% $SiO_2$) igneous rocks through time, discretized in 50 Myr age bins. Rocks with $K_2O/Na_2O$ are rare prior to 3.0 Ga and common post-3.0 Ga. Panel **b** shows a compilation of granitoid geochemistry from various cratons[2]. Sedimentation, high-temperature metamorphism, and emplacement of Neoarchean granites occurred at different times in each craton. Data sources are presented in the supplement. The lowermost green field in c shows the normalized cratonic mantle age distribution[1] with a prominent peak in the Neoarchean. Red and black lines show the number of preserved sedimentary (black) and fine-grained (red) rocks samples in a global database averaged in a moving-window calculation. Sedimentary rocks first appear en masse in the Mesoarchean-Neoarchean.

**Figure 2**: Heat production rates for Archean rocks. All values are back-calculated to 2.8 Ga using measured concentrations of U, Th and K; each curve is a kernel density estimator for the underlying datasets. Vertical lines correspond to heat production of modern upper crust[22] (UCC) calculated at 2.5 and 3 Ga. Source data and calculations are detailed in the Methods. Panel **a** shows heat production rates for Archean crustal rocks with silica content greater ('Archean High $SiO_2$') or less than ('Archean Low $SiO_2$') 62 wt.%; panel **b:** compositional endmembers of Archean sedimentary rocks; panel **c:** Late-Archean granites broken apart into compositional categories and **d** Archean granulite terranes and xenoliths.

**Figure 3**: Thermal evolution of thickened crust in the Neoarchean. Panel **a** shows the thermal evolution of thickened TTG crust. Left-hand plot shows the vertical distribution of heat production; middle plot shows geotherms in 5 Myr increments following instantaneous thickening of 30 km-thick crust by a factor of two ('saw-tooth' initial geotherm). Thick orange line represents the geotherm at 50 Myr, the solid black line shows the TTG solidus ('high-Si' bulk composition) and the gray markers represent a compilation of PT estimates derived from Archean metamorphic terranes[59]. Melt fractions corresponding to the 50 Myr geotherm are shown in the panel farthest to the right. Note that this crustal configuration does not result in significant melt production (<2 volume % melt at 50 Myr). Panel **b** is as for **a** except for distributions of heat production that correspond to sediment burial. Combinations of sediment layer thickness and heat production are chosen to demonstrate the importance of emplacement depth on thermal structure: lower thermal gradients with increasing depth mean that lower rates of heat production are required to cause melting. For the sedimentary layer, melt fractions were calculated using T-X relations derived for the 'Archean shale' composition. Colored lines show the thermal evolution of similar models that contain hot sedimentary layers placed at various depths and thicknesses (heat production shown on the left panel). Black lines correspond to solidus curves for TTG (as in **a**) and 'wet' shale composition. Abrupt changes in melt fractions with depth are caused by the exhaustion of muscovite during melting.

**Figure 4**: Subaerial weathering drives cratonization. Schematic illustration of the key geodynamic processes in craton stabilization. Panel **a**: prior to continental emergence,

thickening of TTG crust did not cause extensive melting, resulting in only limited crustal differentiation. Panel **b**: following continental emergence, weathering of TTG crust concentrated U, Th and K into terrigenous sedimentary rocks, which, when entrained into the deep crust, elevated the crustal geotherm over tens of millions of years and induced large degrees of melting. Subsequent melt-migration redistributed the HPEs from the lower- to the upper crust (potassic and peraluminous granites, yellow and pink, respectively), thermally and mechanically stabilizing the continental lithosphere. Vertical orange lines represent schematic distributions of U, and K.

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**Online Methods section**

**Whole-rock heat production calculations**

For all samples, the modern U, Th, and K concentrations were used to calculate total rock heat production back in time. Each radioactive isotope ($^{40}K$, $^{232}Th$, $^{235}U$, $^{238}U$), along with their respective decay constants, were used to calculate each rock's heat production. This type of element-concentration based calculation is essential to obtain a clear picture of the heat production in Archean rock compositions, as the distribution of U, Th, and, in particular K, vary between rock compositions and over time. For consistency, we calculated the heat producing capabilities of each rock sample at 2.8 Ga, then compiled these values (Fig. 2a-d).

To determine the sample-based heat production of Archean crustal rocks and sediments, we compiled modern (i.e. measured) concentrations of U, Th, K (in ppm) and major elements (in wt.% oxide; $SiO_2$, CaO, $TiO_2$, $Al_2O_3$, $FeO_T$, MgO, $Na_2O$, $K_2O$, where $FeO_T$ corresponds to total Fe) from published whole-rock compilations. The major element composition of each sample was used to define a statistical framework that enables us to compute average and median values of heat production for specific lithologies.

