

Delamination and recycling of Archaean crust caused by gravitational instabilities

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An extensive, thick MgO-rich primary crust underlain by highly residual mantle must have formed during the Archaean as a consequence of higher ambient mantle potential temperatures¹. However, the preserved volume of this crust is low suggesting much of it was recycled². Further, the tonalite–trondhjemite–granodiorites that dominate exposed Archaean crust cannot have been generated directly from MgO-rich primary crust since a hydrated low-MgO basalt source is required³. Here we show that the thermodynamically stable mineral assemblages expected at the base of fully hydrated and anhydrous MgO-rich crust 45 km thick make it denser than the complementary underlying residual mantle. We use 2-D geodynamic models to explore the fate of this gravitationally unstable crust. Our results demonstrate that magmatically-overthickened MgO-rich crust, whether fully hydrated or anhydrous, could have delaminated by Rayleigh–Taylor instabilities for mantle potential temperatures > 1500–1550 °C, depending on rheology. The dripping instabilities generate return flow of asthenospheric mantle that melts adiabatically producing additional primary crust. Melting of overthickened and dripping MgO-rich crust and intracrustal fractionation of primary magmas both may produce the hydrated

27 **MgO-poor basalts necessary to source tonalite–trondhjemite–granodiorite melts, but the**
28 **associated ultramafic residues now must reside in the mantle.**

29 Understanding the formation and reworking of Earth’s early crust is important for models
30 of mantle evolution. Earth’s primary crust is a product of partial melting of the mantle. Most
31 primary (oceanic) crust on modern Earth is produced at spreading ridges, is on average 6–7 km
32 thick and is composed of mid-ocean ridge basalts (MORB; 5–8 wt% MgO), the result of partial
33 melting at melt fractions (F) of 0.08–0.1 and mantle potential temperatures (T_p) of 1280–1400
34 °C⁴. This crust is recycled by subduction during which it is converted to eclogite. On the early
35 Earth things were undoubtedly different. The occurrence of voluminous tonalite–trondhjemite–
36 granodiorite (TTG) crust and komatiites, which are largely restricted to the Archaean⁵, and the
37 volume of continental crust produced during the Archaean⁶ are consistent with a hotter Earth.
38 Although the thermal evolution of Earth and the mechanism of heat loss remain matters of
39 debate^{7,8}, higher T_p during the Archaean would have led to more extensive melting of the mantle
40 and production of a thick (up to 45 km) MgO-rich primary crust^{1,2}. Komatiite magmatism
41 associated with plumes and excess T_p may have contributed to the early crust, but the volume of
42 komatiite in greenstone belts (0–20 vol.%) suggests this played only a minor role⁹.

43 Based on the petrology and geochemistry of volcanic rocks, the composition of model
44 primary melts generated by varying degrees of partial melting of fertile peridotite can be
45 calculated with reference to experiments⁴, from which the composition of the complementary
46 residues are constrained by mass balance^{1,2}. Near-pristine xenoliths of harzburgite that sample
47 the Archaean lithospheric mantle are highly depleted in FeO, Al₂O₃ and CaO relative to fertile
48 mantle and have compositions that correspond well with modelled residua for melt fractions of
49 0.25–0.45² and primary crust 25–45 km in thickness². Samples of high MgO non-arc basalts
50 from Archaean greenstone belts have compositions consistent with their derivation from primary
51 melts¹ (Fig. 1). However, the volume of these rocks in the exposed geologic record is low.

52 Assuming these non-arc basalts are representative of primary crust^{1,2}, then most of this crust is
53 missing, which suggests it has been recycled or sequestered at depth in the mantle.

54 A plot of MgO content versus age for non-arc basalts¹ shows that ambient T_p was >1500
55 °C during the Archaean, with a maximum in the Mesoarchaeon (Fig. 1a). The inferred onset of
56 secular cooling of the mantle broadly coincides with a marked decrease in the rate of crustal
57 growth around 3 Ga⁶, and may record the transition to dominantly subduction-driven plate
58 tectonics on Earth^{10,11}. Prior to this, most primary crust may have formed by intra- and over-
59 accretion of melts above upwelling mantle driven by smaller wavelength convection¹¹. With a
60 higher basal heat flux and increased radiogenic heat production, temperatures close to the base of
61 thick primary crust in the Archaean may have exceeded 1000 °C¹⁰, conditions under which
62 partial melting of ultramafic rocks is likely if they were hydrated^{12,13}.

63 The range in T_p for any given age suggested by the petrological data (Fig. 1a) is similar to
64 temperature variations in mantle upwelling beneath modern mid-ocean ridges⁴, and would have
65 given rise to variations in the composition and thickness of Archaean primary crust (Fig. 1c) and
66 associated residual mantle. Such variations may have allowed different tectonic styles to operate
67 in different parts of the ancient Earth¹¹. Extreme ambient mantle temperatures ($T_p > 1600$ °C)
68 could have produced *in situ* primary crust 40–45 km thick² (Fig. 1c). Subduction similar to that
69 observed in contemporary plate tectonics may not have been possible under these
70 conditions^{8,11,14}.

71 Here we test the postulate that thick primary crust produced in the early Archaean could
72 have become gravitationally unstable with respect to the underlying mantle residues so that if
73 locally thickened it became negatively buoyant and could have delaminated by Rayleigh–Taylor
74 (RT) instabilities. To do this we model the equilibrium (minimum Gibbs free energy) mineral
75 assemblages developed near the base of the crust for a wide range of metamorphosed (hydrated
76 and anhydrous) primary crust compositions (Figs 2a,b and Supplementary Figs S1–S4) and their
77 complementary (anhydrous) residues (Fig. 2c and Supplementary Figs S5 and S6) that account

78 for secular changes in T_p , F and crustal thickness (see Methods). First, we calculate the stability
79 of equilibrium mineral assemblages for a temperature of 1000 °C (Figs 2a,b and Supplementary
80 Figs S1, S3), which corresponds to that expected near the base of 30–45 km thick crust based on
81 the apparent thermal gradients recorded by Mesoarchaeon to Neoarchaeon metamorphic rocks¹⁰
82 and approximates the solidus temperature of hydrated ultramafic primary crust, inferred to lie
83 between the fluid-absent solidi for amphibolite and amphibole-bearing MORB pyrolite and
84 wehrlite^{13,15}. To assess the temperature sensitivity of the results, we also calculate phase
85 equilibria at 900 °C (Fig. 2a and Supplementary Figs S2, S4). We use the results to calculate the
86 density contrast at the crust–mantle interface for the range of compositions and temperatures
87 modelled to assess the stability of the crust relative to the underlying residual mantle (Fig. 3).
88 We incorporate the density data with a parameterized melting and melt extraction algorithm into
89 a 2-D thermomechanical model (see Methods) to evaluate the dynamical plausibility of crustal
90 recycling by RT instabilities for end-member simulations involving either hydrated or anhydrous
91 crust (Supplementary information, 3 Geodynamic numerical modelling), **wet or dry mantle**, and
92 various crustal thicknesses, T_p and **Moho temperatures** (Fig. 4 and Supplementary Figs S8–S12).
93 By integrating results from both thermodynamic and 2-D numerical modelling, we draw several
94 important inferences about Archaean tectonics, partial melting of delaminated primary crust,
95 intracrustal differentiation and the generation of TTG magmas.

96 The variation in density between fully hydrated and anhydrous crust is shown in Figs 3a
97 and b, the density of the complementary residual mantle in Fig. 3c, and the effect on the potential
98 instability of the crust in Fig. 3d. Thick anhydrous crust and hydrated primary crust with high
99 MgO (> 21–22 wt%), generated at extreme T_p (> 1600 °C), would have been gravitationally
100 unstable at its base (Fig. 3d). Primary melts with lower MgO (< or <<21 wt%; T_p <1600 °C)
101 would have produced lower crustal thicknesses (Fig. 1c) and overthickening would be required
102 for the base to become gravitationally unstable (Fig. 3d).

103 Here we illustrate the results of the geodynamic modelling with reference to one
104 experiment for the evolution of primary crust during the early Archaean (Fig. 4). Thick MgO-
105 rich primary crust forms above hot upwelling mantle by near-isentropic melting. Local
106 thickening causes material at the base of the crust to become gravitationally unstable. Using 2-D
107 thermomechanical modelling we are able to show that delamination of this over-thickened
108 primary crust occurs by RT instabilities once the viscosity is sufficiently low (Fig. 4) and occurs
109 for a range of crustal thicknesses and geotherms (see Methods and Supplementary information, 3
110 Geodynamic numerical modelling). **A scaling analysis reveals that the two requirements for the**
111 **instabilities to occur are a sufficiently thick negatively buoyant lower crust and low effective**
112 **viscosities of the uppermost mantle and lower crust (see Supplementary information 3.2.3).** For
113 realistic rheologies this mechanism starts to operate at T_p 1500–1550 °C but is particularly
114 effective at $T_p \geq 1600$ °C, consistent with predicted upper mantle temperatures in the Archaean
115 (Fig. 1). Sinking primary crust could have partially melted directly to produce basalt¹⁶ and/or
116 reacted with and refertilised the underlying residual mantle to induce further melting^{16,17}. Mantle
117 return flow would have caused additional adiabatic melting and further crustal thickening (Fig. 4
118 and Supplementary information, 3 Geodynamic numerical modelling). At the predicted
119 temperatures, partial melting near the base of overthickened hydrated primary crust would result
120 in intracrustal differentiation. These secondary magmas would have had lower MgO contents
121 than the primary crust.

122 The depths at which the base of the primary crust becomes gravitationally unstable
123 represent maxima. Firstly, the modelled crust is taken to be homogeneous. However, similar to
124 present-day primary crust (intra-oceanic arc crust¹⁸, oceanic plateaux^{19,20} and oceanic crust²¹),
125 primary crust in the Archaean is likely to have differentiated into lower MgO and higher MgO
126 (cumulate-rich) portions¹⁹. Secondly, the modelled residua (Fig. 2c) are the average
127 compositions of the melt-depleted mantle and do not account for the decrease in the degree of

128 partial melting with depth. Consequently, the residual mantle immediately underlying the crust
129 would have been more depleted, and marginally less dense, than modelled.

130 We emphasise that high MgO (> or >>18 wt% MgO) primary crust produced during the
131 Archaean would have been ultramafic not basaltic¹⁴. Compositions in the lower portion of over-
132 thickened primary crust would have been silica-undersaturated metapyroxenite or
133 metahornblendite, additionally containing garnet at higher pressures (>1.0–1.5 GPa) or at lower
134 degrees of hydration (Figs 2a,b and Supplementary Figs S1 and S3). High MgO garnet
135 metapyroxenite (16–23 wt% MgO) xenoliths of Archaean age have mantle-like oxygen isotopic
136 compositions²²⁻²⁴ and generally record high equilibrium temperatures²². They are
137 compositionally and mineralogically similar to modelled high MgO compositions²² and may
138 represent near pristine primary crust that foundered into the depleted mantle²⁴. By contrast, low
139 MgO eclogite xenoliths have omphacitic clinopyroxene, highly variable oxygen isotopic
140 compositions and may contain accessory kyanite, quartz/coesite, K-feldspar and/or corundum²³.
141 They likely represent highly fractionated crustal compositions with a multi-stage evolutionary
142 history²².

143 The high SiO₂, Na/K, Sr/Y and depleted HREE contents of sodic Archaean TTGs suggest
144 that most were derived by partial melting of garnet amphibolite^{3,25}. At 1000 °C garnet
145 amphibolite is stable only in hydrated primary crust with MgO < or << 18 wt% at $P > 1.0$ GPa
146 (Fig. 2a, Figs S1), and this stability is further reduced at 900 °C (Fig. 2a, Figs S3). Thus, TTG
147 melts cannot have been generated from partial melting of unmodified Archaean high-MgO
148 primary crust, which would result in basaltic or nephelinitic melt compositions¹⁴, but were
149 derived from crust of basaltic composition³. This basaltic crust could have been produced by
150 differentiation of primary melts by crystal fractionation¹⁴ or by partial melting of over-thickened
151 primary crust. Thus, a two-stage process is required for the production of Archaean TTG
152 magmas.

153 Although the thermal structure of the mantle in the Hadean and Eoarchaeon is poorly
154 constrained, primary crustal compositions indicating $T_p > 1600$ °C appear to be limited to
155 occurrences between c 3.5 Ga (or earlier) and c 1.9 Ga (Fig. 1a). This interval spans the onset of
156 production of ultra-thick primary crust and tectonics dominated by RT instabilities to the
157 transition to a dominantly modern-style of subduction, and may define an early tectonic regime
158 with globally variable tectonic styles controlled by the spatial range in T_p at any point in time¹¹.
159 Rather than subduction-driven plate tectonics, delamination by RT instabilities of overthickened
160 primary crust was responsible for the production and stabilization of the bulk of the continental
161 crust.

162

163 **Methods**

164 The phase equilibria are presented as isothermal pressure (P) – composition (X) pseudosections
165 calculated in the chemical systems $\text{Na}_2\text{O}-\text{CaO}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{O}_2$
166 (NCFMASHTO) for the primary melt compositions (Supplementary Figs S1, S3, S7 and S9) and
167 $\text{CaO}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{O}_2$ (CFMASO) for the complementary residue compositions
168 (Supplementary Fig. S5 and S11) using *Perple-X*²⁶ with internally consistent thermodynamic
169 data for end members²⁷ and the most recently calibrated activity–composition models for solid-
170 solution phases²⁸ (see Supplementary information, 2 Phase equilibria modelling). The modelled
171 compositional range for primary crust is a binary mixture between the primary melt composition
172 for MORB ($X = 0$) and a fictive end member composition (MgO of 25 wt%; $X = 1$) as a proxy
173 for primary crust generated at extreme T_p (Fig. 1). The composition (in mol.%) of these end
174 members in terms of the components SiO_2 , TiO_2 , Al_2O_3 , FeO , MgO , CaO , Na_2O , O_2 is: 49.49,
175 0.66, 10.11, 6.69, 17.72, 12.72, 2.28, 0.33 ($X = 0$) and 42.96, 0.60, 2.61, 10.09, 34.97, 7.53, 0.99,
176 0.25 ($X = 1$). The end member compositions of the binary bulk compositional range for the
177 complementary residue in terms of the components SiO_2 , Al_2O_3 , FeO , MgO , CaO , O_2 is: 38.25,
178 1.61, 5.80, 51.66, 2.53, 0.15 ($X = 0$) and 35.76, 1.28, 4.07, 58.52, 0.27, 0.10 ($X = 1$). All

179 calculations use an $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ ratio of 0.1 (see Supplementary information, 2 Phase
180 equilibria modelling). The phase equilibria modelling considers the phases olivine (ol), garnet
181 (g), clinopyroxene (cpx), orthopyroxene (opx), hornblende (hb), plagioclase (pl), quartz (q),
182 ilmenite (ilm), rutile (ru), spinel (sp), magnetite (mt), hematite (hem) and H_2O . Based on the
183 phase equilibria, densities are calculated for hydrated primary crust, anhydrous primary crust and
184 the complementary residues to determine the relative densities of hydrated or anhydrous crust
185 compared to the underlying residual mantle at 1000 and 900 °C.

186 The parameterized 2-D thermomechanical models solve the governing conservation
187 equations of mass, momentum and energy of slowly creeping fluids on geological timescales
188 using the finite element code MILAMIN_VEP^{29,30}. The code employs realistic rock densities
189 that have been computed as a function of P and T for the given rock types described above, as
190 well as laboratory-constrained creep laws. In addition, we have implemented partial melting of
191 mantle lithosphere and underlying fertile mantle, and added a simplified melt extraction
192 algorithm which removes melt from the mantle as soon as >5% melt is present and emplaces it
193 on top of the crust. We have performed systematic studies to understand the sensitivity of the
194 results to input parameters, and developed a scaling analysis that demonstrates that the
195 underlying physics of the models is governed by density-driven instabilities that erode the base
196 of the crust in relatively small pulses once the viscosity of the mantle lithosphere underneath the
197 crust is sufficiently low (see Supplementary information, 3 Geodynamic numerical modelling).

