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## 1 **Spreading Continents Kick-Started Plate Tectonics**

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6

7 **Plate tectonics characterises the present-day geodynamics of the Earth. Stresses acting on cold,**  
8 **thick and negatively buoyant oceanic lithosphere are key to the initiation of subduction and the**  
9 **operation of plate tectonics<sup>1,2</sup>. In the Archaean, because the Earth interior was hotter, oceanic**  
10 **crust may have been thicker, making oceanic lithosphere possibly more buoyant than at**  
11 **present<sup>3</sup>. Whether subduction and plate tectonics occurred on the early Earth is ambiguous**  
12 **both in the geological record and in geodynamic models<sup>4</sup>. Here we show that because they were**  
13 **thick and buoyant<sup>5</sup>, early continents produced intra-lithospheric gravitational stresses large**  
14 **enough to drive their gravitational spreading, to initiate subduction at their margins, and to**  
15 **trigger subduction episodes. Our model predicts the co-occurrence of deep to progressively**  
16 **shallow mafic volcanics and arc magmatism within continents in a self-consistent geodynamic**  
17 **framework, explaining the enigmatic multimodal volcanism and tectonic record of Archaean**  
18 **cratons<sup>6</sup>. Moreover, our model predicts a progressive petrological stratification and tectonic**  
19 **structuration of the sub-continental lithospheric mantle, two predictions that are respectively**  
20 **consistent with xenolith<sup>5</sup> and seismic studies, and consistent with the existence of a mid-**  
21 **lithospheric seismic discontinuity<sup>7</sup>. The slow gravitational collapse of early continents could**  
22 **have kick-started transient plate tectonic episodes until, as the Earth interior cooled and**  
23 **oceanic lithosphere became heavier, plate tectonics became self-sustaining.**

24 Present-day plate tectonics is primarily driven by the negative buoyancy of cold subducting plates.

25 Petrological and geochemical proxies of subduction preserved in early continents point to subduction-

26 like processes already operating before 3 billion years (Gyr) ago<sup>8,9</sup> and perhaps as early as 4.1 Gyr

27 ago<sup>10</sup>. However, they are not unequivocal, and geodynamic modelling suggests that the thicker

28 basaltic crust produced by partial melting of a hotter Archaean or Hadean mantle would have

29 increased lithospheric buoyancy and inhibited subduction<sup>3,4</sup>. Therefore, mantle convection under a  
30 stagnant lid with extensive volcanism could have preceded the onset of subduction<sup>11</sup>. In this scenario,  
31 it is classically assumed that the transition from stagnant-lid regime to mobile-lid regime and the  
32 onset of plate tectonics require that convective stresses overcame the strength of the stagnant lid<sup>12</sup> at  
33 some stage in the Archaean.

34 On modern Earth, gravitational stresses due to continental buoyancy can contribute to subduction  
35 initiation<sup>2,13</sup>. The role of continental gravitational stresses as a driver of Archaean lithospheric  
36 deformation has been emphasized<sup>14,15</sup>, however their potential to initiate subduction has been  
37 overlooked. Studies of xenolith in Archaean cratons show that the early continental crust was  
38 underlain by a thick (~200 km), moderately- to strongly-depleted and therefore buoyant lithospheric  
39 mantle<sup>5</sup>. A common model for the formation of early continental lithosphere invokes partial melting  
40 in mantle plumes, leading to magnesium-rich mantle residues (e.g. refractory harzburgites and  
41 dunites) under thick basaltic plateaux<sup>5,16,17</sup>. Partial melting of these thick basaltic crusts, at depth  
42 >40 km, further differentiates the crust into tonalite-trondjemite-granodiorite (TTG) and restitic  
43 material<sup>16,18</sup>.

44 First order calculations show that the horizontal gravitational force acting between a continent 200 km  
45 thick and adjacent oceanic lithosphere is of the order of  $10^{13}$  N m<sup>-1</sup> (see Extended Data Fig. 1),  
46 comparable to that of present-day tectonic forces driving orogenesis<sup>1</sup>. To explore the tectonic impact  
47 of a thick and buoyant continent surrounded by a stagnant lithospheric lid, we produced a series of  
48 two-dimensional thermo-mechanical numerical models of the top 700 km of the Earth, using  
49 temperature-dependent densities and visco-plastic rheologies depending on temperature, melt fraction  
50 and depletion, stress and strain rate (see Methods). The initial temperature field is the horizontally-  
51 averaged temperature profile of a stagnant-lid convection calculation for a mantle ~200 K hotter than  
52 at present (Fig. 1a<sub>0</sub> and Extended Data Fig. 2). The absence of lateral temperature gradients ensures  
53 that no convective stresses act on the lid, allowing us to isolate the dynamic effect of the continent. A  
54 225-km-thick, buoyant and stiff continent (170-km-thick strongly depleted mantle root overlain by  
55 40-km-thick felsic crust; see Fig. 1b<sub>1</sub>) is inserted within the lid, on the left side of the domain to  
56 exploit the symmetry of the problem (Fig. 1a<sub>0</sub>). A 15-km-thick mafic crust covers the whole system

57 (Fig. 1a<sub>0</sub>), accounting for the common occurrence of thick greenstone covers on continents, as well as  
58 thick basaltic crust on the oceanic lid<sup>3</sup>.