For Archean crustal rocks, we compiled whole-rock compositions from two sources (Fig. 2): compositions of Archean TTG 'grey gneisses' from ref. [29] were combined with a large compilation of Archean crustal rock compositions from ref. [60] (file 'aad5513-tang.m-sm.database.s2.xlsx' from their Supplementary Material). The conjoined dataset was filtered to remove partial analyses (i.e., those that did not report values for either U, Th or K, or the major elements), reducing the sample set to 6691 samples for which we calculate heat production values. A principal components analysis (PCA) of non-normalized major element data reveals that the majority of compositional variance (73%, Fig. S-1:A-B) is accounted for by classical indices of magmatic differentiation (i.e. $SiO_2$, MgO concentrations). We use non-normalized data for the PCA reflecting our interest in absolute—and not relative—values of compositional variance; all PCA calculations were performed using MATLAB's *princomp* function. Figure S-1:C shows that the aggregate dataset defines a broad positive correlation between heat production and whole-rock $SiO_2$ concentration, as is expected for plutonic rocks[61]. The correlation defines an obvious grouping threshold of 60-65 wt.% $SiO_2$, separating 'high-$SiO_2$' (n = 2046) from 'low-$SiO_2$' (4645) compositions that we use to compute population statistics for heat production (see main text); the distributions defined by these data are plotted in Fig. 2. Exemplar major element compositions for each group are shown as square markers on Figs. S-1:B and C and used in phase equilibria calculations detailed beneath.

Archean sedimentary rock heat production values were determined using the database of compositions from ref. [30]. This compilation was filtered for rock samples with a complete set of major element and U, Th and K concentrations, with depositional ages >2.5 Ga (n = 269). We performed a PCA on the non-normalized dataset, finding that two axes account for over 80 % of the compositional variance. The first axis ($X_1$; Fig. S-2A) separates $SiO_2$-rich compositions (i.e. arenites) from those rich in FeO and $Al_2O_3$ (i.e. mafic sediments and shales); the second axis ($X_2$) separates aluminous from ferruginous sediment compositions. These axes define three obvious end-members: siliciclastic, shale and mafic sediment compositions, between which the aggregate data form mixing arrays. Bulk-rock heat production increases toward the shale end-member (Fig. S-2B). For our calculations of heat production, we partitioned data according to orthogonal distance from each end-member sediment composition (i.e., which 'arm' of the $X_1$ vs $X_2$ structure the sample is aligned to). We then computed population statistics on the heat production for each of the three groupings, siliciclastic, shale, and mafic sediments, respectively; Fig. 2 shows the distribution of these data.

Compositions of late-Archean granite samples were aggregated from three sources: (1) peraluminous granites[15] (n = 84); (2) a Kaapvaal granite database[2] (n = 51), and (3) a Slave craton granite database[16,21] (n = 109). Filtering for analyses with U, Th, and $K_2O$ measurements reduced the database from 160 samples to 159 granite samples from the suite of samples from the Slave and Kaapvaal, and 84 peraluminous Archean samples. Reported (i.e. modern) U, Th, and $K_2O$ values were used to calculate heat production rates for each rock analysis at 2.8 Ga, and these values are plotted in Figure 2.

Heat production rates for Archean lower-crustal rocks from exhumed granulite-facies terrains and xenolith suites were calculated in a similar manner using the compilation of ref. [45]. Of these samples, 36 metasedimentary xenoliths had U, Th, and $K_2O$ values, while 441 metasediments from Archean granulite terrains had all three values allowing for an accurate calculation of the heat production at 2.8 Ga.

Heat production rates for modern continental crust back in time are based on the calculations of ref. [20] who used the composition of modern upper continental crust from ref. [62].

**Phase equilibrium calculations**

The effect of the latent heat of melting on thermal evolution is dependent on the relationship between temperature ($T$), pressure ($P$) and melt fraction ($X$). As the equilibrium melt fraction at any given $PT$ condition is controlled by the availability and compositions of fusible minerals, we calculated $T(P)$-$X$ relations for each of the compositions detailed above.

Phase relations were calculated using Theriak-Domino version 11.03.2020, downloaded from https://titan.minpet. unibas.ch/minpet/theriak/theruser.html[63] combined with the 2011 version of the ref. [64] thermodynamic database (ds62, update February 2012); post-processing was performed within MATLAB. Phase relations and rock properties were calculated between 4-16 kbar and 600-1200 °C.