198

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271

272 **Author Contributions**

273 M.B. and T.E.J. developed the project; T.E.J. calculated the phase diagrams and B.K. developed
274 and ran the numerical models. All authors discussed the results and were involved in writing the
275 paper.

276

277 **Additional information**

278 Supplementary Information is available in the online version of the paper. Reprints and
279 permissions information is available online at www.nature.com/reprints. Correspondence and
280 requests for materials should be addressed to T.E.J.

281

282 **Competing financial interests**

283 The authors declare no competing financial interests.

284

285 **Figure legends**

286

287 **Figure 1 | Calculated primary melt compositions for Precambrian non-arc basalts. a**, MgO
288 (mantle potential temperature) vs age (see Supplementary information, supplementary notes for
289 Fig. 1a); ambient mantle potential temperature peaked during the Archaean. **b**, MgO (mantle
290 potential temperature) vs FeO; the primary melt compositions for MORB ($X = 0$) and a fictive
291 composition with 25 wt% MgO ($X = 1$) define end-members of the binary compositional range
292 used in the thermodynamic modelling. **c**, MgO (mantle potential temperature) vs melt fraction
293 (F) based on the relationship $F = 1.456 - 2.189(\text{FeO}/\text{MgO})$; ~ 1 km of crust is produced per 1
294 vol.% melting².

295

296 **Figure 2 | Results of thermodynamic modelling of primary crust and complementary**
297 **residues at 1000 °C**. Metamorphic mineral assemblage stability fields for hydrated **a**, and
298 anhydrous **b**, primary crust, and **c**, complementary residues (based on Supplementary Figs S1–
299 S6; in **a**, note reduced stability field for amphibolite at 900 °C); line shows approximate
300 thickness of primary crust in relation to MgO content. Primary melt compositions for
301 Precambrian non-arc basalts and inferred mantle potential temperatures (T_p) are shown along the
302 abscissa. Extrapolated pressures at 1000 °C (right-hand ordinate) are based on apparent thermal
303 gradients recorded by metamorphosed Precambrian rocks¹⁰.

304

305 **Figure 3 | Density of primary crust and complementary residues.** Densities calculated at
306 1000 °C plotted against MgO content for **a**, hydrated primary crust, **b**, anhydrous primary crust
307 and **c**, the complementary residues. **d**, Shows the relative density of hydrated crust compared to
308 the underlying residual mantle at 1000 °C against MgO of the primary crust with the equality
309 line for anhydrous crust at 1000 °C and those for 900 °C (based on Supplementary Fig. S7)
310 superimposed. The solid (hydrated) and dashed (anhydrous) lines separate stable primary crust
311 (below) from potentially unstable primary crust (above).

312

313 **Figure 4 | Results of geodynamic modelling: snapshots from an experiment with 45 km**
314 **thick initial primary crust and T_p of 1600 °C.** **a**, Model configuration and initial conditions
315 (dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink-
316 brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; yellow, melt-bearing
317 asthenosphere). **b**, Shows local thickening of the initial primary crust by addition of new primary
318 crust (green) and delamination of negatively buoyant initial primary crust. **Large-scale mantle**
319 **flow is caused mainly by the positive buoyancy of the melt-bearing asthenosphere; crustal drips**
320 **induce smaller circulation patterns.** **c**, Shows density (left) and viscosity (right) for the left-hand
321 and right-hand halves of **b**, respectively. **d**, Shows local thickening and delamination of
322 negatively buoyant new primary crust.

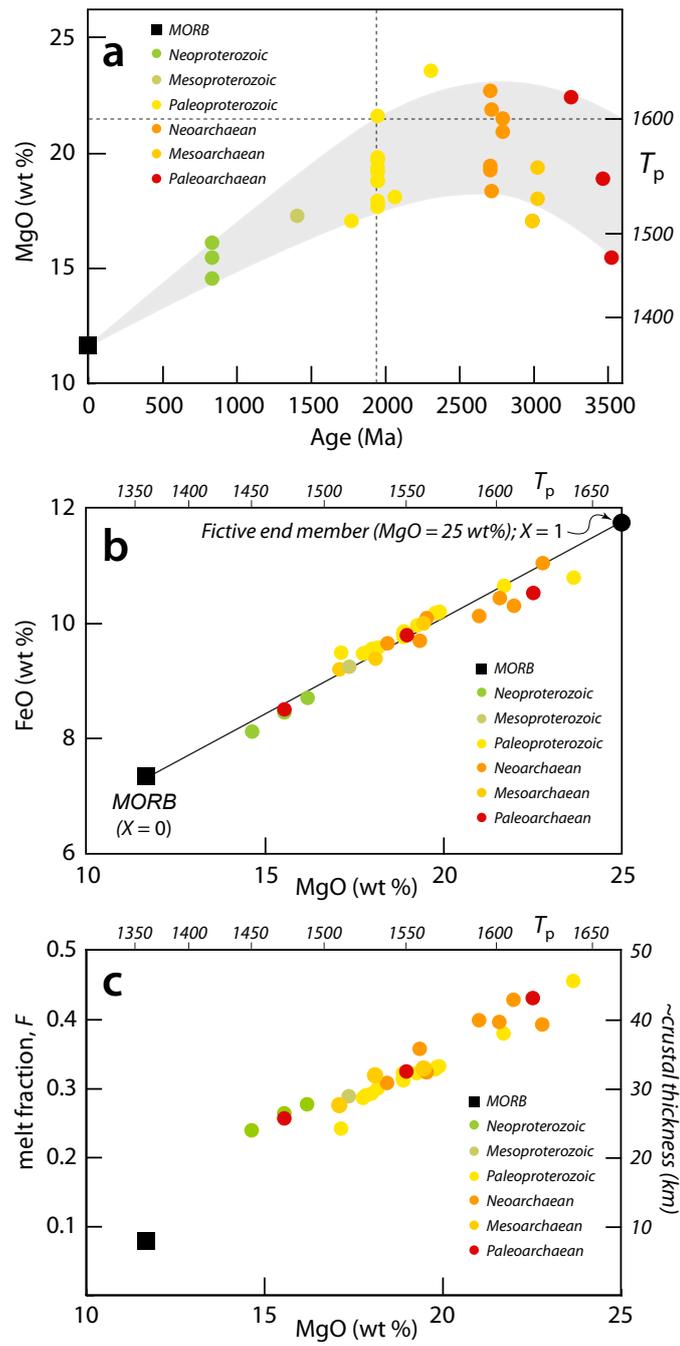


Figure 1

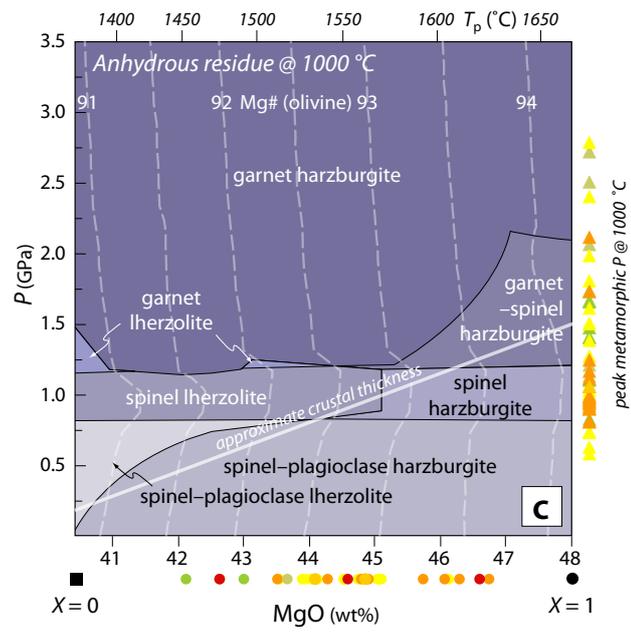
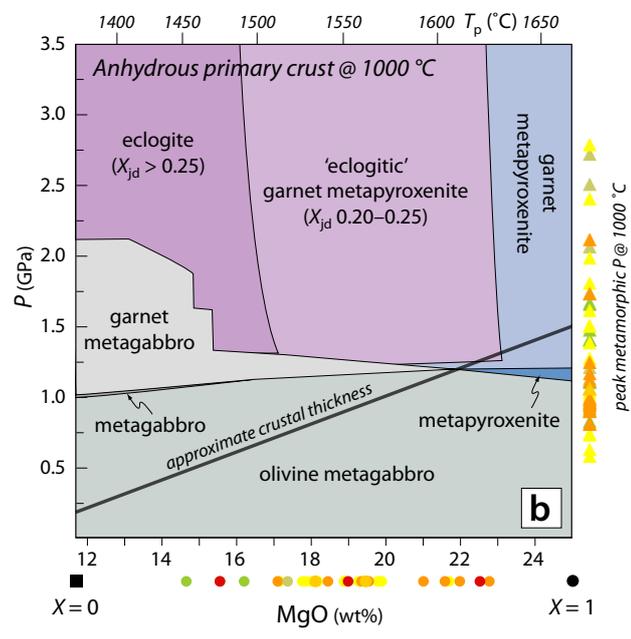
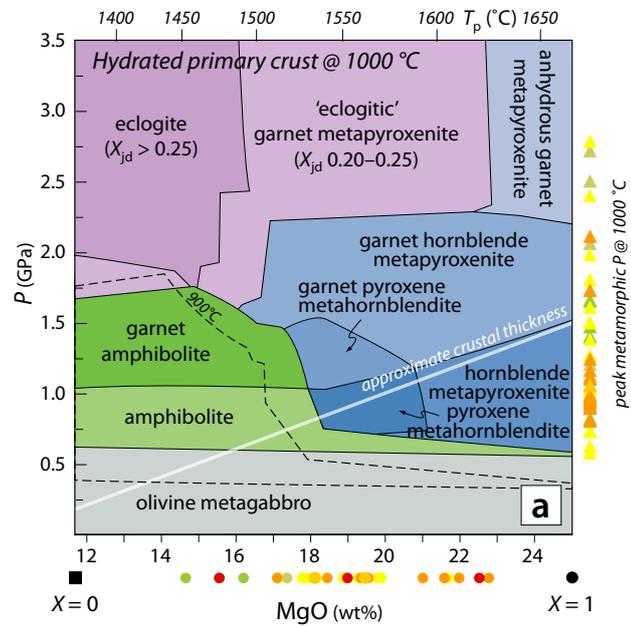


Figure 2

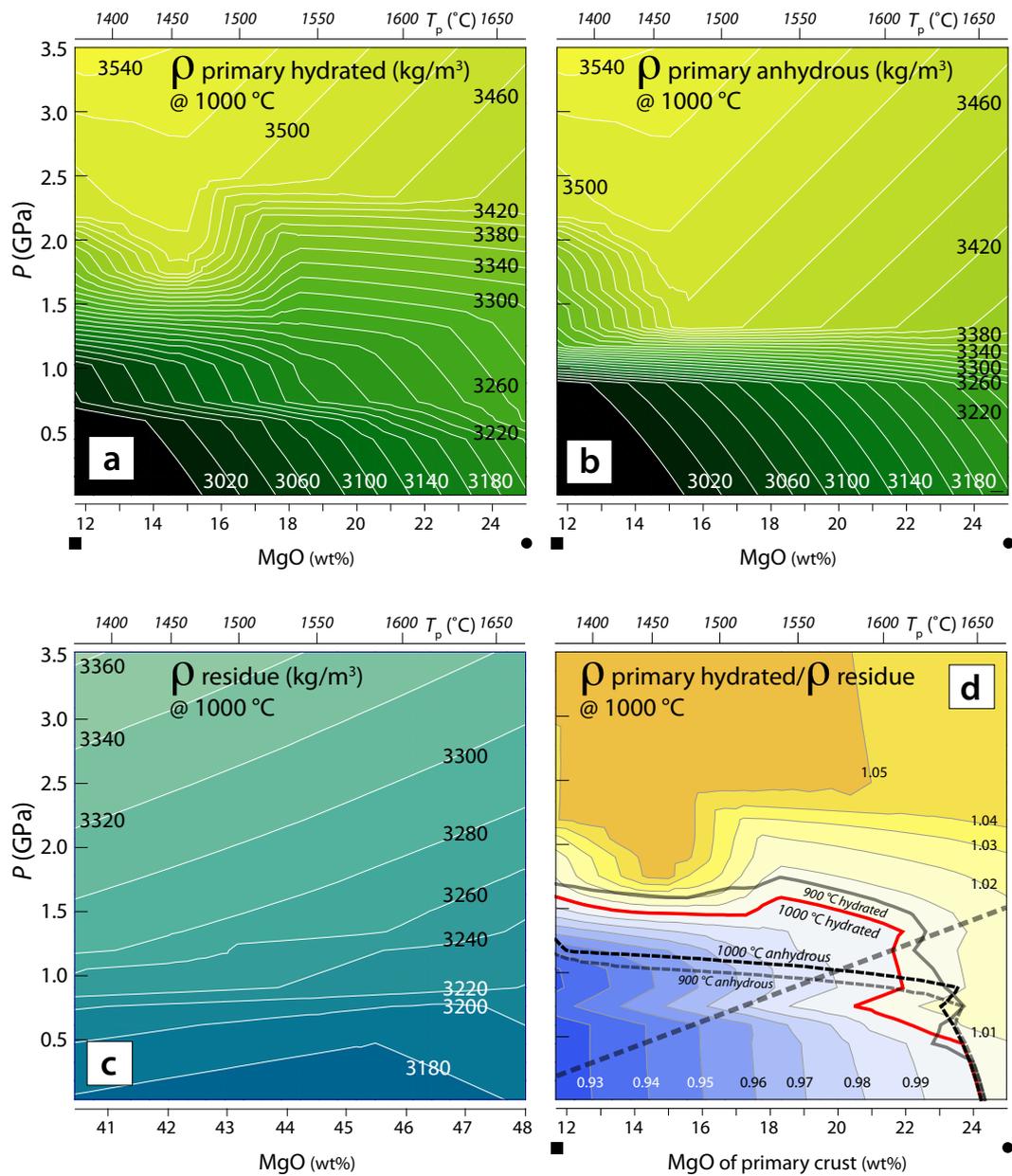


Figure 3

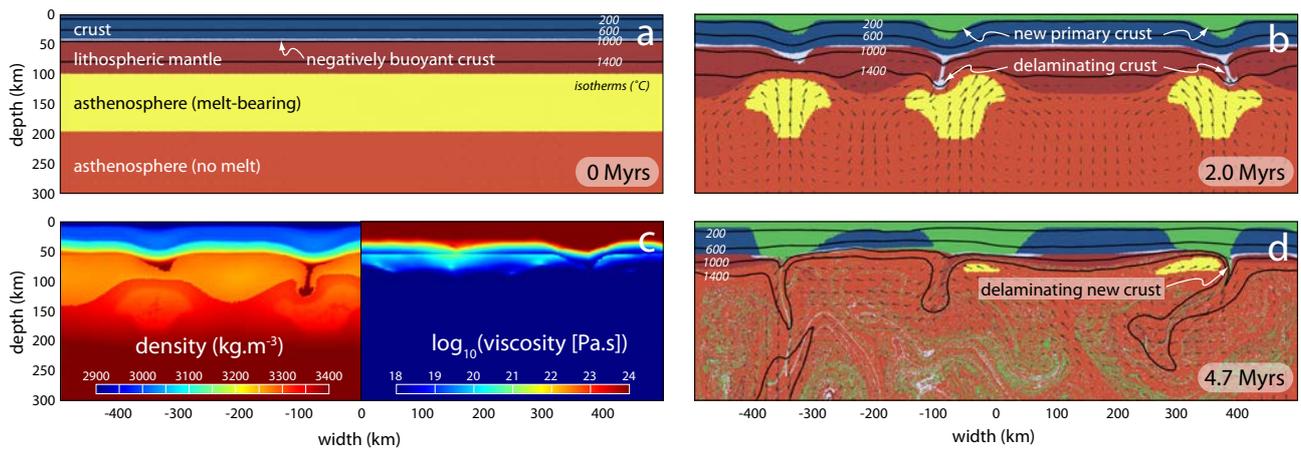


Figure 4

Delamination and recycling of Archaean crust caused by gravitational instabilities

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1. Supplementary notes for Fig. 1a

The data used in Fig. 1a are taken from Herzberg *et al.* (2010) with the following corrections to the ages: Coonterunah Group, Pilbara craton, 3.515 Ga (Green *et al.*, 2000, p. 71); Sulfur Springs Group, Pilbara craton, 3.245 Ga (mean of maximum and minimum age constraints in Smithies *et al.*, 2005, Fig. 2); Warrawoona Group Pilbara craton, 3.459 Ga (mean of maximum and minimum age constraints in Kato & Nakamura, 2003, p. 193); Ivisaartoq, Greenland, 3.018 Ga (mean of maximum and minimum age constraints in Polat *et al.*, 2008, p. 297); Iringora structure, north Karelian greenstone belt, Finland, 2.282 Ga (Shchipansky *et al.*, 2004, p.448); Kushtagi-Hungund greenstone belt, India, 2.707 Ga (Balakrishnan *et al.*, 1999, p. 74); Phulad ophiolite, Delhi Supergroup, Rajasthan, India, 1.4 Ga (minimum age, Volpe & Macdougall, 1990, p. 186); Yiyang, South China Block, 0.823 Ga (Wang *et al.*, 2007, p. 1104); Chukotat Group, Cape Smith belt, Canada (mean of maximum and minimum age constraints in St. Onge *et al.*, 1999, Fig. 3); Nathdwara, Aravalli Supergroup, India, 2.3 Ga (Ahmad *et al.*, 2008, p. 155); Vermillion greenstone belt, Minnesota, USA, 2.7 Ga (Card, 1990).

