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Deep continental roots and cratons

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Summary:

The formation and preservation of cratons - the oldest parts of the continents comprising over 60% of the continental landmass - remains an enduring problem. Key to craton development is how and when the thick strong mantle roots that underlie these regions formed and evolved. Peridotite melting residues forming cratonic lithospheric roots mostly originated via relatively low-pressure melting and were subsequently transported to greater depth by thickening produced by lateral accretion and compression. The longest-lived cratons assembled during Mesoarchean and Paleoproterozoic times, creating the 150 to 250 km thick, stable mantle roots that are critical to preserving Earth's early continents and central to defining the cratons although we extend the definition of cratons to include extensive regions of long-stable Mesoproterozoic crust also underpinned by thick lithospheric roots. The production of widespread thick and strong lithosphere via the process of orogenic thickening, possibly in several cycles, was fundamental to the eventual emergence of extensive continental landmasses - the cratons.

[158 Words]

1. The lithospheric mantle – a brief introduction

The outer highly viscous “skin” of the Earth - the lithosphere - separates the surface from its interior. Lithosphere definitions have many nuances¹. Here, we define the lithosphere simply as Earth's strong outer thermal boundary layer through which heat is primarily transferred by conduction (**Box 1**). The base of the lithosphere can be defined as the point at which a linear extrapolation of this conductive geotherm intersects the mantle isentrope. The cooler temperatures and higher viscosities of lithosphere compared to underlying asthenospheric mantle contribute to it being one of the longest-lived large-scale features of

the solid Earth. The mantle portion of the lithosphere, the “mantle root”, is generally thicker and older beneath continents than oceans².

Given the controversies surrounding the origin of the crust that we walk on and sample readily as geologists, it should not surprise the reader that the origin of the deeper parts of the solid Earth, such as the continental lithospheric mantle, is as controversial and more difficult to constrain. Here, we review some physical and chemical properties of deep lithospheric roots beneath continents and examine their integral role in forming the oldest parts of the continents - the cratons. We explore how these properties arose in the context of mantle melting environments. The melting ages of peridotites forming the cratonic roots, and their temporal relationship with the overlying crust are examined before using geodynamic models to constrain the origin of the large-scale geological characteristics of cratons, i.e., how the cratons were made.

2. Making cratons and their lithosphere

2.1. Defining a craton

Despite lithospheric mantle comprising up to 80% of the thickness of continental plates, the origin and evolution of these deep roots remains contentious. Cratons produce over 90% of the world’s gold and platinum and almost 100% of its diamonds. The properties of cratonic lithospheric roots are becoming increasingly recognized as key factors in the topographical expression of continents³, lithospheric volatile storage⁴ and the location of many metal deposits⁵.

83 Understanding the role of lithospheric mantle in the stabilization and subsequent protection
84 of continents requires clarification of the term craton. The original use - kratogen - from the
85 Greek “kratos” meaning strong⁶, merely implied a continental terrane displaying long-term
86 stability of 100s of Myr, with no age definition. Following Kennedy⁷, Clifford⁸ recognized an
87 association of ancient continental masses (> 1.5 Ga) with certain mineral deposits,
88 especially diamonds, gold and platinum, though more recent widespread use of “craton” has
89 become synonymous with Archean regions. Yet many “cratons” have long-lived tectonic
90 histories that belie the image of post-Archean “stability”.

91

92 Studies of the Kaapvaal craton (**Box. 1**), generally use a craton definition specifying a region
93 where basement crustal rocks are > 2.5 Ga (e.g., ⁹) yet major mid-craton disruption and
94 magmatic addition affected it in Paleo- and Mesoproterozoic times. Studies of the Siberian
95 and Amazonian cratons (**Fig. 1**) have followed the broader definition¹⁰ of “a segment of
96 continental crust that has attained and maintained long-term stability, with tectonic
97 reworking being confined to its margins”. For the 4.1 million km² Siberian craton, much of
98 the crust surrounding two Archean nuclei was either intensely metamorphosed or formed
99 in Paleoproterozoic times¹¹. Similarly, the Amazonian craton has only a small area of clearly
100 established Archean crust¹² within large regions of Paleoproterozoic crust¹³. It, along with
101 numerous other cratons (e.g., Rae, Hearne & Gawler; **Fig. 1**) have been extensively intruded
102 by felsic plutons in the Paleo- and Mesoproterozoic making the restriction of the definition
103 of a craton to geological inactivity since the Archean problematic.

104

105 A thick lithospheric mantle has long been identified as a distinguishing feature of cratons,
106 intimately linked to their stability (e.g., ^{2,14 15 16 17}). Hence, it seems logical to involve the

107 mantle root in the definition of a craton. These deep strong mantle keels play a role in
108 protecting the overlying crust from subsequent reworking and recycling, diverting mantle
109 plume heat and mass fluxes away from the roots¹⁸ without significant modification, unless
110 weakened by metasomatism^{19,20}.

111

112 Seismology can be used to assess the correlation between deep cratonic mantle roots and
113 ancient crust (**Fig. 1**). Areas underlain by cool, thick (>150 km) lithospheric mantle roots
114 with anomalously fast seismic wave speeds extend well beyond the ~55 Archean nuclei of
115 continents²¹⁻²⁵ (**Box 1; Fig. 1**), underpinning the much larger projected outlines of
116 composite terranes amalgamated and stabilised in the Paleo- to Mesoproterozoic, for
117 instance well into the Mesoproterozoic terranes of North America²⁴. It is clear from
118 seismology and geochronology (**Fig. 1; Box 1&3**) that the thickest lithosphere is not
119 exclusive to the oldest crust, and most of Earth's crust older than ~ 1 Ga has >150km thick
120 lithospheric roots²⁶ that have been present since that time¹⁶. This naturally leads to a more
121 practical definition of craton:

122

123 **Cratons** - are coherent blocks of Precambrian lithosphere, typically stable for time periods in
124 excess of a billion years due to protection by deep (> 150km) lithospheric keels. The majority of
125 Earth's Archean crust lies within such terranes and can be referred to as Archean cratons or
126 nuclei. But the moniker "craton" is not exclusive to the Archean. Larger composite cratons may
127 consist of Archean nuclei surrounded by Palaeo- Mesoproterozoic crust, all underpinned by
128 thick cool lithospheric mantle. Supercratons comprise multiple composite cratons (Fig. 1).

129

Cluster analysis of seismic tomographic models²³ or lithospheric thermal properties²⁷, can produce regionalized maps of upper mantle structure by identifying similar types of lithospheric mantle. These maps (e.g., Supplementary **Fig. 1**) successfully match thick lithosphere with the above definition of cratons, significantly increasing the contiguous cratonic area of some regions such as the southern African supercraton²⁶ and in global terms (**Fig.1**). Cratons, as defined here, comprise ~ 63% of the exposed continental surface, or ~ 18% of Earth's surface [Supplementary **Fig. 1**].

Exceptions are areas of Archean crust that have recently lost an old thick lithospheric keel and are now underlain by relatively thin (<100 km) lithosphere, e.g., the eastern North China and Wyoming cratons (**Fig. 1**). Crust in these regions survived for billions of years likely because they were protected by a thick root until lithosphere thinning in Phanerozoic time^{28 29 30 31}. They are classified as “modified cratons”. Regions of young crust, underlain by thick lithosphere such as Tibet, or the Central Asian Orogenic Belt (**Fig. 1**) have not yet achieved cratonic stability, but may do so²⁶.