Bulk compositions for phase equilibria calculations are presented in Table. S-1. For the crustal compositions ('high-Si' and 'low-Si', Fig. S-1), $XFe^{3+}$ ($XFe^{3+}$ = Fe3+/($Fe^{2+}$+$Fe^{3+}$)) was set to 0.2 based on values reported for the EAT composition of  Condie, (1981) in agreement with the general assumption that the Archean surface environment was less oxygenated than modern environments of hydrothermal alteration for which $XFe^{3+}$ typically exceeds ~0.3 (e.g. ref. [66]). For sedimentary compositions, $Fe_2O_3$ was set to 0.1 mol.% and we assume melting occurs under fluid-saturated conditions. Accordingly, bulk-rock $H_2O$ concentrations were adjusted so that minimal free $H_2O$ was present at the solidus at 10 kbar; this approach was adopted to ensure that melting initiated at the fluid-saturated solidus across the P-T range of interest.

All calculations were performed in the $Na_2O$-CaO-$K_2O$-FeO-MgO-$SiO_2$-$H_2O$-$TiO_2$-O (NCKFMASHTO) subsystem. The effect of Mn on phase relations was not considered due to low concentrations of MnO in most of the compositions and uncertainties on the energetics of Mn-mixing between relevant mineral phases. Furthermore, the principal consequence of Mn on phase relations is to stabilize garnet to lower-grade PT conditions[67], such effects are expected to have a minor impact on suprasolidus phase relations. For silica-rich compositions ('high-Si', siliciclastic, shale and Archean shale; Table S-1) the following activity-composition models were used: silicate melt[68,69]; plagioclase feldspar[70]; epidote[64]; chlorite, chloritoid, biotite, garnet and orthopyroxene[68]; white mica[68,71], magnetite and ilmenite[72]. For mafic compositions ('low-Si' and 'Mafic sediment'; Table S-1) we used the following activity-composition formulations: mafic melt, amphibole and clinopyroxene[73]; chlorite, garnet and orthopyroxene[68]; plagioclase[70]; olivine[64]; spinel[74]; magnetite and ilmenite[72]; epidote[64]; white mica[68,71]. The following pure phases were considered in all calculations: $H_2O$, albite, quartz, kyanite, sillimanite, rutile, sphene, clinozoisite and zoisite.

Figure S-3 shows the variation of melt fraction with temperature at 10 kbar for all bulk compositions considered. Assuming minimal saturation at the wet solidus, all compositions yield solidus temperatures between ~620 and 720 °C, but melt fractions diverge at >700 °C between fertile shales and the remaining rock types. Curves for both shale compositions are characterized by abrupt increases in dX/dT <800 °C, caused by muscovite dehydration melting reactions; our exemplar 'Archean shale' composition is predicted to yield a melt fraction of ~0.3 by ~800 °C. Such melt fertility is in contrast to the remaining sediment compositions as well as the ‘high-‘ and ‘low-Si’ bulks that are characterized by melt fractions <0.2 at <800 °C. The broadly basaltic ‘low-Si’ composition is predicted to yield large volumes of melt at elevated temperatures, between ~1000 and ~1150 °C, whereas the TTG-like ‘high-Si’ bulk produces most melt at slightly lower temperatures and over a greater temperature range, from ~850 to ~1100 °C. Note that the mafic sediment and ‘high-Si’ compositions yield broadly similar melting curves at >900 °C. Finally, the siliciclastic composition is the least fertile bulk composition considered, yielding a melt fraction of only 0.6 at 1200 °C, due to the predominance of quartz in all phase assemblages.

To examine the effect of variable $H_2O$ concentrations on melt fertility, we constructed T-X$H_2O$ curves for the ‘low-Si’ and ‘high-Si’ compositions. Solidus temperatures for both compositions range from <650 °C to >850 °C for saturated and dry melting scenarios, respectively. For ‘wet’ melting, we used $H_2O$ concentrations that minimally saturate the system at 10 kbar (‘wet’ melting lines, Fig. S-4); for ‘dry’ melting we adopted values of $H_2O$ that resulted in solidi temperatures around 800 °C. Reduced $H_2O$ concentrations would result in the onset of melting at higher temperatures.

We note that these T-X relations and melting curves are the result of continuous and discontinuous reactions amongst mineral phases and silicate melt. Interested readers are referred to refs. [68, 75] for a detailed discussion of the phase relations of pelitic and TTG compositions at supra-solidus conditions, respectively.