2. Phase equilibria modelling

The phase equilibria modelling uses primary melt compositions calculated by inversion of the major element chemistry of non-arc basalts of varying age as a proxy for the changing composition of the primary crust and by mass balance its complementary residue with time

(Herzberg *et al.*, 2010). The results are presented as isothermal (temperature = 1000 or 900 °C) pressure (P) – composition (X) pseudosections calculated in the Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂–O₂ (NCKFMASHTO) chemical system for the primary melt compositions and the CaO–FeO–MgO–Al₂O₃–SiO₂–O₂ (CFMASO) chemical system for the complementary residue compositions. Calculation were performed using *Perple_X* (Connolly, 2005) and the internally consistent thermodynamic dataset of Holland & Powell (1998) with the following activity–composition models for solid-solution phases: garnet (White *et al.* 2007), clinopyroxene and amphibole (Diener & Powell, 2012), orthopyroxene and spinel–magnetite (White *et al.*, 2002), plagioclase (Holland & Powell, 2003), and ilmenite–hematite (White *et al.*, 2000). FMS olivine incorporates a symmetric Fe–Mg interaction parameter ($W_{fo,fa}$) of 8 kJ.

The primary crust is modelled either as fully hydrated, requiring < or <<1.5 wt% H₂O, in which H₂O is contained within hornblende with ~2 wt% H₂O, or as anhydrous, whereas the residues are modelled as anhydrous. The hydrated and anhydrous end-member cases allow consideration of variable hydration of the primary crust. The modelled compositional range for primary crust is a binary mixture between the primary melt composition for MORB ($X = 0$) and a fictive end member composition (MgO of 25 wt%; $X = 1$) as a proxy for primary crust generated at extreme T_p (Herzberg *et al.*, 2010; Fig. 1b). The composition of these end members in terms of the components SiO₂, TiO₂, Al₂O₃, FeO, MgO, CaO, Na₂O, O₂ expressed as mol.% and normalised to 100% is: 49.49, 0.66, 10.11, 6.69, 17.72, 12.72, 2.28, 0.33 ($X = 0$) and 42.96, 0.60, 2.61, 10.09, 34.97, 7.53, 0.99, 0.25 ($X = 1$). The end member compositions of the binary compositional range for the complementary residue in terms of the components SiO₂, Al₂O₃, FeO, MgO, CaO, O₂ expressed as mol.% and normalised to 100% is: 38.25, 1.61, 5.80, 51.66, 2.53, 0.15 ($X = 0$) and 35.76, 1.28, 4.07, 58.52, 0.27, 0.10 ($X = 1$). All calculations use a constant Fe³⁺/(Fe³⁺ + Fe²⁺) ratio of 0.1 (Berry *et al.*, 2008). The model calculation and phase diagrams include the following phases (and abbreviations): olivine (ol), garnet (g), clinopyroxene (cpx), orthopyroxene (opx), hornblende (hb), plagioclase (pl), quartz (q), ilmenite (ilm), rutile (ru), spinel (sp), magnetite (mt), hematite (hem) and H₂O as appropriate.

Figure 2 shows the hydrated (Fig. 2a) and anhydrous (Fig. 2b) metamorphic mineral assemblage stability fields for the range of primary melt compositions (from $X = 0$ to $X = 1$) and their complementary residues, Fig. 2c, for pressure up to 3.5 GPa at 1000 °C; also shown in Fig. 2a is the reduced stability field of amphibolite at 900 °C. These simplified diagrams are based on the complete phase diagrams shown in Figs S1a–S6a and the rock names used in Fig. 2 are based on the modes shown in Figs S1b–S6b, which are located at the end of this section. In Figs 2a, b, the green field of amphibolite and garnet amphibolite is defined by assemblages containing both hornblende and plagioclase. Ultramafic rocks shown in blue are defined by assemblages lacking plagioclase but containing diopsidic clinopyroxene ($X_{jd} < 0.20$), within which metahornblendites contain >50 vol.% hornblende. In the purple eclogite field garnet and omphacitic clinopyroxene ($X_{jd} > 0.25$) coexist with rutile and quartz/coesite. ‘Eclogitic’ rocks in pale purple are modally dominated by garnet and clinopyroxene of ‘intermediate’ composition ($0.20 < X_{jd} < 0.25$). For primary crust at 1000 °C, clinopyroxene abundance varies widely from <15 to >50 vol.%, but is predicted in all assemblages; it is diopside except at high P (>1.8 GPa) and low MgO (<15 wt%), where it is omphacite. Hornblende contents are up to 60 vol.%, with maximum abundances at around 1.0 GPa and 18 wt% MgO. Garnet is stable above 1.0 GPa at low MgO and above 1.5 GPa at the highest MgO contents modelled. Low MgO (< 15 wt%) ‘MORB-like’ compositions are amphibolites at ~0.6–1.0 GPa or garnet-amphibolites at ~1.0–1.7 GPa in which rutile is stable at $P > 1.3$

GPa, and anhydrous kyanite-bearing quartz/coesite eclogites at $P > 1.7$ GPa. High MgO compositions (>18 wt%) contain olivine, orthopyroxene and ilmenite at all modelled pressures; they are metahornblendites or metapyroxenites above ~ 0.75 – 0.65 GPa to ~ 2.3 GPa and anhydrous garnet pyroxenites at higher pressure. For intermediate compositions (15–18 wt% MgO) the upper pressure stability limit of plagioclase (and hence amphibolite) declines steeply to lower pressures with increasing MgO. Amphibolites will begin to melt at ~ 900 °C, potentially to produce TTG-like melts. Metahornblendites and hornblende-bearing metapyroxenites will begin to melt at ~ 1000 °C, potentially to produce basalt.

In Fig. 2c all compositions are peridotites with >50 vol.% olivine and 10–30 vol.% orthopyroxene; the abundance of both phases increases with increasing MgO. Garnet-harzburgite with trace amounts of clinopyroxene is predicted in all but the lowest MgO compositions at $P > 1.25$ GPa; spinel-bearing peridotites are stable at $P < 1.2$ and additionally contain plagioclase at $P < 0.8$ GPa. Lherzolite with > 5 vol.% clinopyroxene is only stable in relatively low MgO residues and at $P < 1.5$ GPa (because the degrees of melting are lower in this X–P range). Also shown are contours for the forsterite content of olivine (as $Mg\# = 100 * \text{molar MgO}/(\text{MgO} + \text{FeO})$), which increases from 91 to 94 with increasing MgO.

Figures 3a and b show the calculated densities for hydrated and anhydrous primary crust calculated at 1000 °C, whereas Fig. 3c shows the calculated densities for the complementary residues, plotted against MgO of the primary crust. The densities shown in these diagrams are based on the complete phase diagrams shown in Figs S1a, S3a and S5a. The residues have densities ranging from < 3200 kg/m³ at low P to > 3360 kg/m³ at 3.5 GPa for low MgO compositions. At fixed pressure, densities decrease with increasing MgO by up to 2% across the full compositional range modelled. Density contours for the metamorphosed primary crust are more complex reflecting the more variable mineral assemblages. For both fully hydrated and anhydrous primary crust the lowest densities occur at low P and low MgO where the calculated abundance of plagioclase is highest. At low pressures (< 1.0 GPa), densities increase towards higher MgO and, at fixed pressure, may increase by 8% over the full compositional range modelled. In fully hydrated primary crust at low MgO compositions, density increases dramatically between 1.0 and 2.0 GPa recording the transition from amphibolite to eclogite. In higher MgO compositions densities increase less dramatically with increasing pressure, recording a transition from garnet-absent to garnet-bearing metahornblendite and metapyroxenite and eventually to anhydrous garnet metapyroxenite. In anhydrous primary crust at MgO > 15 wt% density increases dramatically from ~ 1.0 to > 1.3 GPa recording the appearance of spinel and garnet at the expense of plagioclase, and at MgO < 15 wt% density continues to increase to > 2 GPa as plagioclase is eliminated over a wide range of pressure. Figures S7a and b show the calculated densities for hydrated and anhydrous primary crust calculated at 900 °C, whereas Fig. S7c shows the calculated densities for the complementary residues, plotted against MgO of the primary crust. The densities shown in these diagrams are based on the complete phase diagrams shown in Figs S2a, S4a and S6a. The changes in density from 1000 to 900 °C are relatively small. Figure 3d shows the relative density of the metamorphosed fully hydrated primary crust with respect to the underlying residual mantle at 1000 °C plotted against MgO of the primary crust with the equality line for anhydrous crust at 1000 °C and those for 900 °C (based on Supplementary Fig. S7d) superimposed. The solid (hydrated) and dashed (anhydrous) lines separate stable primary crust (below) from potentially unstable primary crust (above).

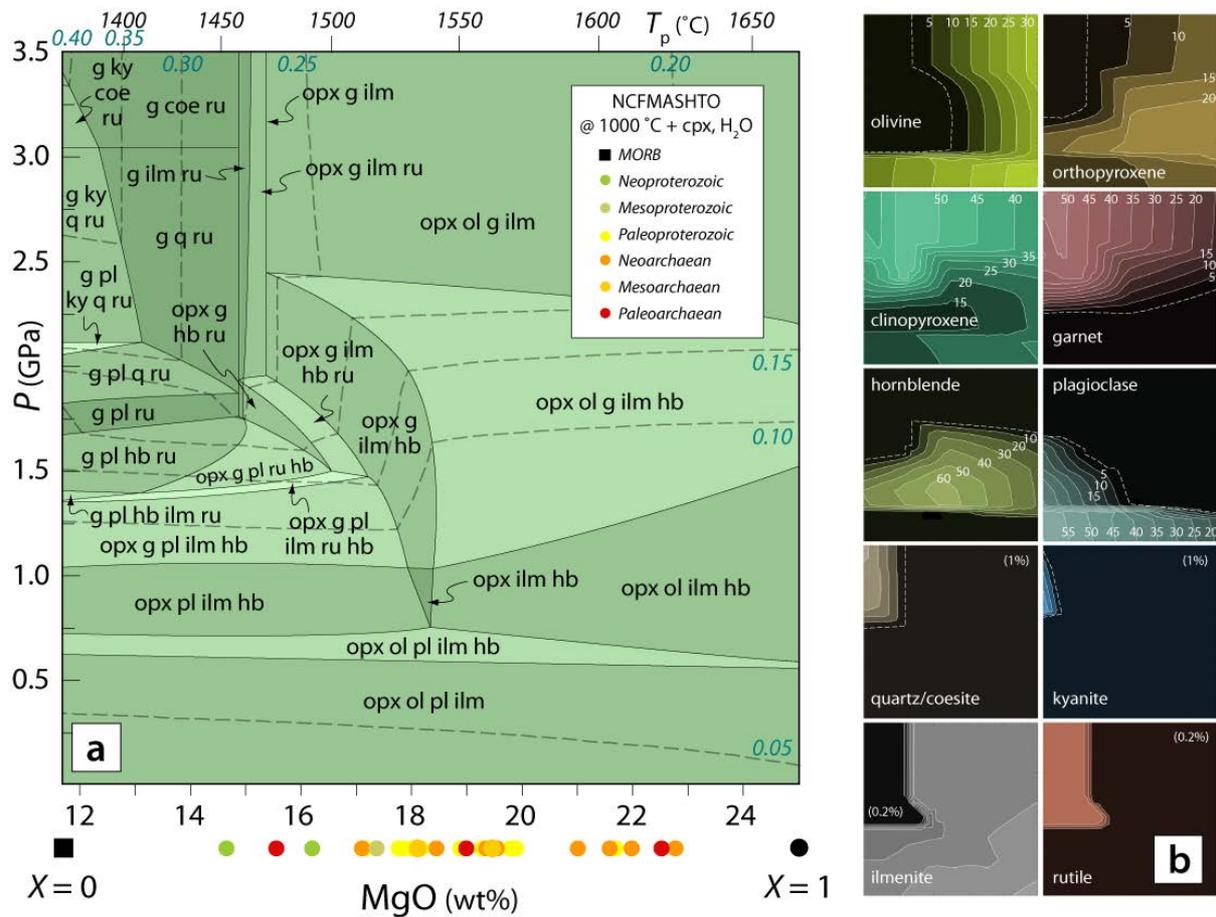


Figure S1. a, Calculated isothermal ($T = 1000\text{ °C}$) P - X pseudosection in the NCFMASHTO chemical system showing the equilibrium (minimum Gibb's free energy) mineral assemblages developed within hydrated primary crust ranging in pressure from 0–3.5 GPa and in composition from the calculated primary melt composition of MORB ($X = 0$) to a fictive end member with 25 wt% MgO ($X = 1$), together with the equivalent mantle potential temperatures (T_p ; Herzberg *et al.*, 2010). The extrapolated peak pressures at 1000 °C based on the apparent thermal gradients recorded by metamorphosed Precambrian rocks are also shown (Brown, 2007). The pseudosection is contoured for the mole fraction of the jadeite component in clinopyroxene (dashed lines). All fields are H_2O -saturated and all contain clinopyroxene in addition to those minerals listed on the diagram. Abbreviations for modelled phases are as follows: olivine (ol), garnet (g), clinopyroxene (cpx), orthopyroxene (opx), hornblende (hb), plagioclase (pl), quartz (q), ilmenite (ilm), rutile (ru), spinel (sp), magnetite (mt), hematite (hem) and H_2O . The depth of shading of the assemblage fields reflects increasing variance; the palest fields have a variance of 3. All phase diagrams were calculated using the software *Perple_X* (Connolly, 2005). Calculations use the ds55 internally consistent thermodynamic dataset (Holland & Powell, 1998) and the most recently calibrated activity–composition models (Diener & Powell, 2012). **b**, Calculated modal abundance (in vol.%) of phases within the modelled primary crust. The pressures and binary compositional range are the same as in a. The intervals between vol.% contours for the minor phases quartz/coesite, kyanite, ilmenite and rutile are indicated.