59 Our numerical solutions show that the presence of a buoyant continent imparts a horizontal force  
60 large enough to induce a long period (~50 to 150 million years - Myr) of slow collapse of the whole  
61 continental lithosphere (Fig. 1 and Extended Data Fig. 3), in agreement with the dynamics of  
62 spreading for gravity currents<sup>19</sup>. Hence, a continent of larger volume leads to larger gravitational  
63 power and faster collapse. Because of lateral spreading of the continent, the adjacent lithospheric lid  
64 is slowly pushed under its margin (Fig. 1a<sub>1</sub>, and Extended Data Fig. 3 a<sub>1</sub>). For gravitational stress  
65 lower than the yield stress of the oceanic lid, thickening of the margin of the lid is slow and viscous  
66 drips (i.e. Rayleigh-Taylor instabilities) detach from its base (Extended Data Fig. 3 a<sub>0-1</sub>). These  
67 instabilities, typical for stagnant-lid convection<sup>20</sup>, mitigate the thermal thickening of the lid.

68 When gravitational stresses overcome the yield stress of the lithospheric lid, subduction is initiated  
69 (Figs. 1a<sub>1-2</sub>). Depending on the half-width of the continent and its density contrast with the adjacent  
70 oceanic lid (i.e. its gravitational power) three situations can arise: i/ subduction initiates and stalls  
71 (Extended Data Fig. 3b), ii/ the slab detaches then the lid stabilizes (Figs. 1a<sub>3-4</sub> and Extended Data Fig.  
72 3c), or iii/ recurrent detachment of the slab occurs until complete recycling of the oceanic lid followed  
73 by stabilisation (Extended Data Fig. 3d). When the slab reaches a depth of ~ 200 km, slab pull  
74 contributes to drive subduction and rapid rollback of the subducting lid, which in turn promotes  
75 lithospheric boudinage and continental rifting (Fig. 1a<sub>2</sub> and Extended Data Fig. 3c<sub>1-3</sub>). Through  
76 spreading and thinning of the continent, its base rises from 225 km to ~75 km deep on average, and  
77 shallower between lithospheric boudins (Figs. 1a<sub>3</sub> and 2b and Extended Data Figs. 3c<sub>2-3</sub> and 3d<sub>1-5</sub>).  
78 This triggers an episode of deep (~150 km) to shallow (<100 km) decompression melting and  
79 progressive depletion of the ambient fertile mantle (Figs. 2 and 3b). Harzburgites of the continental  
80 mantle are too refractory to melt upon decompression. Polybaric melting of fertile mantle produces a  
81 6 km thick basalt cover and a mantle residue ~ 75 km thick with an averaged depletion of 7.5% (Figs.  
82 2 and 3c). The bulk of depletion occurs over a volcanic episode lasting up to ~13 Myr, although  
83 partial melting persists over a duration of 45 Myr (Fig. 2b). In the last ~25 Myr of melting, the sub-  
84 continental mantle cools until melting stops (Figs. 1a<sub>4</sub> and 2). Decompression melting allows the

85 spreading continent to maintain a minimum chemical thickness of at least 140-150 km. Following the  
86 melting phase, conductive cooling results in the thermal thickening and strengthening of a chemically  
87 stratified cratonic lithosphere (Figs. 1a<sub>4</sub>, 2 and 3d). Over the whole process, the buoyancy of the  
88 continent decreases, subduction stops and a stagnant lid regime is re-established (Fig. 1a<sub>4</sub>).

89 Trade-offs between yield stress, gravitational stress and continental volume determine the initiation of  
90 subduction in our models (Extended Data Figs. 3b and 3d). For a yield stress of 150 to 300 MPa,  
91 consistent with recent estimates of rheological parameters of the lithospheric mantle<sup>21,22</sup>, increasing  
92 the continent width favours subduction initiation (Extended Data Fig. 4). These results suggest an  
93 increasing potential for subduction as continental area increased through time.

94 Our models not only confirm important results from previous studies, but they also provide innovative  
95 explanations for key attributes of Archaean cratons. Both transient subduction and dripping styles are  
96 consistent with previous models of Archaean<sup>23</sup> and modern<sup>13</sup> subduction. As observed in previous  
97 work, the length of the detached segments increases with the yield stress of the lid<sup>23</sup>. Our models  
98 confirm that a combination of the buoyancy, the large viscosity of Archaean sub-continental  
99 lithospheric mantle (SCLM), and large plastic strain weakening prevent the recycling of the SCLM,  
100 which explains its longevity<sup>24</sup>. Our models account for the average thickness (~6 km) and duration of  
101 volcanism (~10 to 50 Myr) of greenstones covers, and predict polybaric (5 to 2 GPa) partial melting,  
102 involving deep (garnet-bearing) to shallow sources (Fig. 2). The MgO content of basaltic melts  
103 produced at these pressures ranges between 11 and 17% (Fig. 2a), consistent with komatiitic basalts  
104 (12-18% MgO) typical of Archaean greenstones<sup>17,26</sup>. Tholeiitic basalts (6-12% MgO), abundant in  
105 Archaean greenstones<sup>26</sup>, can be produced in our model in regions of continental necking where partial  
106 melting can occur at pressure < 3 GPa. Figure 2a shows that to account for the formation of  
107 komatiites (MgO>18%) our model would simply require a mantle potential temperature greater than  
108 1820 K. When subduction starts, arc volcanism is expected at convergent margins while continental  
109 extension and rifting still operate (Fig. 1a<sub>2</sub>). Therefore, our model can explain the metasomatism and  
110 production of sanukitoid melts through migration of younger TTG melts generated by partial melting  
111 of subducting eclogitised basaltic crust (Fig. 3c).