2.2 Cratonic mantle lithosphere – composition and properties

Earth's upper mantle, based on sampling, theory and experiment, is composed largely of peridotite³², typically a rock dominated by olivine, along with orthopyroxene, clinopyroxene and an Al-rich mineral that, depending on pressure and bulk composition, is garnet or spinel beneath continents [**Box 1a**], or rarely plagioclase (<1 GPa). Extraction of melt from peridotite is the key process defining the formation of Earth's lithospheric mantle. Mantle peridotites residual after significant melt extraction are referred to as “depleted” whereas compositions similar to estimates of un-melted mantle are known as “fertile” (**Box 2**).

154 Fertile mantle is characterized by the peridotite sub-type lherzolite, which transforms to
 155 harzburgite then dunite as clinopyroxene and then orthopyroxene are consumed by
 156 progressive melting (**Box 2**). The shift from fertile to depleted compositions has a major
 157 influence on peridotite density since Fe is preferentially extracted over the atomically
 158 lighter Mg from olivine, the main mantle mineral. In addition, Al - a key component of the
 159 dense minerals garnet and spinel - is removed in the melt¹⁴. A maximum of ~ 2.5 %
 160 reduction in bulk density occurs from fertile lherzolite to ultra-depleted dunite³³ (**Box 2**).
 161 The resulting systematic variations in Fe and Mg and, in turn, the degree of depletion can be
 162 traced via the Mg# (molar $100 \cdot \text{Mg} / [\text{Mg} + \text{Fe}]$) of olivine, which depends on the upwelling
 163 path of the mantle during melt extraction as well as the efficiency of melt removal³³⁻³⁹.
 164 Different melting paths engender differences of ~ 0.62% in density at 35% melt extraction³³
 165 (**Box 2**).
 166
 167 Compositionally-derived density decrease combined with the higher viscosity of cooler
 168 melt-depleted lithospheric mantle imparts physical stability to continental lithosphere^{2,14}.
 169 Also, water - as dissolved hydrogen - is a primary control on mantle viscosity⁴⁰. Melt-
 170 depleted, dry, cool lithospheric mantle is much more viscous, and hence stronger, than
 171 fertile, warmer upper mantle that has higher water contents, with viscosity and buoyancy
 172 being key to the robustness of continental roots^{41,42} against attack by tectonic processes
 173 and mantle plumes.
 174
 175 Seismic^{22,24-26} and gravity data⁴³ indicate that cratonic lithosphere is heterogeneous and
 176 that the lithospheric mantle sampled by deeply-derived melts such as kimberlites (Mg-CO₂
 177 H₂O-rich low-volume magmas) may be compositionally atypical, showing more extensive

effects from re-fertilisation due to melt/fluid infiltration (metasomatism). For instance, tens of km-scale seismic heterogeneities beneath the Lac de Gras kimberlite field, Canada⁴⁴, likely represent mantle modification by intruded kimberlites. Also, most lithospheric sections sampled as xenoliths/xenocrysts indicate that the lowermost lithosphere in regions of kimberlite activity is more Ti and Fe-rich than mantle overlying it^{45,46} due to metasomatism.

Geophysical observations of cratonic mantle and the lack of kimberlite eruptions in the thickest lithosphere²⁶ imply that the mantle peridotite xenolith record may provide a biased view – as recognised by petrologists⁴⁷. Modified compositions and mineralogies in many xenoliths are recorded by anomalously high clinopyroxene and/or garnet contents relative to their very elevated Mg#^{48,49} and trace element systematics of clinopyroxene and garnet that commonly reflect their formation from, or equilibration with migrating fluids or melts long after lithosphere formation⁵⁰⁻⁵³.

Despite the modified nature of many cratonic mantle xenoliths, no other samples exist. Petrological approaches to correct for the effects of metasomatism can infer the nature of the unmodified peridotite protolith⁵⁴⁻⁵⁶. Compositions derived from these approaches agree with the compositions of the most depleted, least modally metasomatized cratonic peridotite suites, e.g., Murowa, Zimbabwe craton⁵⁷; East and West Greenland^{34,48,58,59} i.e., that much of the cratonic mantle lithosphere, away from effects of kimberlite transit, comprises harzburgites or dunites with high Mg but low Ca and Al.

2.3 Melting conditions

202 How do such depleted peridotite compositions originate? The suggested tectonic melting
 203 regimes vary from deeply-derived mantle plumes (high average pressure of melting: > 5
 204 GPa) building lithosphere by vertical accretion of residues^{55,56,60,61} to decompression
 205 melting in rift-related environments (low average pressure of melting: < 5 GPa), beneath
 206 thin lithosphere, for instance at mid-ocean ridges^{47,62-72}, or hydrous flux-melting in the
 207 mantle wedges of subduction zones⁷³⁻⁷⁶.
 208
 209 The different genetic hypotheses have been supported by different geochemical arguments.
 210 Authors favoring high-pressure (> 5 GPa) mantle plume melting highlight the low FeO of
 211 many Archean peridotites as indicating melt extraction at 5 to 7 GPa^{55,56,77,78}. However, the
 212 spread in FeO contents is large, elevated by some types of melt metasomatism and lowered
 213 by orthopyroxene growth from Si-rich melts/fluids^{67,79}. While deep-plume initiated
 214 polybaric melting beginning at high pressures and ending at low pressures beneath thin
 215 lithosphere is most effective in extracting the highest total melt-fraction from peridotitic
 216 mantle [**Box 2**], melting initiated at lower pressures, under rifts or hot spreading centres,
 217 where melting might begin at 5 GPa due to elevated mantle potential temperatures in the
 218 Archean and Paleoproterozoic^{66,67,70,80} is also effective at removing melt fractions of
 219 between 30 to 40 %. Such melt fractions can generate the typical olivine Mg#s of cratonic
 220 peridotites^{37,66} and reproduce olivine Mg# versus depth trends⁸¹. More definitive is the
 221 systematic variation in heavy rare earth element (HREE)^{62,67,71,72,76,82} (**Fig. 2**) and transition
 222 metal element concentrations⁶³ indicating that garnet was not a residual phase during most
 223 of the melting regime, that began at 5 to 3 GPa.
 224

225 Mineral chemistry is also a powerful indicator of the depth of melting. High-Cr garnets in
226 cratonic peridotites and peridotite-type diamond inclusions are best explained by high-
227 pressure metamorphism of melt residues that originally formed at lower pressures, in the
228 spinel stability field⁷¹. The range of high-temperature garnet compositions produced in
229 experimental melt-residues or via exsolution from high-pressure melting residues is narrow
230 and low in Cr (3 - 5 wt % Cr₂O₃), contrasting strongly with the extended range to higher Cr
231 contents in cratonic peridotite garnets. The higher garnet Cr contents are generated by
232 metamorphic formation of garnet - at high pressures - from melt residues produced as
233 spinel peridotites during melting at lower pressures^{62 83}. Some cratonic garnet may have
234 exsolved from high-T orthopyroxene, but calculated pre-exsolution host orthopyroxene
235 compositions have high Al and low Ca, likely formed in shallow melting residues within the
236 spinel stability field⁸⁴.

237

238 Some have suggested, from bulk rock chemistry, that the parental low-pressure melt
239 residues may have been serpentinites^{85,86} or that parts of the keel formed from hydrous
240 diapirs from hydrated slabs in the transition zone⁸⁷. However, oxygen isotope compositions
241 in cratonic peridotites are remarkably homogenous and “mantle-like”, over a wide range of
242 Cr and Si contents⁸⁸ precluding a serpentinite or hydrated slab origin.

243

244 High-pressure, plume-derived melt residues, produced through Earth history, may also be
245 present in cratonic lithosphere because they too are buoyant and viscous and may be
246 entrained during lithosphere accretion. In one conspicuous case, cratonic lithosphere of the
247 north Slave craton has been thinned by plume erosion and re-thickened ⁸⁹ by the trapping of
248 higher-pressure melting residues in the “thin-spot”, as traced by geochemistry (**Fig. 2**).