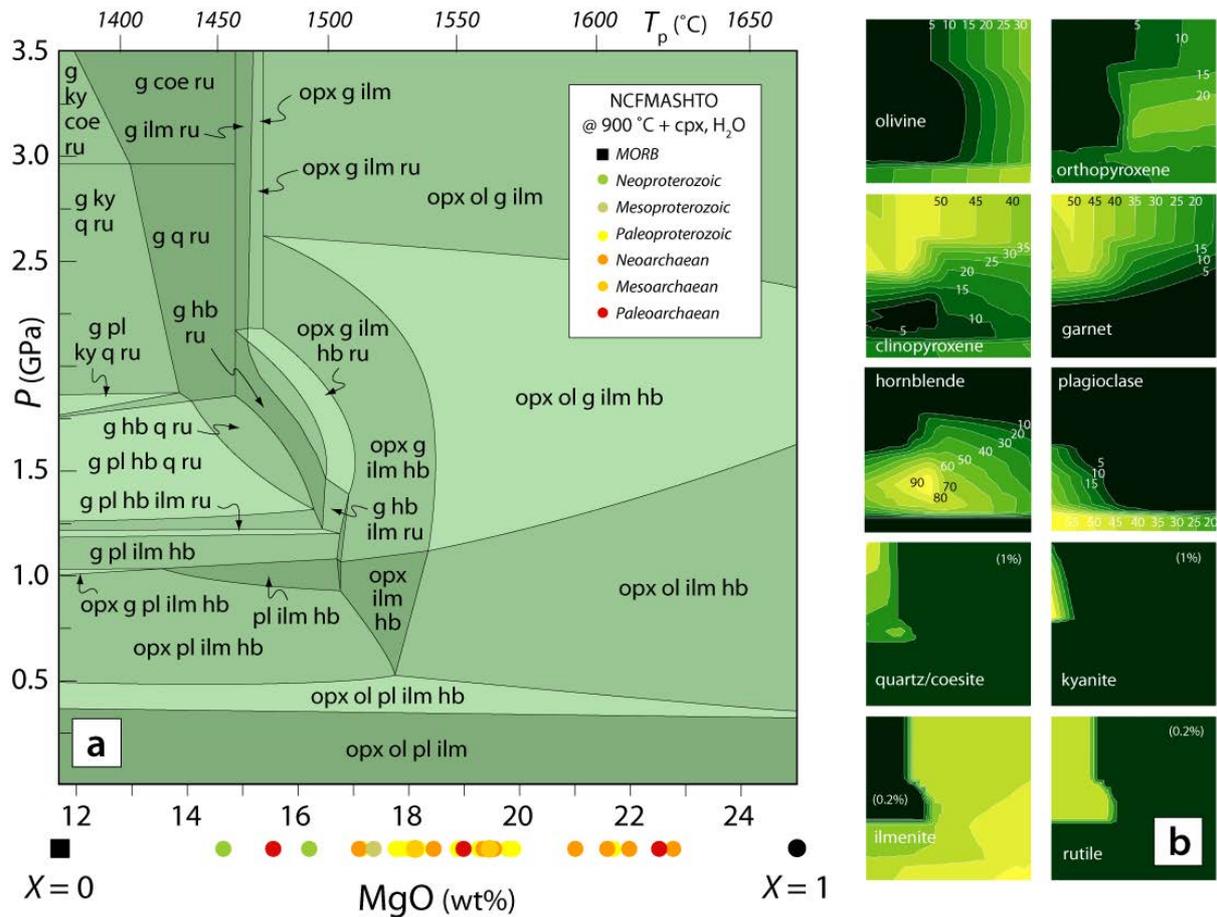


Figure S2. a, Calculated isothermal ($T = 900\text{ °C}$) P - X pseudosection in the NCFMASHTO chemical system showing the equilibrium (minimum Gibb's free energy) mineral assemblages developed within hydrated primary crust ranging in pressure from 0–3.5 GPa and in composition from the calculated primary melt composition of MORB ($X = 0$) to a fictive end member with 25 wt% MgO ($X = 1$), together with the equivalent mantle potential temperatures (T_p ; Herzberg *et al.*, 2010). The extrapolated peak pressures at 1000 °C based on the apparent thermal gradients recorded by metamorphosed Precambrian rocks are also shown (Brown, 2007). All fields are H_2O -saturated and all contain clinopyroxene in addition to those minerals listed on the diagram. Abbreviations for modelled phases are as follows: olivine (ol), garnet (g), clinopyroxene (cpx), orthopyroxene (opx), hornblende (hb), plagioclase (pl), quartz (q), ilmenite (ilm), rutile (ru), spinel (sp), magnetite (mt), hematite (hem) and H_2O . The depth of shading of the assemblage fields reflects increasing variance; the palest fields have a variance of 3. All phase diagrams were calculated using the software *Perple_X* (Connolly, 2005). Calculations use the ds55 internally consistent thermodynamic dataset (Holland & Powell, 1998) and the most recently calibrated activity–composition models (Diener & Powell, 2012). **b**, Calculated modal abundance (in vol.%) of phases within the modelled primary crust. The pressures and binary compositional range are the same as in a. The intervals between vol.% contours for the minor phases quartz/coesite, kyanite, ilmenite and rutile are indicated.

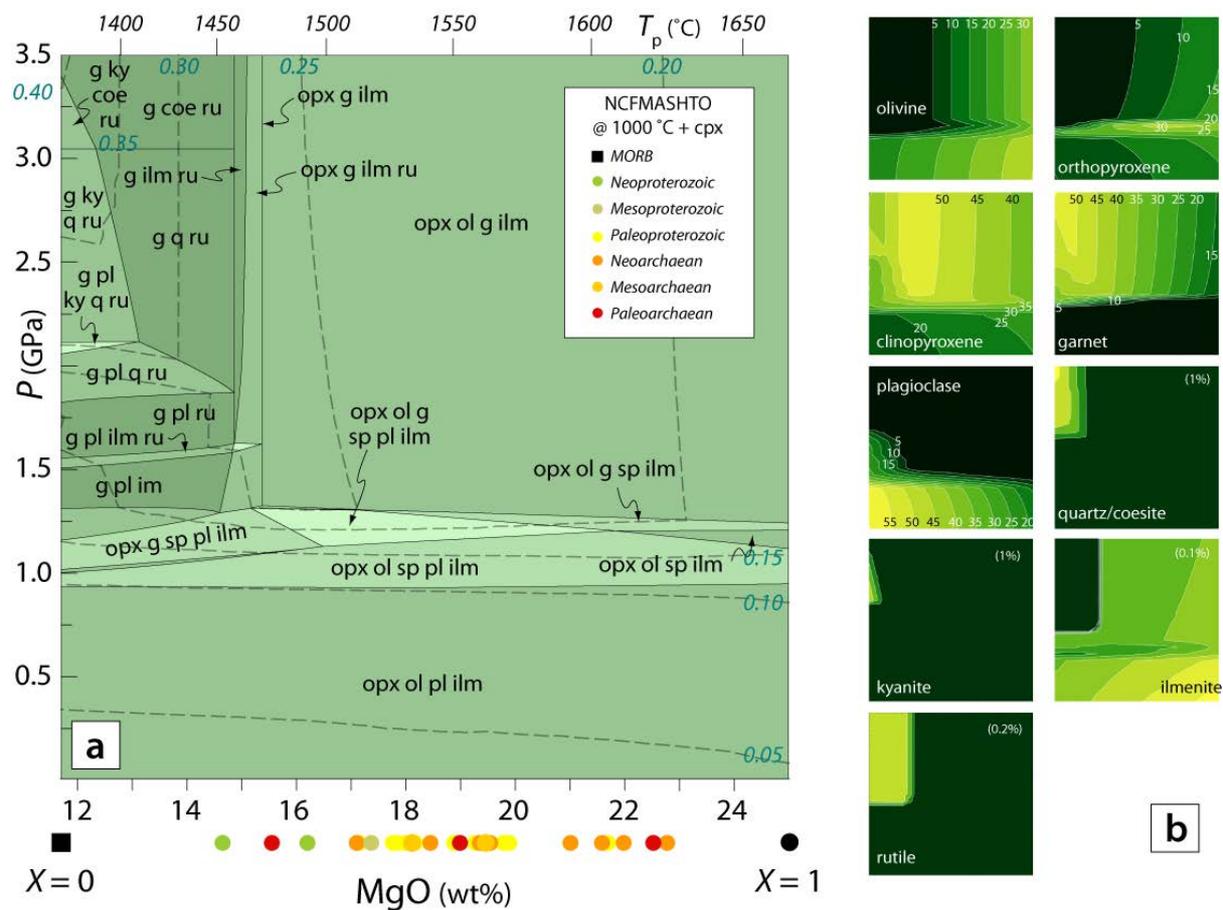


Figure S3. a, Calculated isothermal ($T = 1000\text{ °C}$) P - X pseudosection in the NCFMASHTO chemical system showing the equilibrium (minimum Gibbs free energy) mineral assemblages developed within anhydrous primary crust ranging in pressure from 0–3.5 GPa and in composition from the calculated primary melt composition of MORB ($X = 0$) to a fictive end member with 25 wt% MgO ($X = 1$), together with the equivalent mantle potential temperatures (T_p ; Herzberg *et al.*, 2010). The extrapolated peak pressures at 1000 °C based on the apparent thermal gradients recorded by metamorphosed Precambrian rocks are also shown (Brown, 2007). All fields contain clinopyroxene in addition to those minerals listed on the diagram. Abbreviations for modelled phases are as follows: olivine (ol), garnet (g), clinopyroxene (cpx), orthopyroxene (opx), plagioclase (pl), quartz (q), ilmenite (ilm), rutile (ru), spinel (sp), magnetite (mt), and hematite (hem). The depth of shading of the assemblage fields reflects increasing variance; the palest fields have a variance of 3. All phase diagrams were calculated using the software *Perple_X* (Connolly, 2005). Calculations use the *ds55* internally consistent thermodynamic dataset (Holland & Powell, 1998) and the most recently calibrated activity–composition models (Diener & Powell, 2012). **b**, Calculated modal abundance (in vol.%) of phases within the modelled primary crust. The pressures and binary compositional range are the same as in a. The intervals between vol.% contours for the minor phases quartz/coesite, kyanite, ilmenite and rutile are indicated.

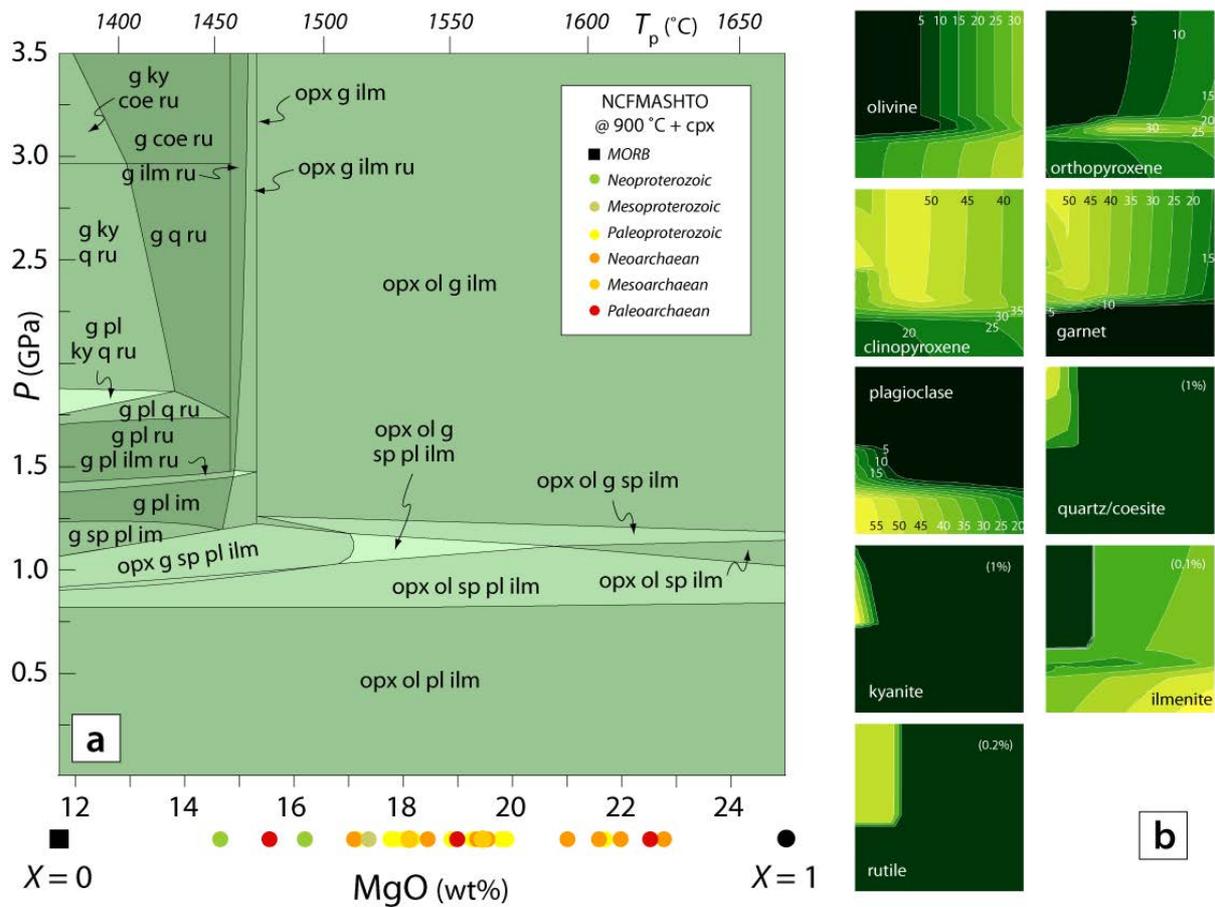


Figure S4. a, Calculated isothermal ($T = 900\text{ °C}$) P - X pseudosection in the NCFMASHTO chemical system showing the equilibrium (minimum Gibb's free energy) mineral assemblages developed within hydrated primary crust ranging in pressure from 0–3.5 GPa and in composition from the calculated primary melt composition of MORB ($X = 0$) to a fictive end member with 25 wt% MgO ($X = 1$), together with the equivalent mantle potential temperatures (T_p ; Herzberg *et al.*, 2010). The extrapolated peak pressures at 1000 °C based on the apparent thermal gradients recorded by metamorphosed Precambrian rocks are also shown (Brown, 2007). All fields contain clinopyroxene in addition to those minerals listed on the diagram. Abbreviations for modelled phases are as follows: olivine (ol), garnet (g), clinopyroxene (cpx), orthopyroxene (opx), plagioclase (pl), quartz (q), ilmenite (ilm), rutile (ru), spinel (sp), magnetite (mt), and hematite (hem). The depth of shading of the assemblage fields reflects increasing variance; the palest fields have a variance of 3. All phase diagrams were calculated using the software *Perple_X* (Connolly, 2005). Calculations use the ds55 internally consistent thermodynamic dataset (Holland & Powell, 1998) and the most recently calibrated activity–composition models (Diener & Powell, 2012). **b**, Calculated modal abundance (in vol.%) of phases within the modelled primary crust. The pressures and binary compositional range are the same as in a. The intervals between vol.% contours for the minor phases quartz/coesite, kyanite, ilmenite and rutile are indicated.