112 On modern Earth, mantle plumes mostly occur away from subduction zones<sup>25</sup>. Hence, the eruption

113 duration and sequential inter-layering of komatiite, tholeiite, calc-alkaline and felsic volcanics,  
114 ubiquitous in Archaean greenstone belts<sup>26-28</sup>, frequently attributed to repeated interaction between  
115 mantle plumes and subduction zones over hundreds of millions of years<sup>6</sup>, remains enigmatic. Our  
116 model predicts the co-occurrence of deep (up to 150 km) to shallow (<100 km) mafic volcanics (Fig.  
117 2b) and arc magmatism in a self-consistent geodynamic framework.

118 Moreover, our model predicts a progressive chemical stratification of the SCLM concomitant with  
119 that of the continental crust and growth of the greenstone cover. This explains the strong geochemical  
120 layering of cratonic mantle inferred by geochemical and petrological studies of mantle xenolith<sup>5</sup>. Pure  
121 shear stretching and thinning during the collapse promotes development of a sub-horizontal litho-  
122 tectonic fabric in the refractory harzburgite and dunite, and to a lesser extent in the accreted  
123 moderately depleted mantle (Fig. 2b). The predicted litho-tectonic layering explains the seismic mid-  
124 lithospheric discontinuity around 100 km depth observed within cratons<sup>29</sup>. This discontinuity<sup>7</sup> could  
125 correspond to the sharp transition predicted by our model between the strongly stretched, strongly  
126 depleted primary root of the continent and the moderately stretched, moderately depleted-to-fertile  
127 mantle accreted through cooling (Fig. 3).

128 We propose that the collapse of early continents was a key process in Archaean geodynamics,  
129 resulting in the concomitant structuration of the mantle root and crust of cratons. This process would  
130 have kick-started transient episodes of plate tectonics, until plate tectonics became self-sustained,  
131 through increasing continental area<sup>30</sup> and decreasing buoyancy of oceanic plates<sup>3</sup>.

132

### 133 **METHODS SUMMARY**

134 We solve the problem of conservation of mass, momentum and energy for incompressible mantle  
135 flow and lithosphere deformation in the top 700 km of the Earth, using the particle-in-cell finite-  
136 element code Ellipsis<sup>12</sup> (freely available at <http://www.geodynamics.org/cig/software/ellipsis3d/>).

137 These equations are solved along with constitutive relations for visco-plastic rheology depending on  
138 temperature, stress and strain rate, and melt fraction and depletion (see Methods and Extended Data  
139 Table 1). Over 200 calculations were conducted varying the lateral extent of the model domain (4200  
140 to 16800 km), the half-width (400 to 1400 km) and thickness (175 or 225 km) of the continents, the

141 limiting yield stress of mantle rocks (100 to 500 MPa), density of the top 75 km of the oceanic lid  
142 mantle (3395 down to 3360 kg m<sup>-3</sup>), and various modes of melt extraction. The reference densities of  
143 all mantle rocks depend on temperature (Extended Data Table 1), and the density of basalts also  
144 depends on pressure to simulate eclogitisation. Our experiments take into account the thermo-  
145 mechanical impacts of partial melting, which allows us to map the evolution of the depletion and  
146 reveal the development of the lithospheric mantle layering. The specific version of Ellipsis and input  
147 scripts used for this work are available from the first author upon request.

148

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231 The authors declare no competing interests. Readers are welcome to comment on the online version of  
232 the paper. Correspondence and requests for Ellipsis code and input scripts should be addressed to  
233 P.F.R. ([patrice.rey@sydney.edu.au](mailto:patrice.rey@sydney.edu.au)).

234

235 **Figure 1 | Numerical solution of an example of continent collapse leading to subduction.**

236 Modelling setup (**a**<sub>0</sub>) and computed snapshots (**a**<sub>1-4</sub>) for a 700-km-deep, 6300-km-long box including a  
237 225-km-thick continent, of half-width 800 km. All mantle rocks have a limiting yield stress of  
238 300 MPa. Mantle cooler than 1620 K is in blue (darker blue is hotter), mantle hotter than 1620 K is in  
239 pink (darker pink is hotter). Regions of depletion due to partial melting of ambient fertile mantle are  
240 hatched. Compositional structure, reference densities and reference rheological profile are shown in  
241 (**b**<sub>1</sub>) for the continent, and (**b**<sub>2</sub>) for the adjacent lithospheric lid. This numerical solution documents the  
242 long phase of slow continental spreading leading to the initiation of a slab (**a**<sub>1-2</sub>). Once the slab  
243 reaches a depth of ~200 km, slab pull contributes to drive subduction, rollback and continental  
244 boudinage (**a**<sub>2</sub>, in some experiments boudinage leads to rifting) and slab detachment (**a**<sub>3</sub>). In this  
245 experiment, the detachment of the slab is followed by a long period of thermal relaxation and  
246 stabilisation during which the thickness of the continent increases through cooling and incorporation  
247 of the moderately depleted mantle (**a**<sub>4</sub>).

248 **Figure 2 | Layering of the continental lithosphere through thinning and progressive accretion of**

249 **moderately depleted mantle.** a/ Points located at depth  $A_0$  to  $A_3$  before spreading, are exhumed  
250 during spreading to locations  $A_0'$  to  $A_3'$  following the blue pressure-temperature-time (P-T-t) path.