However, while such residues contribute to the compositional heterogeneity of cratonic mantle, their contribution, based on the geochemical evidence, appears to be subordinate, with the combined trace-element and mineralogical evidence dominantly reflecting a lower pressure (< 5 GPa) polybaric decompression melting regime. That many cratonic peridotites now record equilibration pressures > 5 GPa requires tectonic transport of these buoyant shallow-formed melting residues to greater depth during thickening of the lithospheric roots during craton-formation processes^{62,65,66,71,72,74-76,83,90}.

2.4. Age structure of cratonic mantle, crust-mantle relations and temporal trends

Having established the presence of widespread thick mantle lithosphere beneath continental crust of Archean through Mesoproterozoic age (**Fig. 1; Box 1**) as a key ingredient of cratons, it is important to examine specific age relationships between cratonic roots and their overlying crust, which can constrain the mode craton formation. A challenge is that peridotite ages recorded by trace-element isotopes are not, in all cases, the age of formation of the lithospheric keels – a fact well illustrated by the very large spectra of melting ages for peridotites found beneath the very young continent of Zealandia⁹¹ or the ancient Os model ages observed in abyssal peridotites^{92,93}.

Early radiogenic isotope studies of cratonic mantle and inclusions in diamonds showed very different isotopic signatures to the asthenosphere, establishing isolation of lithospheric mantle for Gyr periods⁹⁴⁻⁹⁶. The Rb-Sr, Sm-Nd and U-Pb systems used in these studies have very low parent-daughter element concentrations in residual peridotite relative to mantle melts. Hence the ages dominantly reflect “enrichment ages” due to metasomatism rather

273 than the age of the melt extraction relating to lithosphere formation. Subsequent application
274 of the ^{187}Re - ^{188}Os decay system to mantle peridotites produced ages that more likely
275 reflected the approximate age of melt depletion^{29,61,77,97} [**Box 3**].
276
277 Widespread application of the Re-Os isotope system to dating lithospheric peridotites has
278 led to a clearer picture of the age of cratonic mantle roots, though the approach is a blunt
279 tool in terms of providing a precise estimate of when melt depletion occurred^{76,81,98-103}.
280 Critically, because Os isotope heterogeneity in Earth's mantle increases with time,
281 interpretation of Re-Os model ages from peridotites or sulfides younger than ~ 1.5 Ga have
282 large uncertainty because of the isotopic diversity of Phanerozoic mantle [**Box 3**]. Portions
283 of modern oceanic lithospheric mantle frequently produce Re-Os model ages of over 1 Ga
284 and sometimes up to 2 Ga, at the mineral to meter scale^{92,93}. This heterogeneous "age"
285 spectrum for modern mantle lithosphere demands caution when interpreting single
286 occurrences of ancient ages of 1 to 2 Ga in a lithospheric peridotite suite dominated by
287 much younger ages, derived from Os isotope compositions that are statistically identical to
288 modern upper mantle. Instead, some studies infer the ubiquitous presence of Archean
289 mantle, refertilised in Proterozoic or younger times, from the occurrence of a single Archean
290 Re-Os model age in a peridotite suite¹⁰⁴. But it is increasingly clear that mantle peridotite
291 age spectra are strongly influenced by the persistence of ancient ages in the upper
292 mantle^{105,106}. Similarly, while more ancient Os model ages can be sometimes recovered from
293 sulfide or alloy grains in peridotites than from most whole-rock analyses in a given
294 suite^{93,103,107-109}, it is important to better understand the spectrum of Os isotope
295 heterogeneity in Archean mantle before interpreting isolated older ages as the age of

296 lithosphere formation, even though these estimates may well prove accurate with extended
297 studies.

298

299 With the above context, we can now explore questions such as: What are the oldest
300 components of cratonic lithospheric mantle and what is the likely age of formation /
301 stabilization of the lithospheric keels? Six first-order observations are:

302

303 **1)** Some cratons contain crust of Hadean or Eoarchean age, but no such ages are
304 preserved for cratonic mantle, precluding a primary origin for cratonic roots as lithosphere
305 formed, for instance, during any early “stagnant lid” phase of Earth’s evolution e.g.¹¹⁰. Some
306 melting residues from these times may have been incorporated into cratonic roots, their
307 ages over-printed.

308 **2)** Though mantle melt residues were clearly being produced by Earth before the
309 Mesoarchean, the vast majority of the peridotite depletion ages are Meso- to Neoarchean,
310 documenting the rapid growth of thick cratonic keels over a shorter period, and at a
311 different rate than the continental growth curve (**Box 3**).

312 **3)** A peak in depletion ages between 2.0 and 1.8 Ga occurs in mantle lithosphere
313 beneath Archean and Proterozoic cratons (**Box 3**), indicating mantle root formation in the
314 Paleoproterozoic¹¹¹⁻¹¹⁴, either coincident with the assembly of some larger composite
315 cratons e.g., Siberia, or prior to / during the formation of the Nuna supercontinent, which
316 created some of the larger composite cratons observed today.

317 **4)** Despite the association of predominantly Archean mantle underlying Archean
318 nuclei and Proterozoic mantle underlying Proterozoic cratons (Box 3), more age decoupling
319 exists between crust and mantle than originally proposed, e.g.,⁷⁷, with lithospheric mantle

320 recording Archean melt extraction underlying Paleoproterozoic crust^{48,115}, Paleoproterozoic
321 mantle residues underlying Archean crust¹¹⁴, Neoarchean mantle residues underlying
322 Meso- to Eoarchean crust¹¹⁶ and Mesoproterozoic residues underlying Neoarchean crust⁸⁹.

323 **5)** Archean melt residues form some of the lithospheric building blocks to Proterozoic
324 cratons. Vestiges of these foundations are also seen in lithosphere beneath modified
325 cratons, but these lithospheric sections have been mostly replaced by more recent mantle
326 **(Box 3)**

327 **6)** With the exception of Tibet or the Central Asian Orogenic Belt - cratons in the
328 making²⁶ - lithospheric mantle underlying regions of thick (>150km) continental
329 lithosphere formed > 1 Ga ago **(Box 1; Fig. 1)** .

330

331 These points are amplified in the context of specific cratons. The Kalahari composite craton,
332 lying within the southern African supercraton, comprises the Kaapvaal and Zimbabwe
333 Archean nuclei and associated Proterozoic terranes **(Box 1b)**. The nuclei show broad age
334 correspondence between Archean crust and highly depleted Meso- to Neoarchean cratonic
335 mantle^{62,74,76,77,82,98,100,109,117}. Where younger lithosphere ages exist, such as the center of the
336 Kaapvaal nucleus (Premier kimberlite) the Paleoproterozoic melt depletion ages **(Box**
337 **1b)**^{74,77} reflect major rifting, disruption and healing of the craton at ~ 2 Ga, coincident with
338 the formation of the massive Bushveld complex. The overall crust-mantle age relations
339 indicate a diachronous Meso- to Neoarchean stabilisation age for the ~ 200 km thick
340 lithosphere across the Archean nucleus **(Box 1b)**¹¹⁸, reflected by the broad mode of
341 peridotite Re-Os model ages. In the outer portions of the Kalahari craton, crust in the
342 Rehoboth and Namaqua-Natal terranes is ~ 1.3 to 2.2 Ga¹¹⁹ **(Box 1b)**. Peridotite xenoliths
343 indicate the roots beneath these outer cratonic regions are up to 200 km thick and Re-

344 depletion ages range from 1.2 to 2.4 Ga (single locality modes between 1.25 and 1.75 Ga ¹²⁰)
345 potentially reflecting lithospheric mantle produced during pulses of juvenile magmatism¹²¹.
346 Subsequent accretion to the craton during the Namaqua-Natal accretionary orogen created
347 the thick lithosphere observed beneath most of this Proterozoic cratonic region imaged by
348 seismology^{22,25,122} (**Fig. 1**).