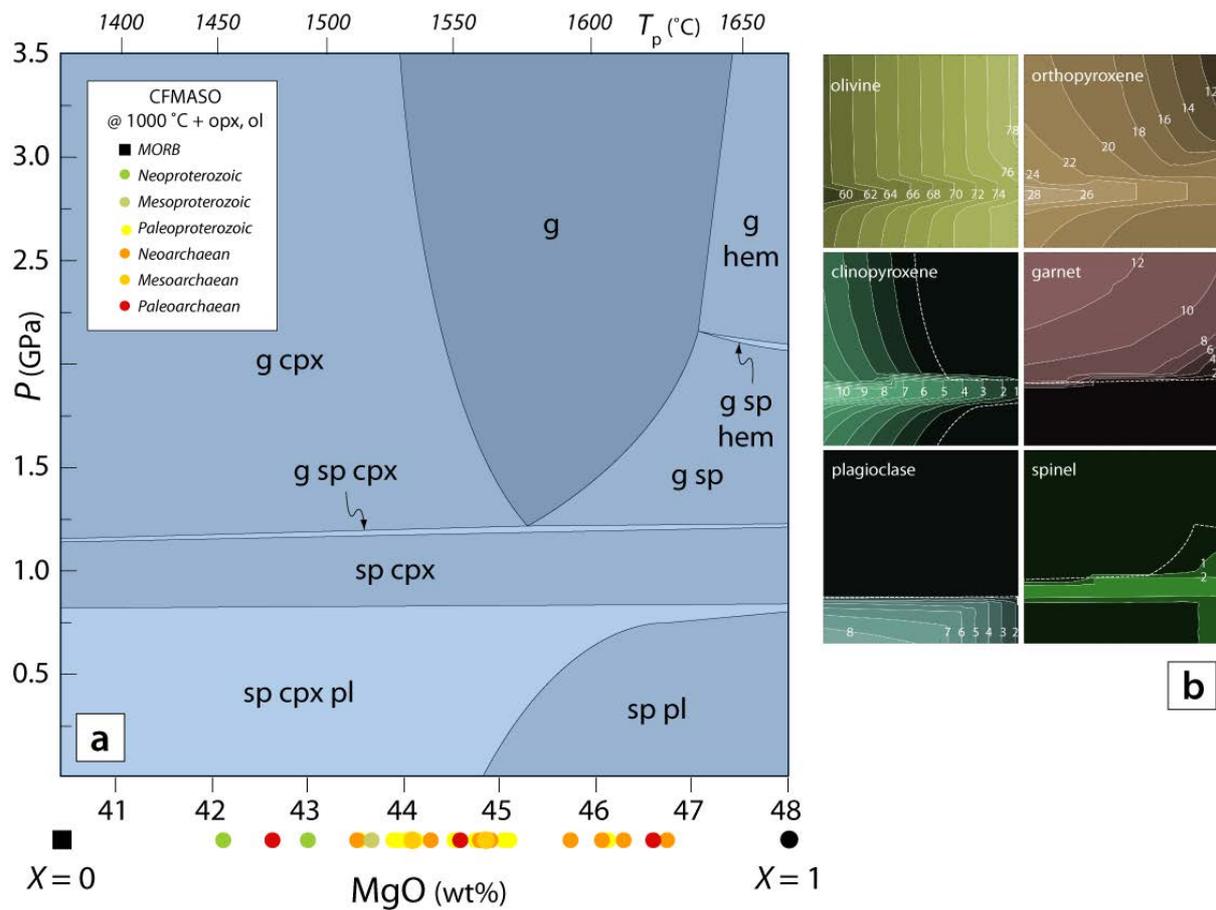


Figure S5. **a**, Calculated isothermal ($T = 1000\text{ °C}$) P - X pseudosection in the CFMASO chemical system showing the equilibrium (minimum Gibb's free energy) mineral assemblages developed within mantle residues ranging in pressure from 0–3.5 GPa and in composition from the calculated residue composition of MORB ($X = 0$) to that of a fictive end member with 48 wt% MgO ($X = 1$), together with the equivalent mantle potential temperatures (T_p ; Herzberg *et al.*, 2010). The extrapolated peak pressures at 1000 °C based on the apparent thermal gradients recorded by metamorphosed Precambrian rocks are also shown (Brown, 2007). All fields contain orthopyroxene and olivine in addition to those minerals listed on the diagram. Abbreviations for modelled phases are as follows: garnet (g), clinopyroxene (cpx), plagioclase (pl), and hematite (hem). The depth of shading of the assemblage fields reflects increasing variance; the palest fields have a variance of 4. All phase diagrams were calculated using the software *Perple_X* (Connolly, 2005). Calculations use the *ds55* internally consistent thermodynamic dataset (Holland & Powell, 1998) and the most recently calibrated activity–composition models (Diener & Powell, 2012). **b**, Calculated modal abundance (in vol.%) of phases within the modelled mantle residues. The pressures and binary compositional range are the same as in **a**.

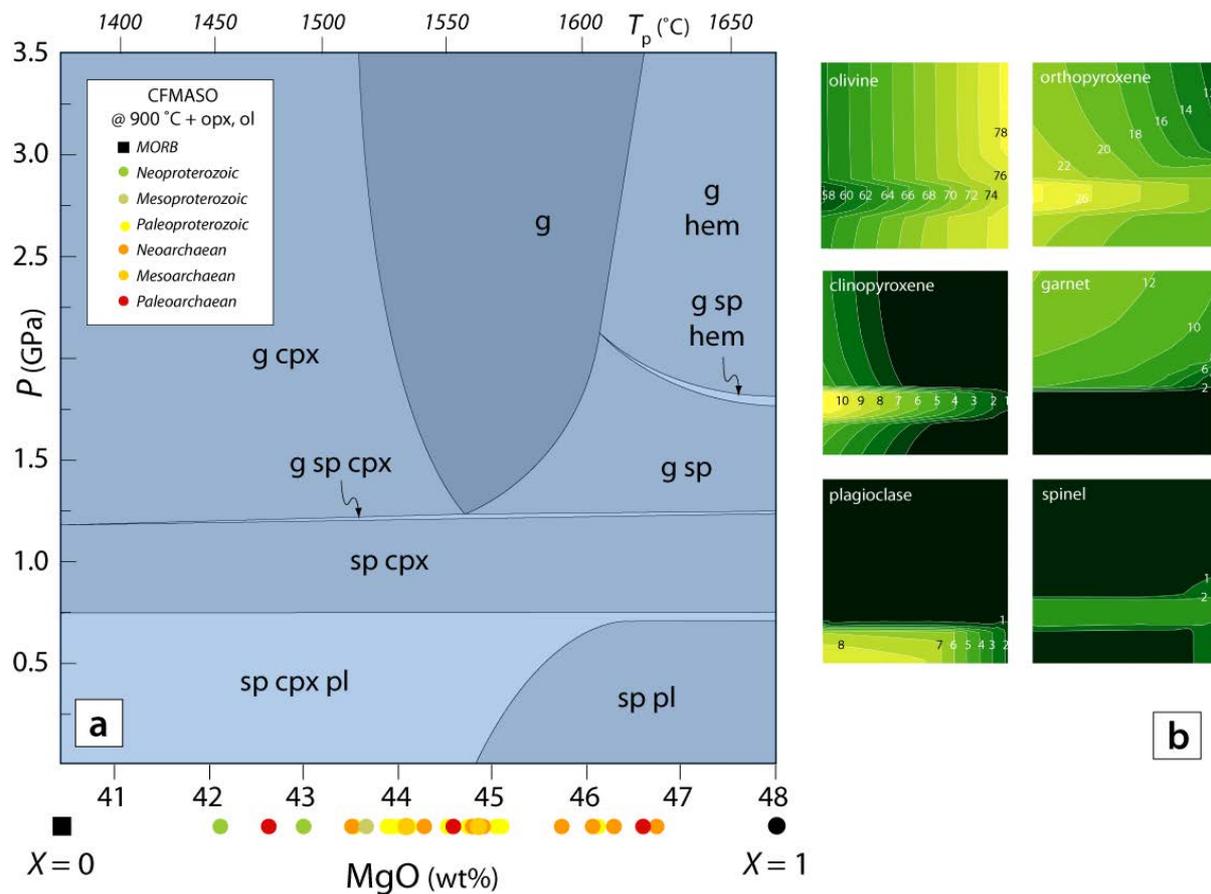


Figure S6. a, Calculated isothermal ($T = 900\text{ °C}$) P - X pseudosection in the CFMASO chemical system showing the equilibrium (minimum Gibbs free energy) mineral assemblages developed within mantle residues ranging in pressure from 0–3.5 GPa and in composition from the calculated residue composition of MORB ($X = 0$) to that of a fictive end member with 48 wt% MgO ($X = 1$), together with the equivalent mantle potential temperatures (T_p ; Herzberg *et al.*, 2010). The extrapolated peak pressures at 1000 °C based on the apparent thermal gradients recorded by metamorphosed Precambrian rocks are also shown (Brown, 2007). All fields contain orthopyroxene and olivine in addition to those minerals listed on the diagram. Abbreviations for modelled phases are as follows: garnet (g), clinopyroxene (cpx), plagioclase (pl), and hematite (hem). The depth of shading of the assemblage fields reflects increasing variance; the palest fields have a variance of 4. All phase diagrams were calculated using the software *Perple_X* (Connolly, 2005). Calculations use the ds55 internally consistent thermodynamic dataset (Holland & Powell, 1998) and the most recently calibrated activity–composition models (Diener & Powell, 2012). **b,** Calculated modal abundance (in vol.%) of phases within the modelled mantle residues. The pressures and binary compositional range are the same as in a.

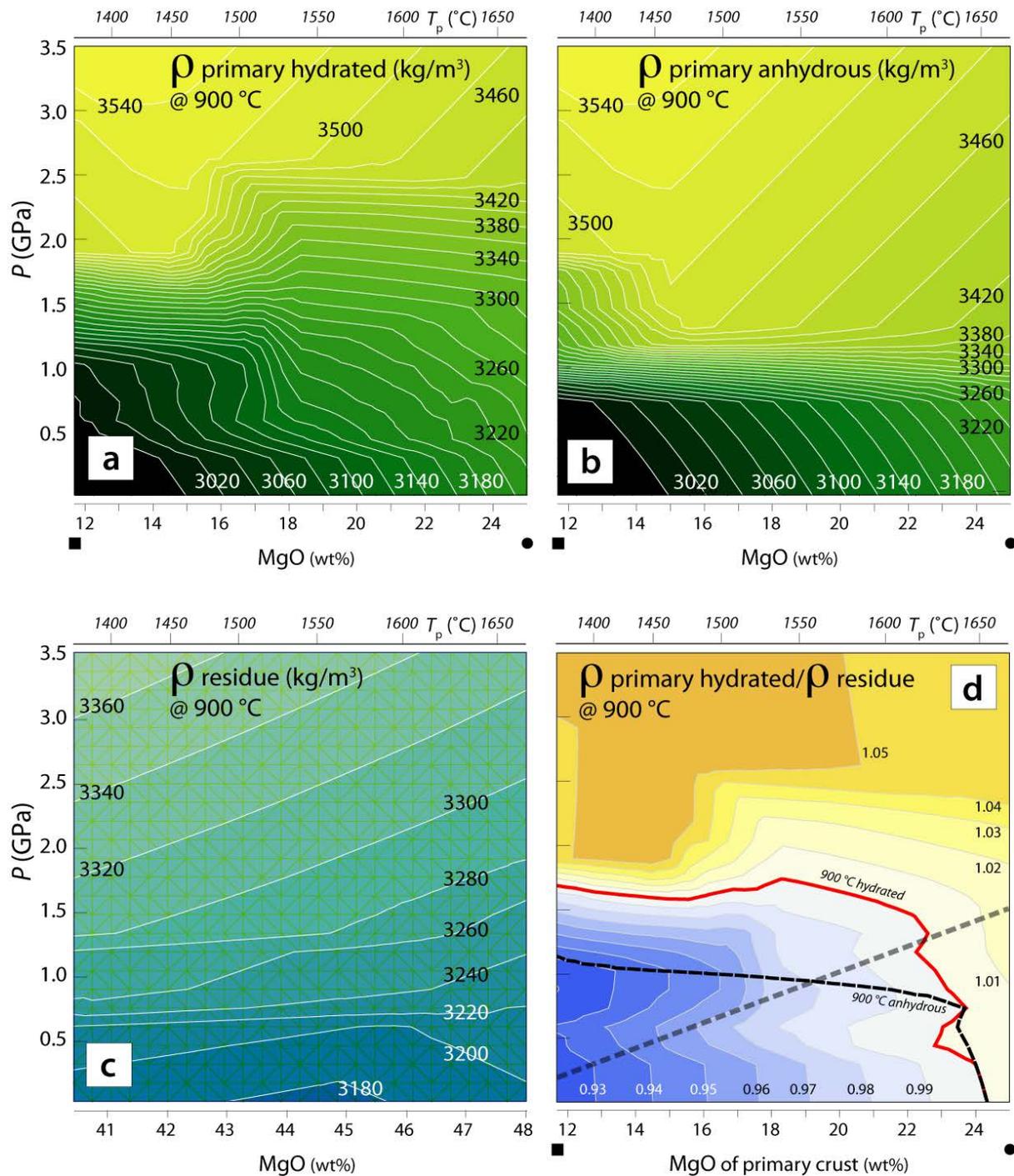


Figure S7. Density of primary crust and complementary residues. Densities calculated at 900 °C plotted against MgO content for **a**, hydrated primary crust, **b**, anhydrous primary crust and **c**, the complementary residues. **d**, Shows the relative density of hydrated crust compared to the underlying residual mantle at 900 °C against MgO of the primary crust with the equality line for anhydrous crust at 900 °C superimposed. The solid (hydrated) and dashed (anhydrous) lines separate stable primary crust (below) from potentially unstable primary crust (above).

3. Geodynamic numerical modelling

3.1. Background

The numerical code employed for the geodynamic simulations has been described in detail elsewhere (Kaus, 2010; Thielmann & Kaus, 2012; Schmeling *et al.*, 2008). For completeness, a brief summary is provided here together with a description of several new features added to the code for the present study, namely melt extraction and evolution.

3.1.1. Governing equations and numerical approach

The incompressible Stokes system of equations are given by

$$\frac{\partial v_i}{\partial x_i} = 0$$

$$-\frac{\partial P}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} = \rho g_i$$

$$\rho c_p \left(\frac{\partial T}{\partial t} + v_i \frac{\partial T}{\partial x_i} \right) = \frac{\partial}{\partial x_i} \left(k \frac{\partial T}{\partial x_i} \right) + \tau_{ij} \dot{\epsilon}_{ij} - \alpha T \rho g v_2$$

Where v_i indicates velocity, P pressure, τ_{ij} deviatoric stresses, g_i gravitational acceleration, ρ density, c_p heat capacity, T temperature, k thermal conductivity, ϵ_{ij} strain rate, and α thermal expansivity, and the Einstein notation is used and both shear heating and adiabatic (de)compression are taken into account in the energy equation.

Density in the models is a function of rock type/composition (X), melt fraction (M), pressure (P) and temperature (T):

$$\rho = f(X, M, P, T)$$

A viscoplastic constitutive relationship is used, as follows

$$\tau_{ij} = 2\eta_{eff} \dot{\epsilon}_{ij}$$

$$\eta_{eff} = \min(\eta_{diff}, \eta_{disl}, \eta_{yield})$$

in which the dominant rheology is diffusion creep, dislocation creep or plastic yielding, as given by

$$\eta_{diff} = \eta_{0,diff} \exp \left(Q_{diff} \left(\frac{1}{T} - \frac{1}{873} \right) \right)$$

$$\eta_{disl} = \eta_{0,disl} \left(\frac{\dot{\epsilon}_{II}}{10^{-15}} \right)^{\frac{1}{n}-1} \exp \left(Q_{disl} \left(\frac{1}{T} - \frac{1}{873} \right) \right)$$

$$\eta_{yield} = \frac{\sigma_{yield}}{2\dot{\epsilon}_{II}}$$

where σ_{yield} is the yield stress, ϵ_{II} the second invariant of the strain rate tensor, η_0 the effective viscosity at 873 K and 10^{-15} s^{-1} , n the power law exponent and Q the temperature-dependency of viscosity. For mantle rocks we take diffusion and dislocation creep into account for both “wet” and “dry” end-members whereas the crust is assumed to have a diabase dislocation creep rheology. Density depends on the rock-type, melt fraction and the pressure–temperature conditions, as described in more detail below.

The governing equations are solved with the 2D finite element code MILAMIN_VEP, and we employ LBB stable Q₂P₁ elements for the Stokes equations and quadratic elements for temperature. The code is written in MATLAB, but nevertheless is made relatively efficient by employing the techniques described in detail in Dabrowski *et al.* (2008). Tracers are used to track material properties and the code is used in an ALE mode with regular remeshing to ensure that elements do not become too distorted. This code has been benchmarked for a large number of test problems including convection, Rayleigh-Taylor instabilities, thermal diffusion, shear-heating, and multilayer folding, as well as versus various analytical solutions (e.g. SolCx or weak/strong inclusion setups), as described in a number of publications (Gerya & Yuen, 2007; May & Moresi, 2008; Schmeling *et al.*, 2008; Cramer & Kaus, 2010; Yamato *et al.*, 2011; Cramer *et al.*, 2012; Thielmann & Kaus, 2012).