**Thermal modeling**

We model conduction and radiogenic heat production along a vertical column through the lithosphere [24,76–79]. The transient thermal field is given by:

$$\rho \frac{\partial [C_p(T)T + LX]}{\partial t} = \frac{\partial}{\partial z}\left[k(T)\frac{\partial T}{\partial z}\right] + A(z,t) \tag{1}$$

where $T$ is temperature, $t$ is time, $\rho$ is density, $C_p$ is heat capacity, $k$ is thermal conductivity, $L$ is the latent heat of melting, $X$ is melt fraction and $A$ is radiogenic heat productivity. The effects of advection (i.e. erosion) and heat of sub-solidus metamorphic reaction on the thermal field were not considered. We use temperature-dependent values of thermal conductivity and heat capacity as this exerts an important insulating effect on the thermal evolution of the deep crust [80]. For crustal depths, we employ the parameterization of ref. [81], whereas that of ref. [82] is used for the lithospheric mantle. The energetics of melt production were simulated assuming a value of 320 kJ/kg for $L$ and the lithology-specific T-X parameterizations derived above. Values of 2800 and 3300 kg/m$^3$ were assumed for crustal and mantle density, respectively. Crustal heat production was varied between model runs, but all calculations used 0.006 $\mu W/m^3$ for heat production within the mantle lithosphere[83]. The upper surface of the model domain was held at 0 °C for all times and the base of the model domain was held at constant temperature, defined by the initial geotherm and assumed lithospheric thickness. The initial (pre-thickening) geotherm was calculated using a steady-state formulation for a layered lithosphere with crustal and mantle heat production of 1.4 and 0.006 $\mu W/m^3$, respectively, and a Moho heat flux of 13 $mW/m^2$ [84]. We assumed a constant thermal conductivity of 1.8 W $m^{-1}$ $K^{-1}$ for the initial crustal geotherm to avoid unreasonably cold geotherms calculated with the temperature-dependent conductivity model of ref. [81]. Equation (S-1) was numerically integrated using an explicit finite difference scheme.

We assume that crustal thickening occurs by instantaneous emplacement of a single thrust sheet of variable thickness onto a continental section comprising crust and mantle lithosphere (see Fig. S-5). This configuration of thickening is similar to that observed along modern convergent margins – e.g. the Himalaya[85], and Andes, (e.g., ref. [86]; see ref. [77] for further details and examples) – in which low-grade rocks, including sediments, are delivered to middle and lower-crustal depths along contractional faults. In our calculations, we consider the thermal effect of underthrusting sediment layers of variable thickness and heat production (see Figs. S-9 and S-10). Following emplacement of the thrust sheet, the initial 'saw tooth'-shaped geotherm evolves in response to conduction and radiogenic heating. We do not consider the effects of heating during crustal thickening, nor the impact of a time-dependent mantle heat flux.

**Enrichment of Archean-Paleoproterozoic sediments in K and U**

Radiogenic heat production in sedimentary rocks is controlled by depositional age and radioelement concentration at the time of sediment deposition. For a sediment of an average modern composition, the former parameter imposes an approximate increase of heat production by a factor of ~2 (at 2.8 Ga) relative to the modern sediment.

Figure S-6 shows measured concentrations of U, Th and $^{40}K$ plotted against the age of sediment deposition for shales and 'non-shale' sediment compositions, as defined in the compilation of ref. [30]. The figure shows that $K_2O$ concentrations in all clastic sediments broadly decrease from peak values at ~2 Ga; U concentrations in shales appear to follow a similar trend, although modern black shale concentrations span a large range of concentrations up to 20 ppm. Thorium concentrations in clastic sediments increased from ~3 Ga to ~2 Ga after which the data are highly dispersed. A consequence of these secular variations in radioelement concentrations is that shales deposited between 2 and 2.5 Ga have elevated heat production relative to Phanerozoic shale compositions (Figure S-7 - the median heat production for Phanerozoic shales is 1.81 μW/m$^3$ [$Q_1$ = 1.47 μW/m$^3$, $Q_3$ = 2.8 μW/m$^3$] compared to 2.76 μW/m$^3$ for Archean shale [$Q_1$ = 1.85 μW/m$^3$, $Q_3$ = 3.75 μW/m$^3$]).

Uranium enrichment in Neoarchean-Paleoproterozoic shales is broadly contemporaneous with oxidation of the atmosphere[87,88]. The onset of oxidative weathering is expected to have solubilized U as mobile U(VI) complexes[89], increasing the continental U flux to marine basins[90]. Prior to ~2.5 Ga, U was likely immobilized as U(IV) in plagioclase in exposed granitoids and detrital mineral phases, such as uraninite, pyrite and siderite, in shallow-marine sediments[91,92].