349

350 The high degree of crust-mantle age “coupling” beneath cratons implied by early work^{16,77}
351 has not withstood further scrutiny (**Box 3b**). For instance, in the Siberian craton, Re-Os and
352 Lu-Hf dating shows that while some remnant Archean mantle is present, the dominant melt
353 depletion event for cratonic peridotites beneath the Daldyn-Markha portion of the Anabar
354 Archean nucleus occurred ~ 2.0 to 1.8 Ga^{111,112,123}, whereas highly-depleted lithospheric
355 mantle beneath the Olenek terrance of the same nucleus is predominantly Archean^{123,124}. A
356 complex age structure is also evident in the mantle beneath the central Rae craton, northern
357 Canada, where older, shallow Archean mantle is underlain by mantle that either formed, or
358 was massively over-printed at circa 1.8 Ga¹²⁵. Similarly, the Archean Sask craton (Canada)
359 is underlain by mantle depleted at ~ 1.8 to 2 Ga, associated with the Trans-Hudson
360 orogeny¹¹⁴, and the central Superior craton experienced significant lithosphere replacement
361 during the Mesoproterozoic ¹²⁶. In contrast, the Paleoproterozoic Halls Creek orogen (West
362 Australia composite craton) is underlain by deep lithospheric mantle of Archean age¹¹⁵, and
363 a similar relationship exists in East⁴⁸ and West¹¹⁶ Greenland, part of the Laurentia
364 supercraton (**Fig. 1**).

365

366 Some Paleoproterozoic cratonic peridotites have highly depleted major-element and mineral
367 compositions resembling those formed in the Archean, e.g., in Arctic Canada¹¹³ (**Box. 1a**).

368 This and other examples^{58,67,91,127} indicate that very depleted melt residues can be produced
369 well beyond the Archean/Proterozoic boundary, in contrast to some proposals^{55,128}. Such
370 highly depleted Paleoproterozoic lithospheric mantle became incorporated into craton
371 roots during major post-Archean craton-forming accretionary orogens¹²⁹.

372

373 The variation in peridotite compositions - and hence melting conditions - with geological
374 time is of interest in understanding the origins of cratonic peridotites as well as mantle
375 thermal evolution^{70,80,130}. A recent approach⁷⁰, augmented here using the most reliable
376 estimates of melting ages for peridotite suites screened via criteria such as extended PGE
377 patterns^{81,103} (**Box 3b**) indicates that the apparent secular decrease in peridotite olivine
378 Mg# with decreasing model age (excluding Phanerozoic arc peridotites) fits well with the
379 expected trend of secular decrease in mantle potential temperature, at Urey ratios of
380 between 0.2 and 0.3. This fit, though imperfect, suggests that no anomalously hot mantle
381 plume is required to explain the melting regime of cratonic peridotite residues, consistent
382 with an origin via relatively shallow decompression melting⁷⁰.

383

384 **2.5 Formation of cratons and cratonic mantle – the importance of** 385 **lateral accretion**

386 In the context of craton formation, the debate over the relative roles of residual peridotites
387 formed by mantle plume melting versus those formed by the thickening of residues of
388 shallow polybaric decompression melting can be addressed through geodynamic modelling.
389 Mantle lithosphere above modern mantle plumes experiences net lithospheric thinning, e.g.,
390 beneath Hawaii, where the maximum lithosphere thickness is equal to or thinner than

391 normal oceanic lithosphere¹³¹. Similarly, in the central North Atlantic craton, the ~ 200 km
392 thick mantle root present throughout the Proterozoic Eon^{132,133} was thinned locally to 60
393 km by ~ 60 Ma plume activity¹³⁴. The Ontong Java plateau is an exception since mantle
394 xenoliths reveal a lithosphere exceeding 120 km¹³⁵. but the uppermost 80 km formed from
395 normal oceanic lithosphere¹³⁶. Beneath Africa, seismology indicates that plumes are the
396 sites of lithosphere erosion¹³⁷ and are implicated in plate destruction, not growth¹³⁸.

397

398 Geodynamic modelling of the dispersion of plume melting residues (**Fig. 3; Supplementary**
399 **Video 1**), shows that excess mantle potential temperatures in the upwelling plume, in an
400 ambient mantle that was ~200 K hotter than the present-day MORB source are sufficient to
401 counteract viscosity increases due to melt depletion, allowing rapid dispersal by the plume
402 mass flux, either back into the upper mantle or forming relatively thin, widespread layers of
403 residual mantle, adding slightly to lithospheric depth but not attaining the 200 km thickness
404 of most cratonic lithosphere. Compressional thickening is required to achieve cratonic root
405 thicknesses. Buoyant plume residues seem to be effective at “re-cratonising” lithosphere,
406 after plume-related thinning, coalescing to re-form > 150 km thick lithosphere⁸⁹. In
407 contrast, the residues of high degrees of decompression melting at low average pressures,
408 in rift environments remain at their sites of generation. These residues form at lower
409 mantle potential temperatures, cooling more rapidly to attain the high viscosities needed
410 for stabilization of cratonic roots^{3,139}. The lithospheric columns produced by such melting
411 must then be thickened to depths seen in cratonic roots.

412

413 The dominant lithosphere during Archean times was unlikely to have been as dynamic as in
414 modern-day ocean basins, with perhaps only episodic mobility and nascent subduction-like

415 features^{139,140}. Hence, while extensive polybaric decompression melting at low average
416 pressure is required by cratonic peridotite geochemistry, long-lived mid-ocean ridge
417 spreading centres may not have been as extensive in the Archean as in modern Earth. Other
418 models of early Earth lithosphere dynamics invoke extensive melt extraction at sites of
419 lithosphere rifting/divergence leading to formation of segments of strong buoyant
420 lithospheric “blocks” via strain localization and cooling, sustaining further extension and
421 melting¹³⁹. The resulting mix of depleted lithospheric blocks can amalgamate and thicken
422 via lateral compression/accretion and further cooling into ~ 200 km thick, depleted, cool,
423 cratonic lithospheric roots¹⁴¹.

424

425 Lateral accretion, by either compression during the formation of accretionary orogens or a
426 shallow subduction-like processes involving slab stacking, has long been invoked to play a
427 role in the thickening and stabilization of old and young continental masses and their
428 lithospheric roots^{14,67,91,142-145} and has been illuminated by recent geodynamic
429 simulations^{141,146}. Starting with even present-day thicknesses of melt-depleted oceanic
430 lithosphere, lateral compression, perhaps driven by the initiation of some form of
431 subduction, can generate stable 200 km-thick mantle keels via tectonic and gravitational
432 thickening associated with cooling (**Fig. 3**). This requires the pre-existence of a strong and
433 buoyant depleted mantle lithosphere, such as that produced at rifted margins or spreading
434 axes, which, along with the crust, thickens by compression.

435

436 What is the evidence for lateral accretion and compressive thickening? Plate-scale
437 deformation imparts anisotropic fabrics on lithospheric peridotite through lattice-preferred
438 orientation of olivine, detectable with seismology¹⁴⁷. Seismic anisotropy typically occurs in

439 the upper 150 km of most cratonic lithospheres and is usually interpreted as a deformation
440 fabric created during the formation and evolution of the craton structure¹⁴⁸. A change in the
441 seismically fast axis of olivine, from horizontal at depths < 150 km to vertical at depths >
442 150 km in cratonic roots¹⁴⁹ has also been proposed as evidence for lithospheric shortening
443 via compression in making the deep roots to cratons.