3.1.2. Phase transitions and partial melting

The variations of crustal density as a function of P and T is computed with *Perple_X*, as described in the main text. We discriminate between initial crust, which has a composition with 23 wt% MgO (at $X = 0.839$ on the binary mixture for the NCFMASHTO modelled primary crust given in the Methods) and “new crust”, which has a composition with 13 wt% MgO (at $X = 0.104$ on the binary mixture for the NCFMASHTO modelled primary crust given in the Methods).

We take into account two types of mantle: lithospheric mantle, which is the residual mantle leftover after forming the initial crust (i.e. it has a composition at $X = 0.839$ on the binary mixture for the CFMASO modelled residua given in the Methods), and underlying asthenospheric mantle, which is more fertile than the lithospheric mantle and has the CFMASO model composition of KR-4003 (Herzberg *et al.* 2010). For both mantle lithologies, initially the density is computed as a function of P and T without melting using *Perple_X*. In a next step, *pMELTS* (Ghiorso *et al.*, 2002) is used to estimate the solidus and liquidus temperatures as well as melt fractions for both (depleted) lithospheric mantle and asthenosphere. Although this gives reliable estimates, *pMELTS* is only well-calibrated for pressures between 1 and 3 GPa and melt fractions smaller than 0.3, whereas the models in this study require information from 0–10 GPa or higher. Therefore, for the fertile asthenospheric mantle the parameterization of Katz *et al.* (2003) is used, but adapted for the appropriate Cpx modal fractions for the composition used. For the more depleted mantle lithosphere, the parameterization of Katz *et al.* (2003) is not well calibrated (as too few

experiments exist), and therefore we rely solely on pMELTS data that have been extrapolated to higher and lower pressures, after adapting the solidus and liquidus parameterization for Katz *et al.* (2003) to fit the pMELTS data in the validity range. In common with most simulations, the lithospheric mantle does not melt extensively; this does not have a large impact on the results of this study, but future experimental work is nevertheless required to better constrain the parameterization.

Melting of crustal rocks is not taken into account in the geodynamic modelling, because a thermodynamic model is not currently available to reliably predict the P - T relations of the solidus and liquidus for these crustal compositions, although, if fully hydrated, one might expect crustal rocks of the compositions modelled in this study to melt above temperatures of 1000 °C. The absence of melting in the crust is a crude first-order approximation. However, we do not expect the inclusion of crustal melting to significantly alter model outcomes because this will produce both melt and residue, and the latter will be even denser than the densities already used, so the drips will likely continue descending into the mantle.

3.1.3. Melting and melt extraction

In reality, melt will be extracted by a variety of mechanisms, such as diking or compaction-driven flow. As these processes mostly happen on a timescale that is significantly smaller than the typical time step employed in the model, we assume that melt in the mantle is extracted instantaneously and emplaced on top of the crust. In doing this the possibility that melt might be trapped in the crust during ascent is ignored, which is justified as both mechanisms will thicken the crust.

The melt extraction algorithm largely follows the approach of Sizova *et al.* (2010), and is done in several steps, as follows.

1) The theoretical melt fraction M_0 of a given rock is computed, which depends on the composition of the rock (asthenosphere or mantle lithosphere), the P - T condition and the melting diagram, which is a lookup table as a function of P, T . In general, melt fractions do not vary linearly between solidus and liquidus, but instead change nonlinearly and in a stepwise manner, for example once Cpx in the rock is consumed (Katz *et al.*, 2003).

2) It is assumed that melt will be extracted once the melt fraction exceeds a critical extraction melt fraction ($M_{extract}$), which is set as 0.05 in this work. After a melt-extraction event, a small amount of non-extractable melt (M_{min}) will remain in the rock, which is here taken as 0.02.

3) The amount of melt that has been extracted is tracked on markers. The amount of melt currently present at each marker takes into account melt that has been extracted from that marker during earlier melt extraction events, and is computed as

$$M = M_0 - \sum_n M_{ext}$$

where $\sum_n M_{ext}$ is the total amount of melt extracted in the previous n melt extraction events.

4) If $M > M_{extract}$, the extractable melt fraction $M_{ext} = M - M_{min}$ moves upwards to create new crust and the remaining melt fraction in the mantle (M) is updated. Simultaneously, the remaining rock will have a reduced volume $1 - M_{ext}$, which is compensated by compaction in the vertical direction. This is computed by interpolating the irregular marker distribution on a regular grid on which the vertical motion due to compaction as well as the thickness of the newly formed crust is calculated for every horizontal grid cell. New markers are randomly inserted in the newly generated crust and existing markers are moved downwards accordingly. The newly formed crust, which typically has a thickness of a few 100 meters, will initially be hot (and have mantle temperature) but will cool rapidly on a timescale smaller than the typical time step used (a few 1000 years). For simplicity, therefore, we assume the temperature of the newly formed crust to be 700 °C. Since the characteristic diffusion time for the new crust is typically smaller than the time step, and the new magma is assumed to be emplaced at the surface where gradients in temperature are large, the precise temperature of the new crust does not have a major impact on the results. This simplification was verified by performing additional calculations in which the emplacement temperature was varied from 20 to 900 °C; these experiments yielded similar results. A more sophisticated approach, in which one would model every magmatic pulse with, for example, a 1-D temperature profile, is expected to have little effect on the key results and was not employed here.

5) If more than a certain amount of melt is extracted from a tracer (M_{max}), the rock is assumed to be depleted and no longer produces melt.

6) The effective density of the remaining partially molten mantle rock is computed according to $\rho = M \rho_{melt} + (1 - M) \rho_{phasediagram}$ where ρ_{melt} is the density of the melt, which is assumed to have a constant value of 2800 kg/m³ for simplicity, although we note this is in good agreement with estimates from pMELTS calculations, and $\rho_{phasediagram}$ is the density taken from the phase diagram without partial melting.

The effective viscosity of a partially molten region depends on the melt fraction in a nonlinear fashion (e.g., Bittner, & Schmeling, 1995; Deubelbeiss *et al.*, 2010). As the mafic melts produced in our model are likely to have had a very low viscosity, similar to that of present-day basaltic melts (on the order of 1 Pa s), and the melting typically occurs in regions of the asthenosphere which already have a low solid viscosity, the effective viscosity of the partially molten rock (with no more than 5 vol.% melt in the experiments) is likely to be lower than cut-off viscosity employed. Therefore, a constant viscosity of 10¹⁸ Pas is assigned to regions that are partially molten.

3.1.4. Model setup

The model consists of a 2D box that is 1000 km wide and 660 km high, and has an initial crust of variable thickness, an underlying depleted mantle lithosphere up to a depth of 100 km, and an underlying fertile asthenosphere. Mechanical boundary conditions are free slip (shear stress free) on the side and top boundaries, and no-slip on the bottom boundary, which mimics a viscosity increase in the lower mantle compared to that of the upper mantle. Thermal boundary conditions are isothermal at the top and bottom (in accordance with the initial geotherm) and flux-free at the side boundaries. As the initial geotherm in the Archaean is uncertain and has evolved dynamically, we use a simplistic approach by assuming it to be initially linear in the crust, with 20 °C at the surface and 1000 °C at the Moho, and linearly increasing below until it intersects the mantle adiabat at a depth of 100 km. The mantle adiabat is computed based on T_p (which is varied in the simulations) and an increase of 0.3

K/km. The temperature condition at the bottom of the model is computed from the mantle adiabat. An overview of model parameters is given in Table S1.

The model employs a computational mesh of 257x257 nodes, which is refined towards the lithosphere, such that there is a resolution of 3.9 km (horizontal) by 2.1 km (vertical) in the lithosphere and 3.9 by 3.2 km below a depth of 330 km. Compositions are tracked on tracers of which there are initially ~0.8 million, which number increases during the simulation as new crust is formed. To initiate convection, random noise with a maximum amplitude of 5 °C is initially added on tracers that are deeper than 130 km. As in any geodynamic model with tracers, the fine scale mixing structure in the mantle that occurs in the later stages of the model is not well resolved and should not be interpreted literally (Tackley & King, 2003). The large-scale features, such as the dripping instabilities at the bottom of the lithosphere, are considered to be robust, as tests with a smaller numerical resolution and fewer particles give similar results.

Earth has a free surface, rather than a free slip upper boundary condition as in the experiments. A simulation with standard parameters and a free surface condition was performed to test the effect of this difference on the model dynamics; the result was that the surface condition had a negligible effect on the model dynamics. Surface topographies are typically less than a kilometre in amplitude.

With the chosen geotherm, the asthenosphere immediately underneath the mantle lithosphere is initially partially molten, which results in the formation of a thin layer of new crust at the onset of the simulation (with thicknesses of 2, 5 and 10 km for simulations with a 45 km thick initial crust and a mantle T_p of 1500, 1550 and 1600 °C, respectively). This is a model artefact as in reality the crust would thicken in a dynamic manner and potentially drip-off at the bottom while growing at the top. Notwithstanding, the subsequent evolution of the models is robust, with new crust preferentially forming above upwellings. Moreover, as shown below, models with thinner initial crust produce similar results. Other modelling choices are possible, such as starting with an initially partially depleted asthenosphere, but this is considered unlikely to change the overall findings of the modelling we report herein.

Parameter	Value	Units
Thermal conductivity k	3	W/m/K
Heat capacity c_p	1050	J/kg/K
Gravitational acceleration g	9.81	m/s ²
Lower cutoff viscosity η_{min}	10 ¹⁸	Pa s
Upper cutoff viscosity η_{max}	10 ²⁴	Pa s
Viscosity partially molten zone $\eta_{moltenZone}$	10 ¹⁸	Pa s
Yield stress σ_{yield}	1000	MPa
Crust¹ : dislocation creep prefactor $\eta_{0,dist}$	2.3x10 ²¹	Pas
Crust¹ : dislocation creep powerlaw exponent n	3.05	-
Crust¹ : temperature dependence of viscosity Q_{dist}	1.09x10 ⁴	K
Wet Mantle² : dislocation creep prefactor $\eta_{0,dist}$	1.08x10 ²³	Pas
Wet Mantle²: dislocation creep powerlaw exponent n	3.5	-
Wet Mantle²: temperature dependence of viscosity Q_{dist}	1.6x10 ⁴	K
Wet Mantle² : diffusion creep prefactor $\eta_{0,diff}$	1.1x10 ²⁹	Pas
Wet Mantle²: diffusion dependence of viscosity Q_{diff}	4.03x10 ⁴	K
Dry Mantle² : dislocation creep prefactor $\eta_{0,dist}$	1.07x10 ²⁴	Pas
Dry Mantle²: dislocation creep powerlaw exponent n	3.5	-
Dry Mantle²: temperature dependence of viscosity Q_{dist}	1.82x10 ⁴	K
Dry Mantle² : diffusion creep prefactor $\eta_{0,diff}$	1.8x10 ³¹	Pas
Dry Mantle²: diffusion dependence of viscosity Q_{diff}	4.5x10 ⁴	K
Minimum leftover melt fraction M_{min}	0.02	-
Extraction limit to extract melt and form new crust $M_{extract}$	0.05	-
Maximum extractable melt fraction from the rocks, before they become depleted (non-meltable) mantle rocks M_{max}	0.5	-

Table S1. Model parameters employed in this work. ¹Rheology for the crust are calculated from experiments for a dry diabase rheology as reported by Carter & Tsenn (1987), ²wet and dry mantle dislocation and diffusion creep parameters are taken from Hirth & Kohlstedt (2003), table 1. We assume C_{OH} of 1000 H/10⁶ Si for wet rheologies and a grainsize of 10 mm for diffusion creep. Density and melt fraction are taken from pre-computed phase diagrams.

3.2. Geodynamic model results

Systematic experiments have been run to test the sensitivity of the key results to changes in crustal thickness, mantle T_p , and mantle rheology (wet vs dry).

3.2.1. Effect of initial crustal thickness

Models with $T_p = 1600$ °C, but with varying crustal thickness for a wet and dry mantle rheology (Figs S8, 9), reveal that the physical mechanism is essentially independent of the assumed initial crustal thickness. For crustal recycling to occur, the crust only needs to be

thickened sufficiently to become negatively buoyant at its base at the same time as it becomes sufficiently weak through heating to drip off.

3.2.2. Effect of mantle potential temperature, Moho temperature and rheology

The effect of changing the mantle T_p for a wet mantle rheology is illustrated in Fig. S10. Higher T_p results in a thicker initial crust and in its more rapid removal. As in the case with $T_p = 1600$ °C, the new crust is ultimately recycled into the mantle where it could partially melt (a process that was not modelled). For $T_p < 1550$ °C, a tiny volume of new crust is formed at the beginning of the model runs but later during the runs partial melting no longer occurs in these experiments, and a stagnant lid mode of convection sets in (note that we did not make an attempt to model plate-like behaviour in this study, which requires the use of much smaller yield stresses).

The same analysis for a dry mantle rheology suggests that recycling of the new crust into the mantle stops at a slightly higher T_p (between 1550 and 1600 °C) in these experiments in comparison to those with the wet mantle rheology, but that nonetheless it is a very efficient mechanism above these temperatures (Fig. S11).

Experiments with the ‘standard’ setup (Fig. 4, main paper) but with different initial Moho temperature yield essentially identical results as the standard model, suggesting that the initial thermal state of the crust is a second-order effect in the experiments (Fig. S12). However, it should be noted that crustal melting is not taken into account – adding this process into the model is likely to have an effect at Moho temperatures larger than 1000 °C.

In all experiments, the mantle lithosphere must be eroded and the initial crust thickened before the lower crust is heated sufficiently to generate a dripping instability. Therefore, most of the drips occur above or in the vicinity of upwelling partially molten zones in the mantle, where the newly generated crust “pushes” the older crust into warmer mantle, where it is negatively buoyant. Simultaneously, the depressed crust is heated, which results in a lower effective viscosity that allows Rayleigh-Taylor drips to form in a relatively rapid manner (in agreement with the scaling analysis shown in 3.2.3) For larger values of T_p , erosion of the mantle lithosphere occurs more efficiently, although the original crust remains present for a longer period of time above regions shielded by any leftover mantle lithosphere. At lower T_p this effect plays a more important role and the high viscosity and relatively low density of the mantle lithosphere stabilizes the crust.

As crustal melting is not taken into account in the experiments, the results of these experiments can be regarded as providing conservative estimates on when crustal recycling occurs. Partial melting of the crust would further increase the overall thickness of the crust, which should enhance the recycling effect.