251 The geotherm intersects the solidus and the temperature in the partially molten column remains close  
252 to the solidus because latent heat is continuously extracted with the melt once melt fraction reaches

253 1%. Melt is extracted from various depths following the melt adiabat (yellow arrows). The region  
254 between the hydrous fertile mantle solidus and liquidus is mapped for MgO content (see Method).  
255 The deeper part of the column produces komatiitic basalts (dark blue shading), while partial melting  
256 at pressure < 3 GPa produces tholeiitic basalts (pale blue shading). b/ Temporal evolution of the  
257 laterally averaged depletion (cyan), partial melting (yellow), and density interfaces (thick dashed  
258 lines). As spreading and thinning proceeds, pure shear fabrics (shown as finite strain ellipses) develop  
259 in the refractory mantle and in the moderately depleted mantle, which records a shorter strain history.  
260 The base of the partially molten column remains close ~150 km whereas its top progressively rises  
261 from ~150 km at the beginning of partial melting (at ~44 Myr) to an average of ~75 km below the  
262 surface (at ~55 Myr). From 55 Myr, spreading slows down and progressive cooling reduces the  
263 amount of melt, until partial melting stops at ~82 Myr. This results in the progressive chemical  
264 stratification of the sub-continental lithospheric mantle.

265 **Figure 3 | Proposed model for the co-evolution of cratonic crust and sub-continental**  
266 **lithospheric mantle.** Integrating the results of our numerical experiments with petrological data  
267 supports a model linking the formation of continents to the initiation of subduction at their margins,  
268 through the process of continent collapse. This model predicts the layering of the sub-continental  
269 lithospheric mantle (SCLM), polybaric and multimodal volcanism recorded in greenstone covers, and  
270 the metasomatism of the SCLM. **a**, Deep mantle partial melting leads to the formation of an oceanic  
271 plateau that differentiates into a continent. The residue of mantle melting forms the strongly depleted  
272 harzburgite root of the continent, while the deeper part of the basaltic crust differentiates via partial  
273 melting into tonalite-trondjemite-granodiorite (TTG). **b**, As the continent grows in thickness and  
274 length, excess gravitational potential energy drives its collapse and the shortening of the adjacent  
275 oceanic lid (in grey). During collapse and thinning of the continent, decompression melting (yellow)  
276 of the fertile ambient mantle and extraction of deep (komatiitic basalt) to shallow (tholeiite) melt  
277 contribute to the growth of the greenstone cover, and to the formation of a moderately depleted  
278 mantle layer. **c**, Due to the horizontal push of the collapsing continent, the thickened margin of the  
279 oceanic lid subducts underneath the continental margin. Partial melting of the thick, eclogitised  
280 oceanic crust produces TTG melts (purple), which metasomatise both the mantle wedge and the

281 lithospheric mantle. Melting of the hydrated and metasomatised mantle wedge produces calc-alkaline  
282 to sanukitoid melts (orange). **d**, Following the detachment of the slab, and once the gravitational  
283 power of the continent is too small deform its surrounding, the continent thicken through thermal  
284 relaxation and cooling, first incorporating the layer of moderately depleted mantle and then a layer of  
285 un-melted fertile mantle.

286

## 287 **METHODS**

288 We use Ellipsis, a Lagrangian integration point finite-element code<sup>12,31</sup>, to solve the governing  
289 equations of momentum, mass and energy in incompressible flow. Our reference Cartesian 2D  
290 numerical model is 700 km deep and 6300 km long. The Stokes equation that balances buoyancy  
291 forces, viscous stresses and pressure gradients is solved on a Eulerian computational grid made of  
292 32 x 288 cells with uniform spacing. We have verified validity of our results by testing a higher  
293 resolution grid of 64 x 576 cells. Each cell is populated by 100 Lagrangian particles (total > 920,000  
294 particles) tracking the material properties and therefore material interfaces during mantle flow and  
295 lithospheric deformation. A free slip condition is applied to the vertical and horizontal boundaries of  
296 the modelling domain. The temperature in the upper horizontal boundary is maintained at 293 K,  
297 while the base is maintained at 1873 K.

298 Each material is characterised by a range of thermal and mechanical properties (Extended Data  
299 Table 1) including: density, coefficient of thermal expansion, heat capacity ( $1000 \text{ J kg}^{-1} \text{ K}^{-1}$ ), heat  
300 diffusivity ( $9 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ ), radiogenic heat production, cohesion, coefficient of friction, limiting  
301 pressure-independent yield stress, dislocation creep parameters ( $A$ ,  $n$ ,  $E$ , Extended Data Table 1), and  
302 limiting lower and upper viscosities respectively  $10^{18} \text{ Pa s}$  and  $5 \times 10^{23} \text{ Pa s}$ .

303 Our strategy consists in setting up a model in which the mantle convects under a stagnant lid, the top  
304 15 km of which are made of weak basaltic crust simulating hydrothermally altered basalts (Extended  
305 Data Fig. 2). After reaching steady-state equilibrium, the temperature field is laterally averaged  
306 (Extended Data Fig. 2) and linearised (i.e. temperature at the surface is 293 K, temperature at 150 km  
307 depth and below is 1820 K). This ensures that no convective stress acts on the lid in the initial  
308 condition. Finally, we replace a segment of the oceanic lid by a continent made of depleted mantle

309 underneath a continental crust. We take advantage of the symmetry of the model, and we insert the  
 310 continent along the left side of the model.