**Geology of Archean cratons**

Here, we provide salient details of the geological histories of several Archean cratons where our proposed model provides a viable mechanism to explain Neoarchean intracrustal melting and granitoid formation.

*The Slave Craton*: contains some of the most expansive Archean sedimentary sequences known globally. The general geology has been summarized by many previous workers[7,42,93–95]. The Neoarchean record is defined by a thick package of tholeiitic submarine volcanic sequences (the Kam group) that erupted onto basement gneisses and thin packages of cover group sandstones, BIFs, and conglomerates. The mafic volcanic rocks generated an extrusive package of 1-6 km thickness that was succeeded by a phase of calc-alkaline volcanism. The predominant sedimentary rock sequences comprise two distinct packages of turbidites that were deposited between 2.66-2.61 Ga[93]. Limited geochemical analyses exist for these rocks, but in some areas the packages consist of >5 km of interbedded sandstones, siltstones and black slates[96]. Lower crustal xenolith suites, overwhelmingly associated with the Lac de Gras kimberlite field, Eastern Slave, are dominated by mafic granulites, but contain populations of metasedimentary granulites[34,42,97]. These granulite xenoliths record peak pressures of 0.8-1.2 GPa, indicating they were sourced from Moho depths during the 55 Ma kimberlite volcanism that brought them to the surface. Metasedimentary granulites have refractory compositions and indicate that they undergone substantial melt loss; they have heat productivities of ~0.29 μW/m$^3$ [34]. Metamorphic zircon growth occurred in Slave craton xenoliths in several intervals between 2.64 and 2.51 Ga[42], overlapping with the timing of plutonism represented by the 2.62-2.58 Ga Granite Bloom event in the Slave province. Thus, it is plausible that Neoarchean plutonism in the Slave province was driven by heat production in response to the addition of sedimentary materials into the lower crust during Neoarchean assembly of the craton.

*The Superior Craton*: is the largest, best-exposed and most intensely studied of the Archean cratons. The craton has been subdivided into E-W trending provinces that are commonly faultbounded, defining a lateral structure that has been used to argue for accretionary orogenic processes that sutured the Superior Province into its present configuration[98]. A period of (ultra)high-temperature metamorphism is recorded in the very large Pikwitonei Granulite Terrane at 2.68 Ga[99] which contains m-to-km scale bands or rafts of metasedimentary protoliths, though the bulk of the terrane is dominated by meta-igneous rocks. The adjacent North Caribou Terrane contains a history of sedimentation dating back to ~3.0 Ga[100], though classic wedge deposits are not typically found until 2.7 Ga. Across other parts of the craton, sedimentation occured between 2.72 and 2.68 Ga, while granitoid magmatism lags by 20-40 Ma. This typically coincides with metamorphism where preserved[101].

*The Amazonian Craton:* is formed by two Archean nuclei (the Guiana and Guaporé shields), separated by the Amazonian cratonic basin. Across the craton, TTG magmatism occurred between ~3 and 2.89 Ga and was succeeded by emplacement of sanukitoids and anatectic granites until ~2.84 Ga. In the Carajás Province, emplacement of these anatectic granites was associated with crustal thickening and granulite-facies metamorphism[102].

*The Pilbara Craton:* is the classic 'granite-greenstone' cratonic structure with domal granitic provinces intruding into older basaltic supracrustal packages. The preservation of the Pilbara craton is exceptional, with limited deformation and erosion as compared to other cratons – thereby providing an excellent window into surficial evolution, but a poorer record of metamorphism in the deep crust. The Paleo-Mesoarchean history of the Pilbara is dominated by plume-vertical processes[103,104] that constructed the East Pilbara Terrane. Up to 9-18 km of stratigraphy was developed by a combination of igneous and sedimentary rocks. Significant shale deposition does not occur until ~3.0 Ga[105], and no high-grade metamorphic sequences are recorded in the Pilbara, apart from contact metamorphism surrounding late granite plutons[104]

*The Karelian Craton:* spans ~400,000 km$^2$ of the Baltic Shield (northeastern Finland and adjacent Russia) and is dominated by late Archean TTG gneisses, greenstone belts, diorite-to-granite plutons and migmatitic metasediments. Emplacement of TTG granitoids occurred between ~2.95 and ~2.75 Ga prior to a phase of sanukitoid magmatism, culminating in biotite, and two-mica granite emplacement between 2.75 and ~2.63 Ma (see reviews by refs. 2 and [106]). Deposition of wackes and shales that form the protoliths of amphibolite-grade paragneisses occurred between 2.71 and 2.69 Ma, swiftly followed by regional amphibolite- and granulite-facies metamorphism from 2.7 to 2.63 Ga[107,108].