444

445 The geological evidence for lateral accretion and compression during craton assembly is
446 equally compelling. Most cratonic crust is constructed from numerous individual “blocks” or
447 terranes, now juxtaposed with relatively high aspect ratios, e.g., the Superior craton. Such
448 complex large-scale linear geological fabric requires either subduction or some other lateral
449 accretion process during assembly in accretionary orogens to construct the final craton,
450 perhaps over multiple cycles ^{129,150,151}. In some cratons, e.g., Pilbara, these relationships are
451 not as clear, though most comprise different blocks/terranes that were not originally
452 contiguous. Thrust-bounded terranes characterize the assembly of the Neoarchean portions
453 of cratons^{17,129}, with large-scale continental thrust structures observed back into the
454 Mesoarchean¹⁵², clearly documenting compression of lithosphere. Compression and
455 thickening of lithosphere have thermal consequences for the crust^{153,154} and offer a
456 mechanism - via consequent crustal melting - to produce the prominent post-orogenic
457 granitic magmatism, sourced in-part or wholly by crustal melting, especially when heat-
458 producing nuclides were more abundant in the Archean. Such post-orogenic magmatism,
459 often of a potassic nature, is widespread in cratons, e.g., 2.61 to 2.58 Ga granites of the Slave
460 craton¹⁵⁵; 3.1 Ga granites of the Kaapvaal craton¹²¹; 2.67 - 2.62 Ga post-orogenic granites of
461 the Superior craton¹⁵¹; 2.6 Ga Snow Island granites of the Rae craton¹⁵⁶, and is incompatible
462 with a low geothermal gradient in the lower crust that would be expected from the presence

463 of already stable thick, cool cratonic mantle roots. High-T granulite-facies metamorphism
464 accompanies such lithospheric thickening and is among the hallmarks of “cratonisation” of
465 the crust. In the Proterozoic, craton assembly continued via lateral accretion during
466 compressive orogens, as is clearly illustrated by the evolution of the Laurentia supercraton
467 ^{148,150,157}, producing striking widespread radial seismic anisotropy in the lower craton
468 root¹⁴⁹, and the Siberian composite craton, where 1.8 Ga granulite-facies metamorphism is
469 widespread.

470

471 These features of cratons and their roots illustrate that whatever the various models
472 invoked for the genesis of their crust and mantle components, the decisive final phase of
473 assembling and stabilizing cratons, from the Archean through to the Mesoproterozoic, is
474 lateral accretion, compression and lithospheric thickening, as originally envisioned by
475 Jordan ¹⁴. It should be no surprise that the thickest parts of Earth’s lithosphere on the
476 modern Earth, outside the cratons are in zones of continental convergence^{158, 144}.

477

478 3. **Broader implications and directions**

479

480 Through the Archean, the relationship between peridotite melt-depletion ages, which
481 broadly track the melting that formed the cratonic roots, and the continental growth curve
482 [Box 3] indicates a disconnect in the genesis of the continental crust and underlying mantle
483 root. Continental crust genesis began much earlier, growing through a longer time interval,
484 at a different rate. Since the end of the Archean, the cratonic mantle depletion curve and the
485 continental growth curve are mirror images. Assembly and stabilization of thick, viscous
486 cratonic roots were critical to the preservation of Earth’s continents. This is supported from

487 the first appearance, ~2.8 Ga, of mature sediments in the stratigraphic record, with great
488 diversity in zircon ages [**Box 3**], likely tracking the first significant rise of continents above
489 sea level¹⁵⁹ due to the stabilisation of protective cratonic mantle roots in the Meso- to
490 Neoarchean.

491

492 Cratonic root formation continued to take place through the Proterozoic but the genesis of
493 highly melt-depleted peridotites that formed the craton roots waned significantly after 1 Ga
494 (**Box 3**), perhaps due to mantle cooling. However, mantle residues produced in some
495 Phanerozoic oceanic arcs are as depleted as cratonic peridotites (**Box 3**). Future cratons
496 may be underpinned by the depleted residues of arc melting, swept up during continental
497 assembly, for example, during the formation of Earth's newest continent, Zealandia, a 4.9
498 million km² block of continental crust created in Pacific arc settings and underlain by locally
499 very depleted lithospheric mantle⁹¹ extending over 150km deep where collisional
500 thickening is greatest¹⁶⁰. Understanding these recently-formed continental masses and their
501 collisional roots will be key to understanding how thick lithosphere grows, stabilizes and
502 evolves. If the majority of cratonic lithosphere is more depleted than the peridotite record,
503 mantle roots will be more buoyant, affecting hypsometry.

504

505 Defining what cratons are, and how their lithosphere evolves is critical to finding mineral
506 deposits. For instance, the spatial association between base metal deposits and the
507 transition between thick and thin lithosphere provides a basis for further exploration⁵.

508

509 Our definition of cratons as stable regions of ancient crust, protected by ≥ 150 km thick
510 lithospheric mantle roots, which have experienced minimal tectono-magmatic disturbance

511 since the end of the Mesoproterozoic (1 Ga), (**Box 1, Fig. 1**) is clearly a pragmatic
512 generalization. It is clear that cratons and their roots are periodically disrupted by rifting
513 and invaded by magmatic products. It is also evident that some cratons lose part of their
514 mantle root through disruption and weakening and then may “re-cratonise”⁸⁹, re-
515 establishing thick lithosphere and long-term stability. This re-healing process contributes to
516 the patchwork of lithosphere observed and also is associated, in some cases, with world-
517 class mineral deposits, such as the Bushveld intrusion or Premier/Cullinan diamond mine.
518 Understanding what controls the extent of these craton-disrupting events and how they
519 localize mineralization is important for mineral exploration.

520

521 More fundamentally, searching for evidence of a role for Hadean /Eoarchean melting
522 residues in the evolution of lithospheric mantle will help understand the fate of Earth’s
523 earliest lithosphere, possibly produced in a stagnant lid tectonic regime, but conspicuously
524 absent from the rock record. These early melting residues may have been quickly recycled
525 back into the upper mantle because too rapid compression/accretion of lithosphere and
526 insufficient depletion-related strengthening prevented deep lithospheric roots from
527 stabilising^{139,141}. What remain now as the long-lived lithospheric roots to continents are the
528 successful attempts to stabilize these masses. While evidence of very early melt residues
529 remains elusive, the increasing occurrence, within young lithosphere of oceanic^{92,105,106} and
530 continental⁹¹ affinity, of peridotites or mantle minerals with Archean melt-depletion ages
531 may represent the “ghosts of lithospheres past”, recycled back into the upper mantle and re-
532 housed in later lithosphere construction. Understanding what these vestigial ages tell us
533 about how lithosphere is destroyed and re-constructed, billions of years apart, should be a
534 future endeavor for geochemists.

535

536 The links between hypsometry, continental crust and its underlying lithosphere require
537 further scrutiny. The coincidence between a change in continental freeboard towards
538 presents levels by Meso- to Neoarchean times, reflecting the emergence of major
539 continental masses, and the establishment of thick cratonic roots is striking (Box 3)¹⁵⁹. This
540 implicates a central role for newly formed, stable continental masses – cratons – with
541 special attributes, i.e., thick crust comprising a strong lower crust that is dehydrated and
542 depleted of its heat producing elements by thickening-induced melting or granulite facies
543 metamorphism. How these or other attributes were able to offset any negative thermal
544 buoyancy imparted by the thick cratonic mantle roots underlying this crust requires further
545 clarification. Such cratonic lithosphere, for the first time, appears to have been able to
546 support orography similar to that of modern continents.