311 We model the mantle as a visco-plastic material with temperature, stress, strain-rate and melt  
 312 dependent viscosity. The viscous rheology of the mantle is based on dry olivine<sup>32</sup> (Extended Data  
 313 Table 1). Its plastic rheology is approximated with a Coulomb failure criterion<sup>33</sup> with cohesion  
 314 40 MPa, a coefficient of friction ( $\mu_{ref}$ ) 0.268, and a weakening factor dependent on the accumulated  
 315 plastic strain ( $\epsilon_p$ ). The coefficient of friction ( $\mu$ ) evolves as a function of the total plastic strain as:

316 for  $\epsilon_p < 0.15$ :  $\mu = \mu_{ref} \times (1 - (1 - 0.0373) \times (\epsilon_p / 0.15)^{0.25})$

317 for  $\epsilon_p > 0.15$ :  $\mu = \mu_{ref} \times 0.0373$

318 This weakening leads to fault-like strain localisation with nominal viscosity  $\sim 25$  times weaker than  
 319 that of the surrounding rocks when plastic strain reaches 15%. This weakening, which is a key factor  
 320 for the operation of plate tectonics on present Earth<sup>20</sup>, simulates the formation of phyllosilicates  
 321 (serpentine, talc, micas) during the strain-induced hydration of mantle rocks and basalts. For semi-  
 322 brittle deformation independent of pressure, we impose an upper limiting yield stress<sup>34,35</sup> in the range  
 323 of 100 to 500 MPa.

324 The top 15 km of the model consist of a thick layer of basalts, which simulates both the oceanic crust,  
 325 and the greenstone cover on the Archaean continent. Because these basalts were largely emplaced  
 326 below sea level<sup>36</sup>, they are strongly hydrothermally altered. For simplicity, we assume that this layer  
 327 has a nominal viscosity 1000 times weaker than the underlying mantle, a weak cohesion of 1 MPa, a  
 328 reference coefficient of friction of 0.134 and the same weakening properties than the rest of the lid.

329 This weak layer of basalt helps the decoupling between the subducting slab and the upper plate  
 330 allowing for one side subduction. It also mitigates the absence of free surface, by allowing the upper  
 331 plate to bend downward more realistically to form a deeper trench. The reference density of this  
 332 material ( $3000 \text{ kg m}^{-3}$ ) is sensitive to pressure, and increases upon burial up to  $3540 \text{ kg m}^{-3}$  to  
 333 simulate eclogitisation<sup>37</sup>. When subduction initiates, this weak material lubricates the Benioff plane,  
 334 which, with eclogitisation, facilitates subduction by decoupling the slab from the continental margin.  
 335 This simulates the impact of slab dehydration, which in nature lubricates the Benioff plane.

336 Modelling continent differentiation is beyond the scope of this paper, and we simply replace a 400 to  
337 1200 km long segment of the oceanic lid by a continent, which includes a 40-km-thick crust (density  
338  $2850 \text{ kg m}^{-3}$ ), covered by a 15 km thick flood basalt (density  $3000 \text{ kg m}^{-3}$ ). This crust stands above  
339 170 km of depleted mantle (density  $3310 \text{ kg m}^{-3}$ ) with a nominal viscosity 100 times stronger than the  
340 adjacent lid to account for dehydration<sup>38</sup>. We assume the same limiting yield stress than that of other  
341 mantle rocks that we vary in the range of 100 to 500 MPa. The continental crust is assumed to be 100  
342 times more viscous than a weak mafic granulite<sup>39</sup> (Extended Data Table 1). A limiting yield stress of  
343 250 MPa is imposed in the crust to simulate semi-brittle regime. This strong continent represents a  
344 significant buoyant anomaly, imparting a horizontal gravitational force on its surrounding.

345 We use the hydrous mantle solidus and liquidus of ref. 40 to model partial melting in the sub-  
346 continental fertile mantle. For the depleted root of the continent we increase the hydrous mantle  
347 solidus by 200 K, a reasonable assumption for the harzburgite solidus<sup>41</sup>. Melt fraction is calculated at  
348 each time step using eq. 21 of ref. 42, whereas MgO content of accumulated melt in Fig.2 is  
349 calculated using ref. 43. Because decompression and exhumation of the mantle occur very slowly ( $\sim$   
350 to  $4 \text{ mm y}^{-1}$ ) at the pace imposed by the spreading of the continent, one can expect the melt to  
351 segregate and to pond at the top of the partially molten column, before it escapes to the surface  
352 through dikes. Therefore, we assume that melt in the partially molten column is extracted when the  
353 melt fraction reaches 1%. In this case, latent heat escapes with the melt, and the ascending depleted  
354 residue follows the solidus closely. We assume the fusion entropy to be  $400 \text{ J kg}^{-1} \text{ K}^{-1}$ . We have also  
355 tested the isentropic case in which the melt stays in the source. In this case, lithospheric boudinage  
356 controls continental spreading rather than homogeneous thinning.

357 The small fraction of melt ( $<1\%$ ) has only a modest impact on the buoyancy of the partially molten  
358 region. However, the density of the residue decreases as it becomes more depleted. We assume a  
359 maximum density decrease of 1.5% upon full depletion. Because even a small fraction of melt  
360 lubricates grain boundaries, it affects the viscosity of the partially molten column. Hence, we impose  
361 a linear viscosity drop to a maximum of one order of magnitude when the melt fraction reaches 1%.

362 On the other hand, partial melting drains water out of the solid matrix and reduces the number of

363 phases. Although the impact of dehydration on the viscosity may not be as significant as previously  
364 thought<sup>44</sup>, both processes should contribute to increase the viscosity of the depleted residue once its  
365 temperature drops below the solidus. In our experiment we impose an increase in viscosity  
366 proportional to depletion, assuming a viscosity increase of two orders of magnitude for 100%  
367 depletion.