*The Kaapvaal Craton:* comprises an older-joined crust and mantle root compared to the Slave craton as evidenced by the geochronological investigations that constrain the last Archean magmatic event to ~3.1 Ga[109], apparently due to intracrustal melting during amalgamation of

Mesoarchean continental blocks. After this time, the continental block, or tectosphere, was stable and provided freeboard to erode and deposit large sedimentary basins such as the Witwatersrand Basin at 2.7 Ga. A large, near-complete, cross section of Archean Kaapvaal crust was exposed during the Vredefort impact event, 2.0 Ga[110]. Important to the proposed hypothesis is the fact that a significant amount of the Mesoarchean lower crust exposed by the Vredefort impact is metasedimentary. The exposed rocks include sapphirine-bearing granulites, interlayered felsic gneiss, felsic charnockites, and paragneisses[35,43]. Detailed mapping and geochronology of this exposed lower crust provided evidence for large volumes (~40%) of melting of pre-existing mafic and felsic lower crust in the craton ca. 3.1 Ga. Geochronological investigations constrain the age of crust-mantle coupling to between 3.09 and 3.07 Ga. However, ref. [35] argued that the degree of thermal reworking at 3.08 Ga was inconsistent with the presence of a deep, cool mantle root at that time on the grounds that a root would have served to impede heat transfer. Thus, the crust and mantle were likely aggregated independently. Our proposed mechanism, where intracrustal melting is instead driven by radiogenic heat delivery to the lower crust by underthrusting of sedimentary rocks, provides a viable alternative to independent formation and aggregation of Archean crust and mantle components. This model draws strong support from the Kaapvaal craton, where there is clear evidence for large volumes of residual metasedimentary rocks being present in the lower crust prior to emplacement of granite intrusions at 3.1 Ga[35,43]. Indeed, ref. [44] invoked a comparable model to explain zircon Hf- and O-isotope trends from 3.1 Ga peraluminous granitic rocks exposed in the Grunehogna Craton in East Antarctica – interpreted to be a part of the Kaapvaal craton lithosphere in the Archean.

*The Dharwar Craton:* is the largest of the five cratonic shields that collectively form Peninsular India and comprises a western shield (3.3-2.7 Ga) and a younger eastern shield (3.0-2.5 Ga). The western shield is dominated by TTG gneisses that are overlain by greenstone belts; in contrast, greenstone belts are rare in the eastern shield where late Archean (2.6-2.5 Ga) granitoids intruded older TTG (2.9-2.7 Ga) gneisses (see refs. [111,112] and references therein). Amalgamation of the eastern and western shields is constrained to 2.7 Ga. Supracrustal rocks of the Dharwar Supergroup were deposited between 2.9 and 2.72 Ga based on ages of metavolcanic horizons. Biotite and two-mica granites were emplaced between 2.54 and 2.51 Ga, contemporaneous with regional amphibolite- to granulite-facies metamorphism at 2.51-2.52 Ga[113,114].

*The North China Craton:* spans ~1.7 million $km^2$ across northeastern China, Inner Mongolia and North Korea, and is formed from a mosaic of microblocks that amalgamated prior to 2.5 Ga. Predominant lithologies vary considerably between microblocks, but all rocks with ages >2.5 Ga are affected by Neoarchean metamorphism and were intruded by 2.5-2.45 Ga granitic bodies. Across the craton, TTG plutonism occurred between 2.75 and 2.55 Ga; sanukitoids were emplaced in a short time period 2.55-2.52 Ga that overlapped with emplacement of biotite and two-mica granites (2.55-2.44 Ga). Metapelitic granulites preserve evidence for the incorporation of near-surface rocks into the lower crust along clockwise pressure-temperature paths, culminating at peak conditions at 2.49 (Qingyuan terrane, eastern North China Craton; e.g., ref. [115]) and ~2.52 Ga (Yinshan block, western North China Craton; e.g., ref. [116]).