547

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DGP wrote the manuscript, with major contributions to the various concepts covered by JS, TC, JL, AS, LHW, JvH, PK & KS. AS provided the seismology models, HLW performed the geodynamic models, augmented by JvH. JL performed the trace element modelling.

Competing interests:

The authors declare no competing interests.

DISPLAY ITEM CAPTIONS

Box 1 *Craton definition and Earth's mantle lithosphere and crust-mantle relationships*

SEE FIGURE FOR BOX 1

Box 1a: Schematic depiction of Earth's mantle lithosphere defined here as the outer layer of Earth where heat is lost by conduction (the "tectosphere" of Jordan^{2,14}). Geotherms depict pressure - temperature relations for thermally equilibrated lithosphere, with surface heat flow in mWm^{-2} . Depth to the base of the lithosphere is taken as the intersection of the conductive geotherm with the typical mid oceanic ridge (MOR) isentrope. Adjacent idealised cross section through northern Canada starting in the SE Slave craton ($\sim 60^\circ\text{N}$) to the Sverdrup basin $\sim 80^\circ\text{N}$, with generalized locations of kimberlite fields and other magmatic rocks that provide constraints on lithospheric thickness through xenolith-derived mantle geotherms and magma geochemistry. Typical range in olivine compositions given as the fosterite content (Fo# in blue text). Lithospheric thickness varies between crust of different age, though lithospheric thicknesses of up to 200 km are present beneath crust of Proterozoic as well as Archean age^{24,89,113} (See also Fig 1). Lithospheric thickness is much thinner beneath Phanerozoic continental crust and the oceans. **Box 1b:** 3-D perspective of the lithosphere beneath southern Africa showing the Archean nuclei and Paleo to Mesoproterozoic domains of the "Kalahari craton" that comprises the Kaapvaal and Zimbabwe Archean nuclei and intervening Paleo- to Mesoproterozoic regions. Location of kimberlite pipes supplying mantle xenolith data for lithosphere ages given by circles. Colour

of circles denotes median age of lithospheric mantle (data from ^{57,74,77,98,120}). Crustal ages are representative only. Present day depth of lithosphere taken from seismology^{25,74}.

FIGURE 1 Seismic imaging of continental mantle lithosphere – defining cratonic regions.

Global S-wave tomographic slice (oceans excluded) through Earth at 150 km depth displaying the wide-spread high wave-speed anomalies of deep cratonic mantle roots - in shades of blue - that extend far beyond the boundaries of the exposed/inferred crust of 55 identifiable Archean nuclei (% Vs anomalies relative to a modified AK135 reference model²⁴). Cratons, with crust stable since ~ 1 Ga and underpinned by lithosphere > 150km thick occupy ~ 63% of the continental landmass. Asterix represents modified cratons where ancient crust is underlain by recently thinned lithosphere. Composite cratons comprise multiple Archean nuclei/cratons. The 5 Supercratons comprise multiple composite cratons. Diagonal stripes cover regions strongly influenced by subducting slabs and are not classified as cratons. For a 200 km tomographic slice, and a similar image coloured according to conventions for scientific maps, see Supplementary Material.

Box 2 Mantle peridotite density and mineralogy as a function of melt depletion: Bulk density variation, as relative % change from a fertile (un-melted) mantle peridotite (lherzolite) as a function of fraction of melt extracted, for polybaric perfect fractional melting of 3 different pressures of melt initiation: 3 GPa, 5GPa and 7 GPa, following³³. Green horizontal bars show the variation in residual (melt-depleted) peridotite mineralogy (for pyroxenes and olivine normalized to 100%) and hence lithological change, as extracted melt fraction increases. The most residual (melt-depleted) mantle peridotite is a dunite.

FIGURE 2: Estimating depth of melt extraction for lithospheric peridotites. Peridotite bulk rock molar Mg/Al ratio variation with bulk rock Yb concentration along with polybaric, perfect fractional melting curves for mantle melting beginning at different depths, following similar approaches to⁶⁷. Residual mantle melting mineralogy variations with pressure and melt fraction from³⁶. Melt fraction extracted (in wt. %) given along evolution lines for residual mantle. Data points shown for cratonic peridotites, data fields for oceanic mantle (abyssal and ocean island peridotites) and modified cratonic mantle, e.g., the eastern North China Craton. The majority of cratonic peridotites plot between the trends for melting beginning at 7 and 3 GPa melting, indicating typical starting melting pressures of 5 to 3 GPa. A much smaller fraction of cratonic peridotite residues, e.g., those from Artemisia⁸⁹ (north Slave craton) began melting at 7 GPa or deeper and are likely plume-derived melting residues.

Box 3 Dating lithospheric mantle: Lithospheric mantle age versus composition and relationship to crust

General text: Os is a compatible element, remaining in the mantle residue during melt extraction^{29,77,97}. Much higher Os concentrations in the melt-depleted peridotites versus any later metasomatic melts make Re-Os model ages more robust than many ages constraints on melting provided by incompatible element-based systems such as Rb-Sr, Sm-Nd and Lu-Hf, though the

system is not immune from disturbance^{101,103,161}. The use of whole rock platinum group element (PGE) patterns to evaluate later disturbance of peridotite Re-Os systematics by metasomatism^{49,162} is essential. Emphasis is placed on ages from the most melt depleted, sulfide-absent peridotites, with the lowest Re and Pd concentrations. The advent of in-situ Re-Os dating of sulfides in peridotites^{99,163,164}, or mechanical extraction and dissolution of whole sulfides¹⁰⁷⁻¹⁰⁹ increased the utility of Re-Os dating in peridotites, though sulfide should be absent in highly depleted peridotite compositions. For discussion of the relative merits of these approaches see^{47,101,103}.

Box 3 CAPTION: Dating mantle melt depletion, mantle ages versus craton crust ages and evolution, age versus melt residue composition for mantle peridotites: A) *Left panel* - Re-Os model age calculation assuming that during extensive (>25%) melt extraction from a peridotite, all Re goes into the melt, yielding a melt residue with Re/Os = 0. Extrapolation to a “mantle evolution curve” (black, blue and green lines – representing different models of mantle evolution) give “Re depletion model ages” (vertical axis). *Middle and right panels* - Probability density plots (bandwidth = 100 Myr) for melt depletion ages of mantle beneath cratons dominated by Archean crust (Archean cratonic mantle), Proterozoic cratons (Proterozoic cratonic mantle), Modified cratonic mantle (>1Ga crust with thin lithospheric roots, e.g., eastern North China craton, East Greenland), and Modern oceanic mantle (abyssal peridotites and Phanerozoic ophiolites). See Fig. 1 and text for craton classifications. *Far-right panel* - Model continental growth curve¹⁷ (green), plus (yellow shaded area) a moving average of the proportion of sediments with a given age distribution¹⁵⁹ for “mature sediment sources”, with a relative age spread between the modal zircon U-Pb age and the age of the sediment of 0.4 (normalized to the age of the Earth at the time of deposition). Note the appearance of these mature sediments with the dramatic increase in cratonic mantle as documented Re-depletion ages. **B)** Variation in olivine composition (box and whisker plots) in suites of mantle peridotites (single locations) versus Re-Os depletion age, where the most reliable age is based on combined Os - PGE systematics. Red boxes = mantle xenoliths erupted through crust of Archean nuclei, light-blue = mantle xenoliths erupted through crust of Proterozoic cratons, green = massif peridotites, deep-blue = Phanerozoic oceanic mantle. Note the occurrence of Archean mantle beneath Paleoproterozoic crust (e.g., locations 2 & 7) and Paleoproterozoic mantle beneath Archean crust (locations 16, 17, 19). Ages for locations where lithosphere formation is <100 Ma are given in arbitrary order in the expanded blue inset. Key to location numbers is given in the Supplementary Information. Salmon-coloured curves denote different present-day Urey ratio curves¹⁶⁵ (the ratio of internal heat generation in the mantle over the internal heat flux).