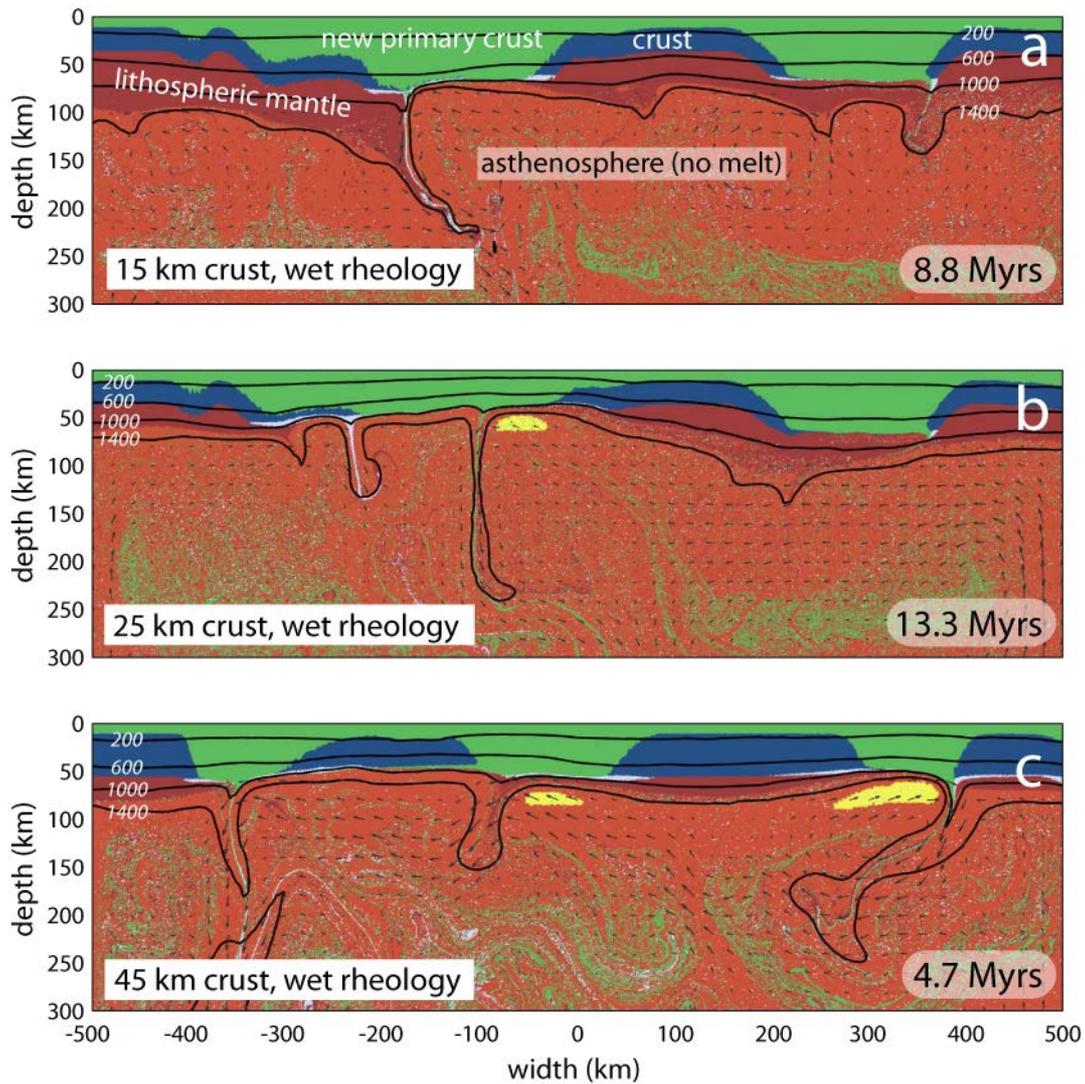


Figure S8. Snapshots from experiments with T_p of 1600 °C and wet mantle rheology. **a**, Initial crustal thickness = 15 km; **b**, initial crustal thickness = 25 km; and, **c**, initial crustal thickness = 45 km. Colours: dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink-brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; green positively and negatively buoyant newly-created crust; and yellow, melt-bearing asthenosphere).

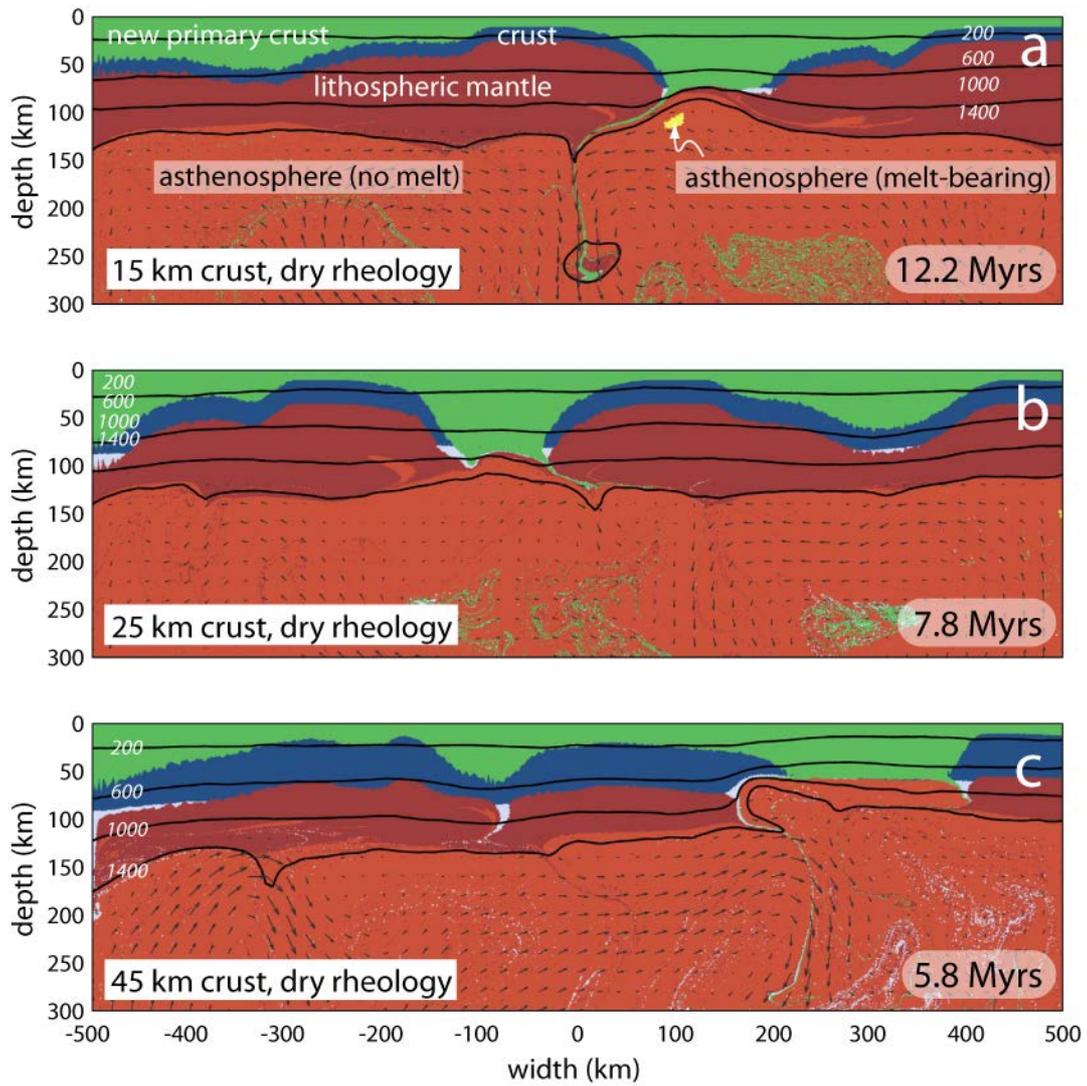


Figure S9. Snapshots from experiments with T_p of 1600 °C and dry mantle rheology. **a**, Initial crustal thickness = 15 km; **b**, initial crustal thickness = 25 km; and, **c**, initial crustal thickness = 45 km. Colours: dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink-brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; and yellow, melt-bearing asthenosphere).

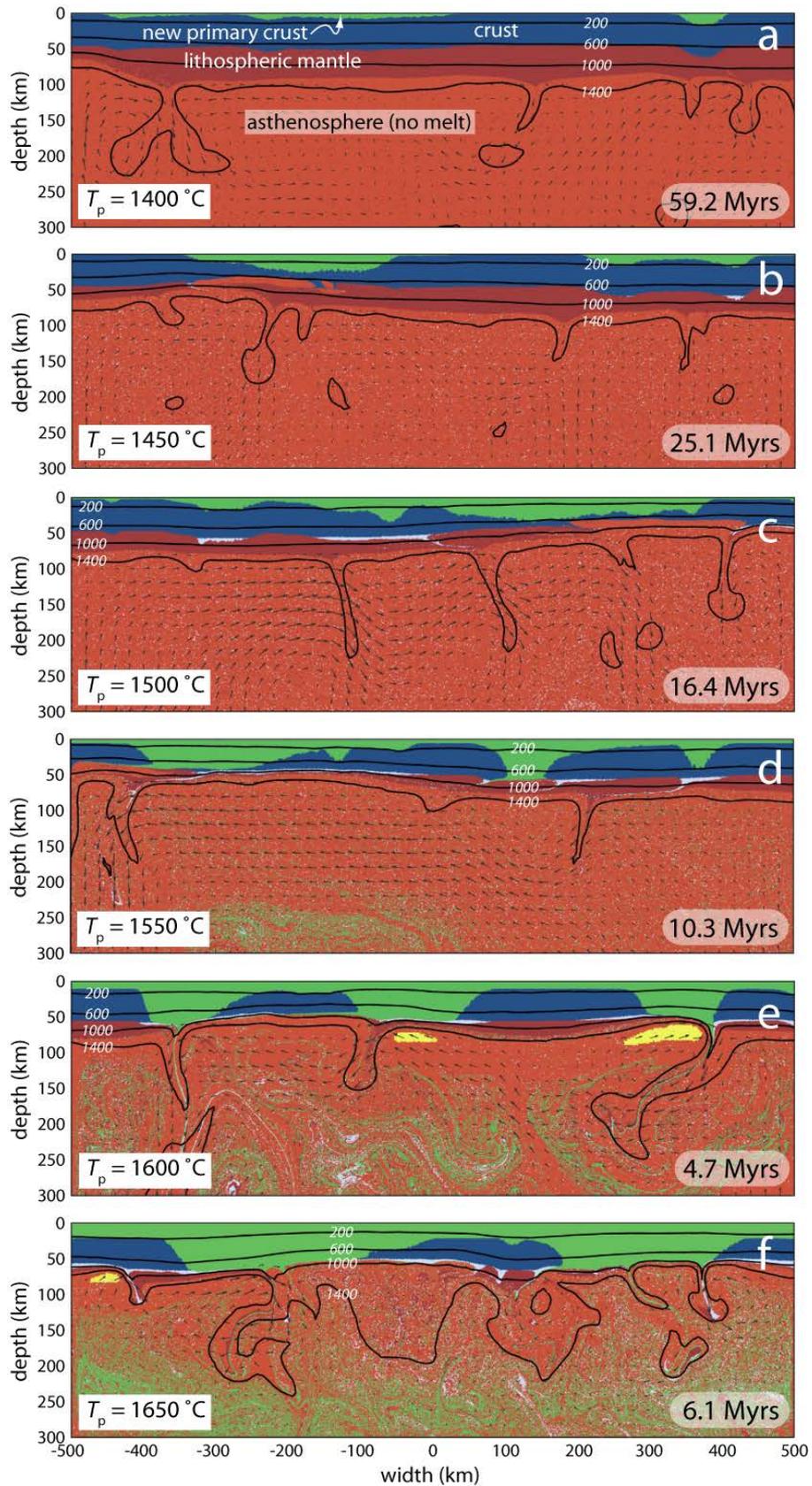


Figure S10. Snapshots from experiments with 45 km thick initial primary crust and wet mantle rheology. **a**, $T_p = 1400\text{ }^\circ\text{C}$; **b**, $T_p = 1450\text{ }^\circ\text{C}$; **c**, $T_p = 1500\text{ }^\circ\text{C}$; **d**, $T_p = 1550\text{ }^\circ\text{C}$; **e**, $T_p = 1600\text{ }^\circ\text{C}$; and, **f**, $T_p = 1650\text{ }^\circ\text{C}$. Colours: dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink-brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; and yellow, melt-bearing asthenosphere).

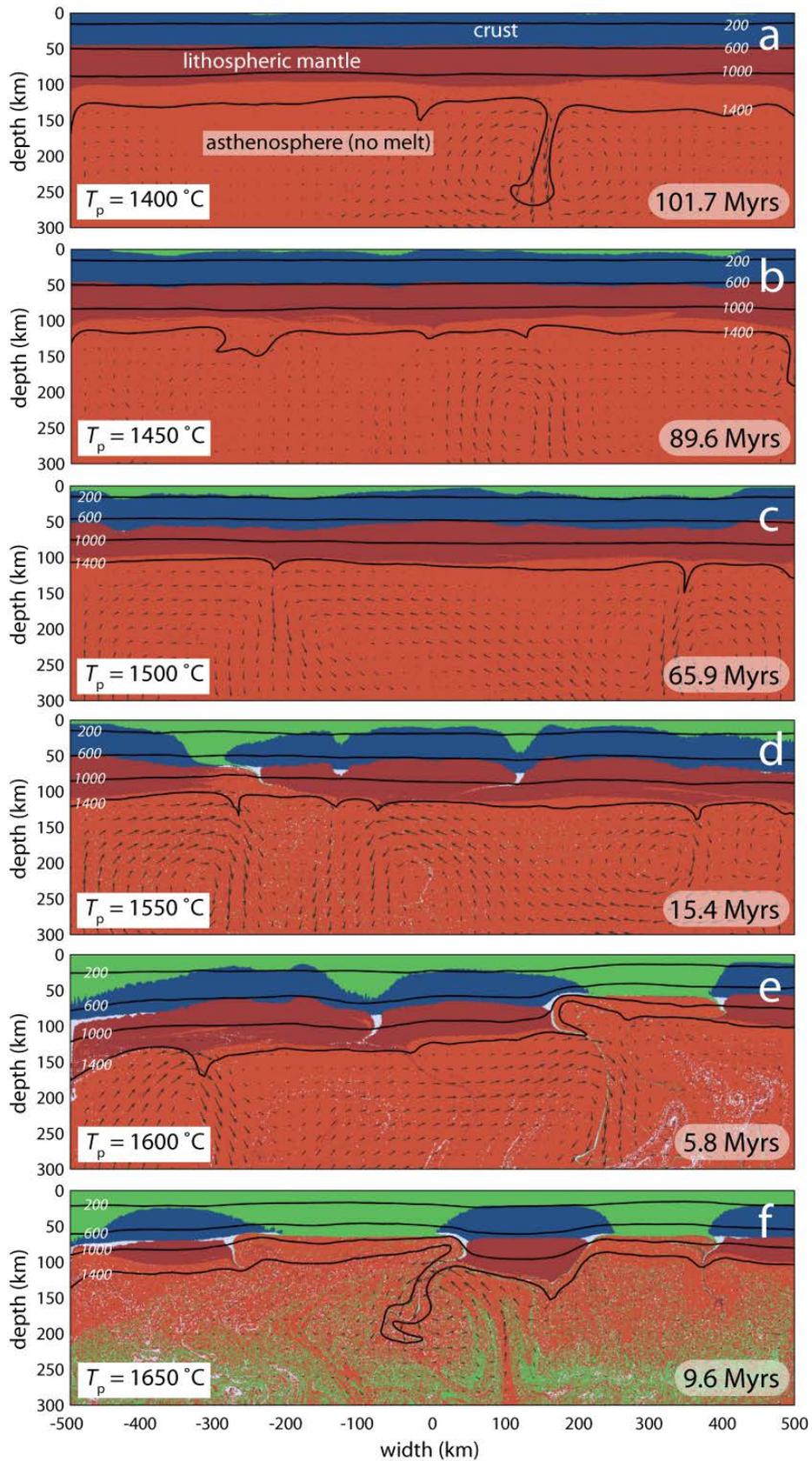


Figure S11. Snapshots from experiments with 45 km thick initial primary crust and dry mantle rheology. **a**, $T_p = 1400\text{ }^\circ\text{C}$; **b**, $T_p = 1450\text{ }^\circ\text{C}$; **c**, $T_p = 1500\text{ }^\circ\text{C}$; **d**, $T_p = 1550\text{ }^\circ\text{C}$; **e**, $T_p = 1600\text{ }^\circ\text{C}$; and, **f**, $T_p = 1650\text{ }^\circ\text{C}$. Colours: dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink-brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; and yellow, melt-bearing asthenosphere).