**Predictions about volumes of peraluminous melts**

In the mechanism proposed here, intracrustal differentiation and craton stabilization is facilitated by radiogenic heat produced in metasedimentary rocks. An obvious prediction is that peraluminous – that is sedimentary – melts would be expected to dominate the Neoarchean rock record. Compilations of Neoarchean granitoids show, however, that such peraluminous granitoids are subordinate to I-type, metaluminous, melts across the majority of cratons (e.g., Bucholz and Spencer, 2019). There are several factors that reconcile this observation with sediment-driven crustal differentiation:

**i.** Neoarchean detrital zircons tend toward elevated $\delta^{18}O$ values relative to those preserved by the magmatic rock record (ref. [49], their Fig. 1), consistent with the erosional removal of differentiated and, in our model, high-heat production peraluminous melts from the rock record;

**ii.** There are cratonic locales where substantial volumes of peraluminous granites are locally preserved. Fig. S8 shows superimposed granite samples (colored symbols) onto a geological map[31] from the Slave craton. Red data points represent rocks mapped as sediment-derived melts while the blue points are mapped as metaluminous or potassic granites. Clearly, a substantial volume of peraluminous granite is exposed in this large (>1500 $km^2$) region, consistent with constraints from seismic wavespeeds and metamorphic rock compositions (see main text).

**iii.** Our calculations show that the incorporation of radioactive sedimentary material into the deep crust stimulates the production of melt from proximal—but non-metasedimentary—lithologies. Figure S-9 shows the volumes of melt produced from metasedimentary (yellow circles) and TTG sources (blue circles) for various burial depths and metasedimentary layer-thicknesses. Indeed, certain configurations predict a 1:1 ratio of metasedimentary and TTG melts (e.g., 10 km of 'Archean shale' incorporated at 40 km depth). Due to decreasing dT/dz with depth, the deeper a metasedimentary package is buried, the more potential there is for the stimulation of melt from proximal TTG crust. This mechanism would account for the observed dilution of pure S-type melts with I-type granitoids in Neoarchean cratons. Furthermore, metasediment-derived melts generated in the deep crust will also be susceptible to assimilation, the generation of mixed melts upon ascent and the potential erasure of their hallmark peraluminous nature.

## Further results from thermal modeling

Figures S-9 and S-10 show the quantities of sediment- (S-type) and crustal-derived (I-type) granitoid melt produced for various combinations of crustal hydration state, sediment composition, layer thickness and burial depth. Melt quantities are calculated by vertical integration of the melting column (e.g. Fig. 3) and are expressed as circular markers with radii scaled to the total km of melt produced after 50 Myr. Visual inspection of these results elucidates that thickening of pure TTG or basaltic crust (corresponding to layer thicknesses of 0 km) produces minimal melt (<2 km for 'wet' melting, Fig. S-9; <1 km for 'dry' melting, Fig. S-10) and such melting is restricted to depths > 30 km. Shales ('Shale' and 'Archean shale' compositions) are the most fertile of the sediment compositions due to mica-rich phase relations that result in characteristic steep dT/dX melting curves (Fig. S-5) at temperatures proximal to the wet solidus. In contrast, burial of mafic sediment produces the smallest melt fractions as a combined result of shallow dT/dX melting curves at <800 °C (Fig. S-4) and low radiogenic heat production (Figs. 2 and S-2).

Figures S-9 and S-10 also show that burial of shales and siliciclastic sediments into the deep crust stimulates the production of granitoid melts from adjacent TTG and basaltic rocks. For example, crustal thickening of 30 km of dry TTG ('high-Si' composition) results in 0.4 km of I-type granitoid melt in comparison to 1.5 km when 15 km of Archean shale (3.4 $\mu W/m^3$) is buried to 30 km. Stimulation of such melt occurs predominantly in response to downward conduction of radiogenic heat produced in the sediment layer—an effect that could plausibly lead to melt production in the lithospheric mantle beneath orogens. Note that this effect does not occur with the burial of mafic sediments within TTG crust (Fig. S-9, left column, third row), due to the limited difference in heat production of mafic sediments compared to TTGs (1.65 $\mu W/m^3$ vs 1.7 $\mu W/m^3$, respectively) and due to the structure of the melting curve for TTG at depth.

## Data and Code Availability statements

Computer codes replicating the results here can be obtained from the authors upon request.

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**Author contribution statements**

J.R.R. and A.J.S. conceived the project. J.R.R. curated geological and compositional datasets; A.J.S. conducted thermal and petrological calculations. Both authors contributed equally to the formal analysis and writing, reviewing and editing of the final draft of the paper.

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**Ethics declarations**

Competing interests

The authors declare no competing financial interests.