FIGURE 3: Geodynamic modelling of possible craton formation processes: Lateral compression and plume residue dispersal.

Panels A-B: Lateral compression (pure shear) model of Wang et al.¹⁴¹ showing the lithosphere temperature and melt depletion fields at ~52 Myr (immediately after shortening) and at 503 Myr, after the start of the model. After initial compressional shortening the lithosphere thickens by cooling, becoming denser to the point where negative thermal buoyancy starts to exceed the inherent chemical buoyancy and results in further thickening. The melt depletion field is

converted from the compositional field by assuming a maximum melt depletion of 35% for mantle peridotite.

Panels C-D: Dynamic model of plume upwelling tracking the path of melt residues. The temperature and melt depletion fields for residual peridotite comprising the lithospheric mantle keel shown at ~33 Myr and ~102 Myr after model initiation. Isotherms plotted at T=350 °C, 550 °C, 900 °C, 1300 °C. The calculation includes rheological strengthening due to melt depletion in the residual mantle. Note how the depleted plume residues rapidly disperse, partially accumulating beneath lithospheric traps, some being recycled back to the asthenosphere. Rheological strengthening due to melt depletion does not prevent plume melting residues from dispersing laterally by plume introduced flow, although a significant fraction of these buoyant residues remain in the uppermost mantle and may become accreted to craton roots during accretion or beneath much younger continents⁹¹. Model colours follow recommendations of ¹⁶⁶ scientifically derived colour maps. Model details in Supplementary Methods.

On-line Methods

Numerical models

We use finite element code “Citcom” ¹⁶⁷⁻¹⁶⁹ to solve the thermochemical flow with extended Boussinesq approximation^{170, 171}. A particle-tracking technique is used to track the movement of different chemical fields, including cratonic root and melt depletion. For detailed description of numerical method, including governing equations, rheology, and mantle plume setup, please refer to Wang et al. 2015¹⁷². The temperature plot is the model temperature after removing the adiabatic gradient, in order to make the plume clear in the temperature field.

We further include melting of mantle peridotite and depletion-dependent buoyancy and strengthening in the model. The anhydrous melting parameterization of ¹⁷³ is used for mantle melting. The amount of melting is calculated for each particle, and any melt generated is removed immediately, assuming instantaneous extraction to the surface. The depletion field is updated with accumulated melting degree with each particle, which is then advected with particles just as with other chemical fields.

We apply an exponential change of mantle peridotite due to melt depletion¹⁷⁴

$$\rho = \rho_0 \exp(\alpha_d D)$$

where α_d defines the rate of density change due to melt depletion D. $\alpha_d = -0.0003$ is used as reference value, which leads to 1.04% of density change at a melt depletion of 35%.

A composite rheology of non-Newtonian and Newtonian viscosity is used, with consideration of composition-dependent effects. The strengthening factor for the cratonic root and plume residue can be calculated as:

$$\Delta\eta = \Delta\eta_0^{\min(1, \frac{C}{C_\eta})}$$

where C is the composition field, including cratonic root and melt depletion. C_η is the composition threshold for the maximum strengthening. A “constant strain-rate” definition is used for $\Delta\eta$, which would lead to an increase of non-Newtonian viscosity by $\Delta\eta^n$ in the definition of “constant

stress”¹⁷². For depletion related rheological strengthening, we use a strengthening factor $\Delta\eta = 3$ in the reference model and vary $\Delta\eta$ between 1 and 10 in other models.

Model setup: The model domain is 2640 km wide and 660 km deep. A no-slip boundary condition is used on the top boundary, while other boundaries are at a free-slip condition. We use a total number of 384-by-96 elements with vertical mesh refinement at depth between 60 ~240 km depth. This provides a spatial resolution of 6.9 km by 4.3 km in this region. To track the composition field, ~1.7 million particles are used in the model domain, resulting in an average particle density of ~47 particles per element.

We start the models with a quasi-steady state thermal field that has ~100 km thick lithosphere. A hot plume with a maximum temperature anomaly of 250 °C rises up from the basal 660 km depth and the residual mantle from plume melting is produced and dispersed. The mantle potential temperature is 1550 °C, which leads to about 1800 °C at the base (660 km) due to adiabatic heating. However, the adiabatic thermal gradient is removed in Fig. 3C-D in order to emphasize the thermal anomaly introduced by the mantle plume.

Fig. 3C-D shows how the plume residue greatly disperses with the plume induced flow even with the rheological strengthening of 3 due to melt depletion. We varied the rheological strengthening factor from 1 to 10 and found that it did not prevent the dispersion of the residual mantle. The videos (PrTempDepletion.avi and PrTempVisc.avi) show the evolution of temperature, depletion, and viscosity field through time.

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Archean Nuclei	
A, Atlantic	Nc, North China*
Aa, Angola	Ni, Nimrod
Ad, Aldan	Np, Napier
An, Anabar	Om, Omolon*
Av, Aravelli	P, Pilbara
B, Bundelkand	R, Rae
Ba, Bastar	Ra, Rio Apa
Ca, Carajas	Re, Reguibot
D, Dharwar	Rp, Rio de la Plata
G, Gawler	S, Sask
Ga, Guiana	Sa, Slave
Gb, Gabon	Sf, Sao Francisco
Gr, Grunehogna	Si, Singhblum
H, Hearne	Sl, Sao Lucia
Ha, Hall	Sm, Sarmatia
K, Kimberley	Su, Superior
Ka, Karelia	T, Tanzania
Kb, Kibalia	Ta, Terra Adélie
Ke, Kemp Land	Tm, Tarim Basin
Ko, Kola	V, Volga-Uralia
Kp, Kaapvaal	VH, Vestfold Hills
Ks, Kasai	Uk, Uweinat-Kamil
La, Lius Alves	W, Wyoming*
L, Lewisian	Y, Yilgarn
M, Mackenzie	Ye, Yemen
Md, Madagascar	Yz, Yangze*
Mn, Man	Z, Zimbabwe
N, Nain	
*Modified craton	
Composite cratons	
Super-cratons

Accretion model

Plume model

SI Guide

Supplementary Notes:

Supplementary Figures 1a & b and captions – a: Global seismic tomography section at 150km coloured according to recommendations for scientific maps; b: Regionalisation of lithospheric seismic properties.

Supplementary Location Notes to Box 3a (locations used in craton and oceanic age plots) & Box 3b (locations used in age versus Mg# plot).

Supplementary Videos (4)

LateralCompression_TempDepletion.avi

Video of dynamic model of lateral compression of cratonic mantle showing temperature and degree of melt depletion in the mantle. See “On-line Methods” for model methodology and caption to Figure 3 for further details.

LateralCompression_TempVisc.avi

Video of dynamic model of lateral compression of cratonic mantle showing temperature and viscosity in the mantle. See “On-line Methods” for model methodology and caption to Figure 3 for further details.

PlumeResidue_TempDepletion.avi

Video of the dynamics of the dispersion of mantle melting residues produced by mantle plumes showing temperature and extent of melt depletion in the mantle. See “On-line Methods” for model methodology and caption to Figure 3 for further details.