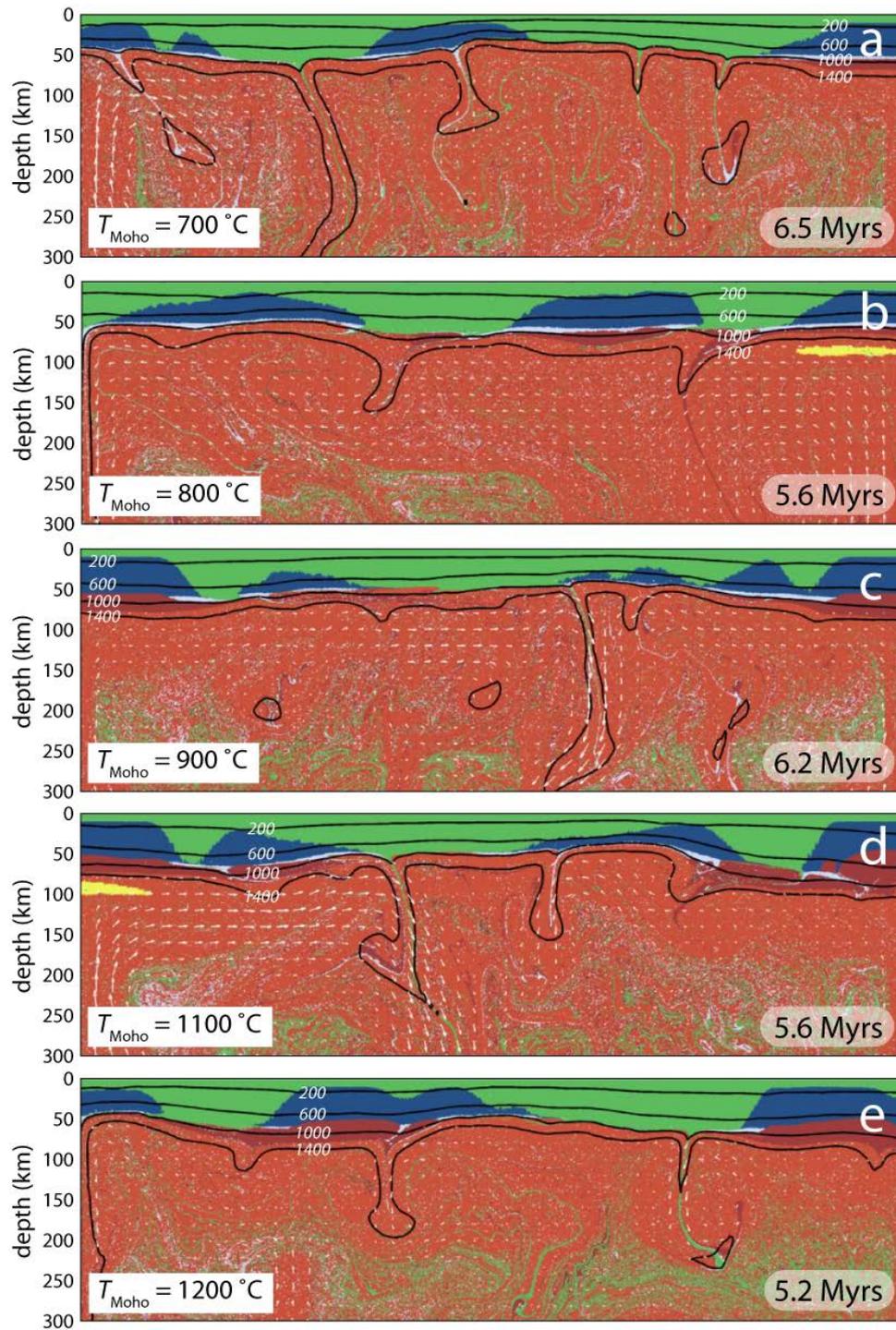


Figure S12. Snapshots from experiments with 45 km thick initial primary crust and wet mantle rheology. **a**, $T_{\text{Moho}} = 700\text{ °C}$; **b**, $T_{\text{Moho}} = 800\text{ °C}$; **c**, $T_{\text{Moho}} = 900\text{ °C}$; **d**, $T_{\text{Moho}} = 1100\text{ °C}$; and, **e**, $T_{\text{Moho}} = 1200\text{ °C}$. Colours: dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink-brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; and yellow, melt-bearing asthenosphere).

3.2.3. Physics of dripping instabilities

All simulations described above are performed with a hydrous crust, which is a worst-case scenario as the crust becomes negatively buoyant only once it is thickened sufficiently (as described in the main text). We have performed additional simulations with an anhydrous crust, in which almost the full crustal thickness is negatively buoyant. However, since the anhydrous crust is relatively cold, the effective viscosity is too high to allow formation of rapid crustal-scale dripping instabilities. Instead, the experiments behave in a fairly similar manner to those described above in which only parts of the base of the crust drip off, generally above or in the vicinity of upwelling partially molten zones in the mantle.

Insight into the basic physics of the dripping instability is obtained using dimensional analysis. Assuming a layer at the base of the crust with thickness H , it will approximately affect a region of width H and sink with the Stokes sinking velocity, given by

$$V_z \propto \frac{C\Delta\rho g H^2}{\eta}$$

where C is a correction factor, $\Delta\rho$ the average density difference between the crust and the underlying mantle, and η the effective viscosity immediately above or below the Moho.

The characteristic velocity of isotherms due to diffusion is given by

$$V_{diff} \propto \frac{\kappa}{H}$$

For a dripping instability to occur, lateral variations in density should exist, which implies that the instability must grow faster than the characteristic diffusion time (so $t_{adv} > t_d$) otherwise the geotherm at the Moho would be equalized. This results in a non-dimensional critical buoyancy Rayleigh number, given by

$$R_b = \frac{V_z}{V_{diff}} = \frac{C\Delta\rho g H^3}{\eta\kappa}$$

which should be larger than 1 for a dripping instability to occur. Some parameters in this equation are well constrained (namely $\kappa = 10^{-6} \text{ m}^2/\text{s}$, $g = 10 \text{ m/s}^2$, $\Delta\rho \sim 50\text{--}200 \text{ kg/m}^3$), whereas others, such as viscosity and layer thickness, are more poorly defined, and C is likely to be a constant based on earlier work (Conrad & Molnar, 1997; Thielmann & Kaus, 2012).

Analysing the numerical experiments at the stage where the dripping instabilities start to form provides a test of the validity of this simple approximation. This task has been undertaken for a number of the experiments in which vertical profiles at the location where a dripping instability occurs were analysed by computing the local thickness and average density difference of the negatively buoyant layer, and determining the maximum viscosity in this region, which generally occurs within the mantle lithosphere. The results of this analysis are summarized in Table S2.

Simulation name	Time [Myrs]	x [km]	Av. $\Delta\rho$ [kg/m ³]	Max η [Pas]	H [km]	T _{moho} [C]	$\Delta\rho g H^3 / (\eta \kappa)$
<i>T_p 1600, Crust 45 km, Wet</i>	1.92	-83	79	5.2x10 ¹⁹	10	996	15.0
<i>T_p 1600, Crust 45 km, Wet</i>	1.92	378	67	4.9x10 ¹⁹	10	986	13.6
<i>T_p 1600, Crust 45 km, Wet</i>	1.97	-320	85	2.5x10 ¹⁹	11	1014	45.6
<i>T_p 1600, Crust 45 km, Wet</i>	2.22	-22	185	1.3x10 ¹⁸	14	1064	3409
<i>T_p 1600, Crust 45 km, Wet</i>	2.20	350	109	1.0x10 ¹⁸	10	1097	1094
<i>T_p 1650, Crust 45 km, Wet</i>	2.61	112	69	3.5x10 ¹⁹	13	994	42.3
<i>T_p 1650, Crust 45 km, Wet</i>	2.56	352	94	3.0x10 ¹⁹	9	964	23.2
<i>T_p 1650, Crust 45 km, Wet</i>	2.56	-127	108	1.9x10 ¹⁹	10	977	56.4
<i>T_p 1500, Crust 45 km, Wet</i>	3.23	-52	62	8.9x10 ¹⁸	11	1041	93.1
<i>T_p 1550, Crust 45 km, Wet</i>	1.67	-128	61	4.0x10 ¹⁹	13	1033	33.23
<i>T_p 1550, Crust 45 km, Wet</i>	1.59	122	85	9.9x10 ¹⁸	13	1036	189
<i>T_p 1550, Crust 45 km, Wet</i>	1.59	369	73	1.4x10 ¹⁹	15	1061	175
<i>T_p 1550, Crust 45 km, Wet</i>	1.63	-318	82	9.5x10 ¹⁸	15	1078	290
<i>T_p 1650, Crust 45 km, Dry</i>	4.32	-127	139	1.4x10 ¹⁹	15	1039	336
<i>T_p 1650, Crust 45 km, Dry</i>	4.33	355	127	5.3x10 ¹⁹	15	1031	81
<i>T_p 1650, Crust 45 km, Dry</i>	4.47	330	152	3.0x10 ¹⁸	20	1162	4056
<i>T_p 1650, Crust 45 km, Dry</i>	4.48	-80	172	2.6x10 ¹⁸	20	1173	5231
<i>T_p 1650, Crust 45 km, Dry</i>	4.69	300	130	1.9 x10 ¹⁹	25	1096	986
<i>T_p 1600, Crust 45 km, Dry</i>	3.32	-80	121	2.5x10 ²⁰	13	1027	11
<i>T_p 1600, Crust 45 km, Dry</i>	3.46	379	125	8.6x10 ¹⁹	15	1046	49
<i>T_p 1600, Crust 45 km, Dry</i>	3.78	270	148	1.1x10 ²⁰	17	1027	64
<i>T_p 1550, Crust 45 km, Dry</i>	3.53	-310	31	5.7x10 ¹⁹	14	1049	15
<i>T_p 1550, Crust 45 km, Dry</i>	3.53	122	126	5.1x10 ¹⁹	15	1095	83
<i>T_p 1600, Crust 25 km, Dry</i>	5.20	-68	165	1.4x10 ¹⁹	17	1147	566
<i>T_p 1600, Crust 25 km, Wet</i>	3.01	260	132	2.5x10 ¹⁸	12	1094	927
<i>T_p 1600, Crust 25 km, Wet</i>	3.12	-190	89	1.3x10 ¹⁹	7	978	23

Table S2. Measured parameters at the onset of a dripping instability from the numerical simulations. Black colors indicate the first dripping instabilities that occur in the model. Red colors indicate drips that form at a later stage. In most cases these instabilities form over significant upwellings.

The results indicate that the onset of the first instabilities is well captured with the expression for R_b given above, provided that a value for C of 0.1 is used, after which drips are predicted to develop as soon as $R_b > 1$. Later instabilities commonly form above the larger partially molten upwellings. In this case, the velocity of the drips has to be not only faster than diffusion but also faster than the upward motion of the isentropically melting mantle. As a result, the condition $R_b > 1$ is insufficient and $R_b > 100$ is typically required.

The simple scaling law may be used to obtain insight into the maximum viscosity and minimum layer thicknesses that are required for a drip to develop at the Moho. Solving for H gives

$$H = \left(\frac{R_{b,crit} \kappa \eta}{C \Delta \rho g} \right)^{\frac{1}{3}}$$

where $R_{b,crit}$ is the critical number (between 1 and 100 for most of the experiments), $C = 0.1$, and all other parameters as described above. Figure S13 shows a comparison of the analytical predictions with the numerical results at the onset of dripping. The agreement is reasonable and shows that a hot mantle lithosphere, which has smaller effective viscosities, will develop smaller drips. As the surface/volume ratio of smaller drips is larger, we expect them to heat in an efficient manner and therefore potentially melt (cf. Elkins-Tanton, 2005).

Our results indicate that layers thinner than 5 km would need a very low viscosity to become negatively buoyant and drip-off, or, alternatively, they would need to be thickened by another process (for example, by an adjacent hot upwelling). Yet, once a drip starts going, our simplified analysis strongly underestimates the velocity of the drips, and numerical simulations show that the resulting drips take most of the negatively buoyant crustal material with them into the mantle. This is the reason that nearly all of the negatively buoyant crust disappears. The stable crust does *not* disappear into the mantle (there is almost no “dark blue” material present in the mantle in the snapshots from the experiments), which further supports our interpretation that the lower crust is indeed removed by RT instabilities. Both positive and negatively buoyant “new” crust is coloured green; the negatively buoyant crust sinks into the mantle through the same mechanism where it is likely to melt as discussed in the main text.

The effect of phase transitions on dynamics of the early Earth has been discussed before. Particularly, Davies (1992) conjectured that the basalt–eclogite transition could play a role by adding negative buoyancy to a plate with a thick crust and thereby add to the negative buoyancy of those plates. Yet, he concluded that:

“It is quite possible that the effect of the basalt–eclogite transformation could have increased the rate of plate tectonics, but perhaps not dramatically and only intermittently” (Davies, 1992).

Our models show that the phase transitions indeed play a major role in the early Earth. Yet, rather than extending the plate-tectonics regime by adding additional negative buoyancy, the phase transitions help to drive rapid crustal recycling of an Earth that was otherwise in the drip tectonics mode. The exact physics of why the Earth switched from a drip-tectonics to a plate-tectonics regime remains to be understood, but our models do suggest that petrological constraints should be taken into account in such models.

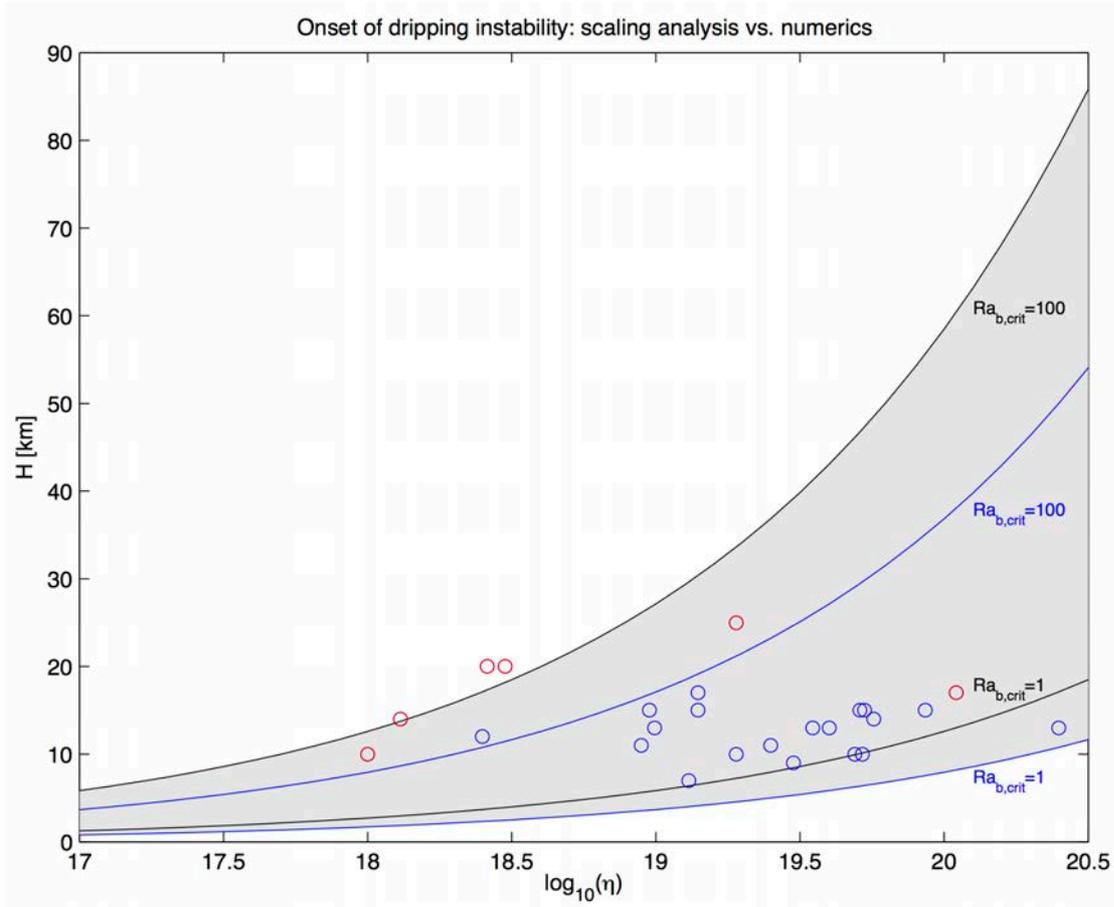


Figure S13: Analytical predictions for the minimum negatively buoyant layer thickness as a function of viscosity that is required for a dripping instability to develop (for this $Ra_b > 1$). Solid lines are predictions assuming $\Delta\rho=50$ (black lines) and $\Delta\rho=200 \text{ kg/m}^3$ (blue lines). Circles are the results from the numerical simulations, where blue indicate the first drips that form in the simulations and red circles indicate drips that form at a later stage (typically above large scale upwellings). The agreement is quite reasonable given the simplicity of the scaling analysis, and shows that only small drips will form in a hotter mantle (which has a lower viscosity).

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