**Extended data figures and tables**

Figure S-1: Principal components analysis (PCA) of Archean crustal rocks. Panel **a** shows principal component loads for Archean igneous and metamorphic rocks (referred to as ‘crustal rocks’ throughout). The first principal component ($X_1$) dominates the compositional variance (73%) and separates silica-rich from mafic compositions; the second principal component accounts for a subordinate proportion of the total variance (20%) and separates compositions enriched in CaO, $Al_2O_3$ and $FeO_T$ from those enriched in $SiO_2$ and MgO. Panel **b** shows principal component scores for the compiled dataset of 6691 whole-rock analyses shaded according to radiogenic heat production at 2.8 Ga. Square markers correspond to exemplar ‘high-Si’ and ‘low-Si’ compositions defined in panel **C** that, in turn, shows heat production calculated at 2.8 Ga plotted against bulk-rock $SiO_2$ wt.%.

Figure S-2: Principal components analysis of Archean sedimentary compositions. Panel **a** shows principal component loads calculated from non-normalized bulk-rock compositions. The first principal component ($X_1$) accounts for 63% of the compositional variance and separates silica-rich sediments from other compositions; the second principal component ($X_2$) accounts for 24% of the variance and separates aluminous sediments from more ferruginous compositions. Panel **b** shows PCA scores for 269 Archean sedimentary compositions shaded according to heat production calculated at 2.8 Ga. End-member compositions are shown by the red (siliciclastic), green (shale) and yellow (mafic sediment) square markers; the blue marker corresponds to our exemplar ‘Archean shale’ sediment composition used in the phase equilibria calculations.

Figure S-3: Variation of melt fraction (X) with temperature. Curves show calculated melt fractions plotted against temperature at 10 kbar for all compositions considered. See text for discussion.

Figure S-4: Effect of bulk-rock water concentration on solidus temperature for ‘low-Si’ and ‘high-Si’ compositions. Panel **a** shows T-X pseudosection calculated for ‘low-Si’ bulk composition; abscissa is bulk-rock molar % $H_2O$. Red line represents the solidus and blue line represents the phase stability limit of free $H_2O$. Thin lines correspond to mineral stability fields; labels omitted for clarity. Vertical dashed lines correspond to molar $H_2O$ concentrations used for ‘dry’ and ‘wet’ melting scenarios. Panel **b** is as for A except for the ‘high-Si’ composition; both panels calculated at 10 kbar.

Figure S-5: Physical sketch of thermal model. Prior to thickening the crust and mantle are in thermal equilibrium (**a**). Subsequent crustal thickening occurs along a crustal-scale thrust fault (**b**) and is approximated by instantaneous emplacement of a thrust sheet onto a continental section (**c**). The thermal structure then evolves via conductive relaxation of isotherms and radiogenic heating (**d**). Note that burial of HPE-enriched sediments (yellow layer) by this mechanism results in a different equilibrium thermal structure to that prior to deformation.

Figure S-6: Secular variations in concentrations of radioelements, K, Th and U, in shales and non-shale sediments. Lowermost panels show volumetric heat production calculated at the age of sediment deposition. Data are from the global compilation of ref. [30].

Figure S-7: Probability density plots of radiogenic heat production for Archean and Phanerozoic sediments. Shales are plotted as solid lines, whereas ‘non-shale’ clastic sediments are plotted as dashed lines. Data are from the global compilation of ref. [30].

Figure S-8: Map of a granitoid province in the NE Slave craton, after ref. [16]. Data points are granitoid samples colored according to their composition and source characteristics.

Figure S-9: Quantities of melt produced from hydrous melting of crustal and sedimentary lithologies. The columns show the background composition of the crustal column, while the rows indicate the composition of sedimentary layers imposed on the thermal models. For each individual plot, the x-axis indicates the depth of burial for a given thickness of sedimentary layer, shown on the y-axis. Melt volumes for each composition are shown by blue and yellow circles.

Figure S-10: Quantities of melt produced from dry melting of crustal and sedimentary lithologies. Structure of the plots are identical to Figure S-9 except they are conducted for dry melting scenarios.

Table S-1: Bulk-rock compositions used for phase equilibria calculations (weight % oxides).Definition of suprascript symbols: $^{\Psi}$ $Fe_2O_3$ set to 0.1 mol.%; $^{*}$ $XFe^{3+}$ set to 0.2, where $XFe^{3+} = (2(Fe_2O_3)/(FeO + 2(Fe_2O_3))$; $^{\phi}$ $H_2O$ concentration required for minimal saturation at the solidus, P = 10 kbar**.** ‘Archean shale’ composition derived from shales with depositional ages >2.5 Ga, whereas ‘Shale’ composition include rocks of all ages.