PlumeResidueTempVisc.avi

Video of the dynamics of the dispersion of mantle melting residues produced by mantle plumes showing temperature and viscosity in the mantle. See “On-line Methods” for model methodology and caption to Figure 3 for further details.

SUPPLEMENTARY NOTES

Supplementary Figure 1a: Seismic imaging of continental mantle.

Global S-wave tomographic slice (oceans excluded) through the Earth at 150 km depth. The scale bar displays % Vs anomalies relative to a modified AK135 reference model [References 23 and 24 of the main manuscript]. Colour scale is the “Vik” scheme derived using the recommendations for colour scales on maps [Reference 166 of the main manuscript].

Supplementary Figure 1b : Seismic tectonic regionalisation of lithosphere.

Tectonic regionalisation [using the method of reference 23 – Main Manuscript] of the seismic data to objectively classify seismic velocities into regions of similarity within the Earth's upper mantle; falling out naturally from this is the division of continents from oceans (not shown), and the subdivision of continents into cratonic very deep lithosphere (green) that broadly coincides with the Archean crustal nuclei to cratons, cratonic deep lithosphere (150 km thick or more – blue colour), broadly coinciding with the Super-cratons and Composite cratons defined in Fig. 1 of the main manuscript, and non-cratonic mantle underpinned by lithospheric mantle extending to <150km to the lithosphere-asthenosphere boundary (grey shading). Diagonal stripes cover regions with apparent thick lithosphere that are strongly influenced by subducting slabs and are not classified as cratons. The boundaries of the **cratons as a whole**, as defined as the edges of the blue shaded regions, underpinned by lithosphere > 150km thick, **comprise 63% of the total continental landmass**. **Cratonic nuclei** with the deepest lithospheres, from this classification occupy ~ **34% of the total continental landmass**.

Box 3a: Locations grouped by type of craton in Re-Os model age compilation

Names are either the name of a given mine, or the location of the host kimberlite, lamproite or basalt

Archean Cratonic Mantle:

Kalahari craton Archean nuclei – Letseng Mine, Lhiqhobong Mine, Matsoku, Mothae Mine, Thaba Putsoa, Newlands Mine, Monastery Mine, Kimberley Mines (Bultfontein), Jagersfontein Mine, Finsch Mine, Murowa Mine.

Karelian craton – Kaavi, Kaavi-Kupio

Rae craton – Somerset Island, Repulse Bay, Pelly Bay, Darby

Slave craton – Jericho Mine, Diavik Mine, Ekati Mine, Artemisia, Gahcho Kué

North Atlantic Craton – Chidliak Mine (Baffin Island), Safartog (W. Greenland), Nigerlikasik and Pyramidefjeld (W. Greenland)

Superior craton – Attawapiskat

Eastern North China craton – Fuxian, Mengyin (Pre-lithosphere thinning kimberlites).

Wyoming craton – Eagle buttes, Williams

Sask craton – Fort a la Corne

Proterozoic Cratonic Mantle

Kalahari composite craton – Venetia Mine, Orapa Mine, Lethlakane, Namaqua-Natal terrane kimberlites (Ramatseliso, East Griqualand, Abbotsford, Melton Wold, Uintjesberg, Markt, Hebron, Hoedkop, Gansfontein, Klipfontein), Rehoboth terrane kimberlites (Gibeon field)

Congo composite craton – Usagaran terrane (Labait, Tanzania)

Western Australian composite craton – Argyle Mine

Laurentia Super-craton – Mackenzie craton (Parry Peninsula), Victoria Island; Hearne Terrane (Buffalo Head Hills), Central Plains Orogen (Sloan)

South American Super-craton - Alto Paranaíba

Siberian composite craton – Udachnaya Mine, Obnazhennaya

Modified Cratonic Mantle

South East Siberia – Tok

East China craton – Longquanlongwan, Kuandian, Hanuoba, Qixia, Yangyuan, Jining, Datong, Fansi, Fushan, Hebi, Dalongwan, Penglai, Shanwang, Baekryeong Island, Jeju Island, Pyeongtaek

Colorado Plateau - Big Creek

Mojavia – Cima

West Greenland – Ubekent Ejland

East Greenland - Wiedemann Fjord

Phanerozoic Oceanic Mantle (Abyssal, Arcs, Ophiolites)

Abyssal – Mid Atlantic Ridge Hole 1274A (ODP Leg 209), Kane Fracture Zone, SW Indian and American-Antarctic Ridges, MAR 15 20 FZ Cruise: Academic Boris Petrov, SWIR Atlantic II FZ Cruise: Robert Conrad 27-9, SWIR Du Toit FZ Cruise: Protea 5, Gakkell ridge (Arctic Ocean), Lena Trough, Arctic basin, Mid Atlantic Ridge (MAR) Saint Paul Fracture Zone

Ophiolites – Tiebaghi massif, New Caledonia; Ouassé Bay, New Caledonia; Me Maoya massif, New Caledonia; Massif du Sud, New Caledonia; Poum, New Caledonia; Babouillat, New Caledonia; Poum, New Caledonia; Babouillat, New Caledonia; Kopeto, New Caledonia; Mamonia complex, Cyprus, Troodos, Cyprus; Oman Ophiolite, Shetland Ophiolite, Taitao ophiolite; southern Chile, Koycegiz,

Turkey; Tekirova, Turkey, Marmaris, Turkey; Macquarie Island, Australia, Ligurian ophiolites, Italy; Totalp, Austria; Purang ophiolite, southwestern Tibet; Central Tibet - Bangong-Nuijiang suture zone; Yarlung-Zangbo ophiolite, Tibet;
Arcs – Kamchatka (Bakening, Avachinsky, Valovayam; Cascade arc (Simcoe); Japan arc (Ichinomegata); SW Japan (Kurose); Bismark Archipelago (PNG); Izu-Bonin-Mariana Forearc, Conical seamount.

Box 3b: Locations plotted on Age versus olivine composition plot

- 1) Siberia (Udachnaya, Archean samples, plus Obnazhennaya)
- 2) Argyle, Australia
- 3) Kalahari craton - Kaapvaal nucleus (excluding Kimberley & Premier)
- 4) Tanzania craton
- 5) Kalahari Craton - Kaapvaal nucleus (Kimberley)
- 6) Wyoming craton
- 7) North Atlantic Craton (E, W. and SW Greenland)
- 8) Slave craton
- 9) Congo craton
- 10) Kalahari Craton - Zimbabwe nucleus (Murowa-Sese)
- 11) North China craton (Archean samples)
- 12) Rae craton - Somerset Island
- 13) Kalahari craton - East Griqualand
- 14) Kalahari craton - Namaqualand
- 15) Kalahari craton - Namibia
- 16) Kalahari Craton - Kaapvaal nucleus (Premier)
- 17) Siberia (Udachnaya, Proterozoic samples)
- 18) Pyrenees orogenic massifs
- 19) Parry Peninsula & Victoria Island
- 20) North China Craton (Post-Archean samples)
- 21) Ronda Massif, Spain
- 22) China - Songshugou
- 23) San Carlos - USA
- 24) Bay of Islands Ophiolite
- 25) Iwanadake ophiolite, Japan
- 26) Dun Mountain ophiolite, New Zealand
- 27) Ontong Java Plateau, Pacific Ocea
- 28) Anita peridotite, New Zealand
- 29) Mariana Arc, Pacific Ocean
- 30) Kamchatka, Russia
- 31) South Sandwich Islands, S. Atlantic Ocean
- 32) Lihir Island, Papua New Guinea
- 33) Japan arc-related mantle xenoliths - various locations