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401 **Extended Data Figure 1 | Gravitational force between continent and oceanic lid. a**, Depth profile  
402 of the difference in lithostatic pressure  $\sigma_{zz}$  between an oceanic lid 150 km thick and a continent (1)  
403 150 km, (2) 200 km and (3) 250 km thick. The vertical integration of the lithostatic pressure  
404 difference ( $\Delta\sigma_{zz}$ ) is the resulting gravitational force  $Fg$  acting between the Archaean continent and the  
405 adjacent lithospheric lid. In all cases, this force is  $> 7 \times 10^{13} \text{ N m}^{-1}$ , comparable to or larger than the  
406 present-day tectonic forces driving orogenesis<sup>1</sup>. **b**, Reference density structure of the continent and  
407 oceanic lithosphere (densities of depleted and fertile mantle are from Ref. 5). All densities vary with  
408 temperature with a coefficient of thermal expansion  $\alpha = 3 \times 10^{-5} \text{ K}^{-1}$ . We assume a linear geotherm in  
409 the oceanic plate ( $T_{(z=0)} = 293 \text{ K}$ ,  $T_{(z=150\text{km})} = 1820 \text{ K}$ ) above a convective mantle with an average  
410 temperature of 1820 K.

411 **Extended Data Figure 2 | Stagnant lid convection model before lateral averaging and**  
412 **introduction of a continent.** The temperature field in our experiments derives from the lateral  
413 averaging of an experiment in which the mantle is let to convect under constant top boundary  
414 temperatures (293 K) and internal radiogenic heat production ( $1.36 \times 10^{-8} \text{ W m}^{-3}$ ). The coefficient of  
415 thermal expansion is equal to  $3 \times 10^{-5} \text{ K}^{-1}$ , the thermal diffusivity to  $0.9 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$  and the heat  
416 capacity to  $1000 \text{ J kg}^{-1} \text{ K}^{-1}$ , and the Rayleigh number of the convecting mantle is between  $10^6$  and  $10^7$ .  
417 The snapshot shows the temperature field after  $\sim 1$  Gyr of evolution. In this experiment, a lid develops  
418 that remains stagnant. The formation of cold drips from the lower, unstable, part of the stagnant lid,



419 and conductive cooling balance each other to maintain the thickness of the stagnant lid. In the lid, the  
420 conductive geotherm is such that a temperature of 1620 K is reached at ~100 km depth, and 1820 K  
421 (the average temperature in the convecting mantle) is reached at ~150 km depth.

422 **Extended Data Figure 3 | Numerical solutions for various models showing contrasting tectonic**  
423 **evolution. a**, In this experiment, all parameters are as in Fig. 1 except the continent half-width that is  
424 500 km. In the case of stable continental collapse, cold drips form faster than subduction can initiate,  
425 which stabilises the oceanic lid. **a<sub>1</sub>**, During and after spreading and thinning of the continent, a layer  
426 of mostly fertile mantle is accreted at the base of the continent through cooling. **b**, This experiment is  
427 in all aspects similar to that presented in Fig. 1, but for the oceanic lid that includes 75 km of buoyant  
428 lithospheric mantle (in purple) with a reference density of  $3365 \text{ kg m}^{-3}$  (i.e.  $35 \text{ kg m}^{-3}$  less dense than  
429 non-depleted mantle rocks). **b<sub>0-1</sub>**, shows i/ homogeneous continental spreading with decompression  
430 melting, and ii/ the initiation of a slab which stalls underneath a long-lived orogenic wedge. The same  
431 experiment with a buoyant mantle lid with a reference density of  $3370 \text{ kg m}^{-3}$  leads to subduction. **b<sub>2</sub>**,  
432 **b<sub>3</sub>**, However, because of its buoyancy the oceanic slab stalls, although the very base of the oceanic lid  
433 is dragged into the asthenosphere. **c**, In this experiment, the continent has a half-width of 600 km and  
434 all mantle rocks have a limiting yield stress of 200 MPa. **d**, In this experiment, the limiting yield  
435 stress of the strongly depleted continental mantle (in green) is increased to 500 MPa to take into  
436 account the possible plastic strengthening of the depleted, and therefore dry, mantle (all other mantle  
437 rocks have a limiting yield stress of 300 MPa). Comparison to Fig. 1 - which shows the same  
438 experiment but with all mantle rocks with a limiting yield stress of 300 MPa - illustrates that a  
439 stronger continent deforms in a more heterogeneous manner since following an episode of spreading  
440 and thinning (**d<sub>1</sub>**), strain localisation and rifting divide the continent in two (**d<sub>2</sub>**), before stabilisation  
441 and cooling (**d<sub>3</sub>**). **d<sub>0-2</sub>**, The continental spreading is heterogeneous, subduction of the oceanic lid  
442 initiates, the slab detaches and a continental rift initiates. **d<sub>3</sub>**, Continental rifting, recurrent slab  
443 detachment and trench-retreat occur, and two continental blocks move away from each other with  
444 little internal deformation. **d<sub>4,5</sub>**, Rifting and subduction stop, while cooling re-establish a stagnant lid.  
445 In all cases, a protracted phase of decompression melting lasting several tens of Myr is coeval with  
446 spreading and rifting.

447 **Extended Data Figure 4 | Tectonic phase diagram: subduction versus stable lid regime as a**  
 448 **function of yield stress and continent width.** Two series of calculations were carried out (with  
 449 continental thickness of 175 km and 225 km), systematically varying the half-width of the continent  
 450 (from 400 to 1200 km) and the limiting yield stress of all mantle rocks (from 100 to 500 MPa).  
 451 Depending on the competition between the gravitational driving power of the buoyant continent and  
 452 the combined viscous resistance of the continent and oceanic lid, the continental collapse either leads  
 453 to the subduction of the oceanic lid under the continental margin, or not. Coloured dots and the  
 454 continuous black thick line represent outcomes of numerical experiments for the 175 km thick  
 455 continent. The thick dashed line separates the stable-lid domain from the subduction domain in the  
 456 case of the 225 km thick continent. The arrow illustrates that continental rifting, because it reduces the  
 457 half-width of continents, stabilises Archaean oceanic lids.

458

459 **Extended Data Table 1 | Thermal and mechanical parameters**

	Reference Density ( $\text{kg m}^{-3}$ )	Thermal expansion ( $\text{K}^{-1}$ )	Radiogenic Heat ( $\text{W kg}^{-1}$ )	Cohesion (MPa)	Coefficient of friction	Limiting yield stress (MPa)	$A$ ( $\text{MPa}^{-n} \text{s}^{-1}$ )	$E$ ( $\text{kJ mol}^{-1}$ )	$n$
Continental crust	2850	0	$2.52 \cdot 10^{-10}$	40	0.268	250	$2 \cdot 10^{-2}$	244	3.2
Basalts	3000	0	$2.52 \cdot 10^{-10}$	1	0.134	50	$7 \cdot 10^7$	520	3
SCLM	3310	$3 \cdot 10^{-5}$	$4 \cdot 10^{-12}$	40	0.268	100-500	$7 \cdot 10^2$	520	3
Mantle	3395	$3 \cdot 10^{-5}$	$4 \cdot 10^{-12}$	40	0.268	100-500	$7 \cdot 10^4$	520	3

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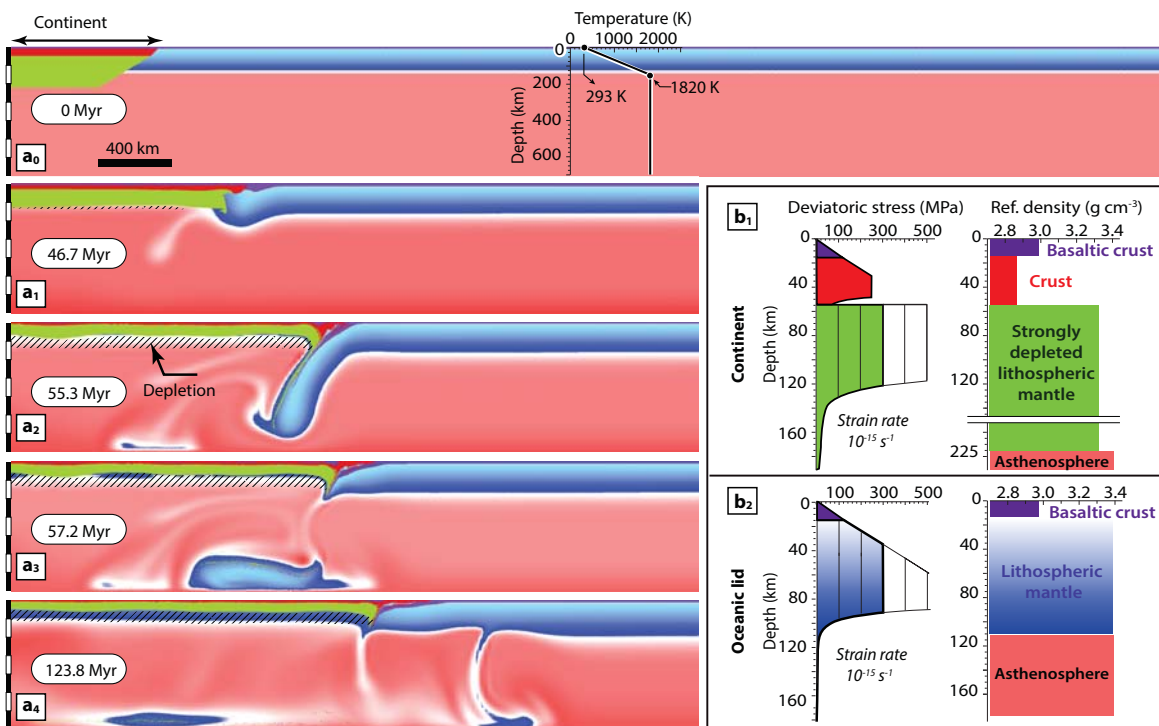
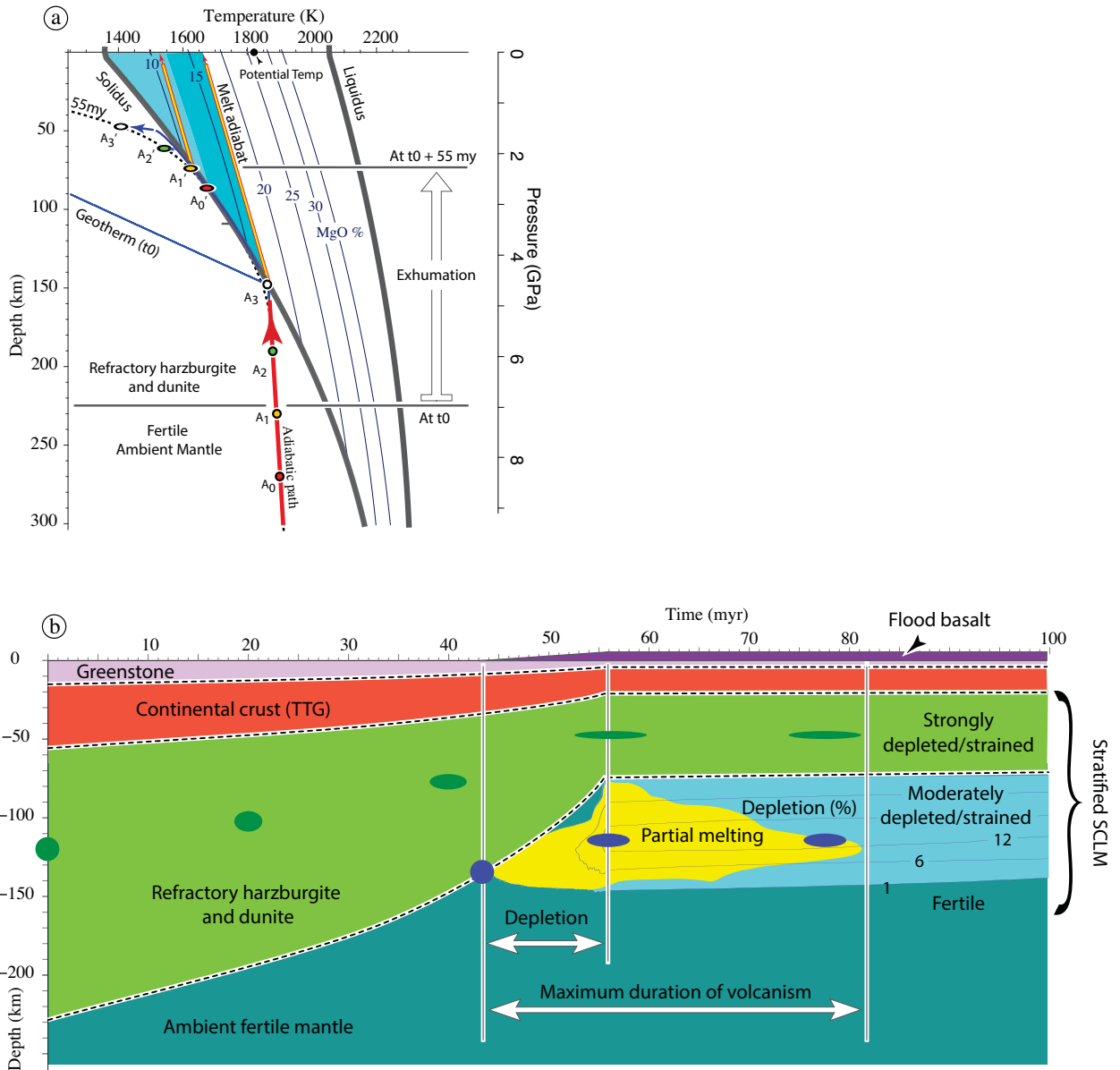
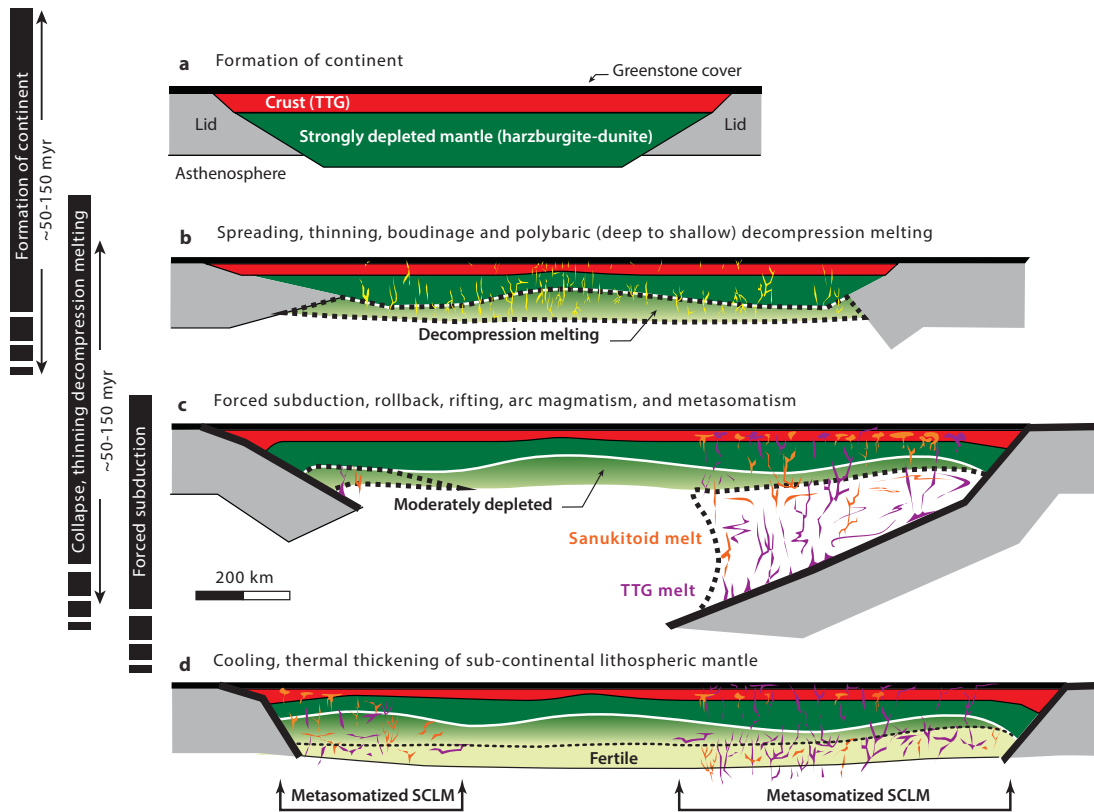


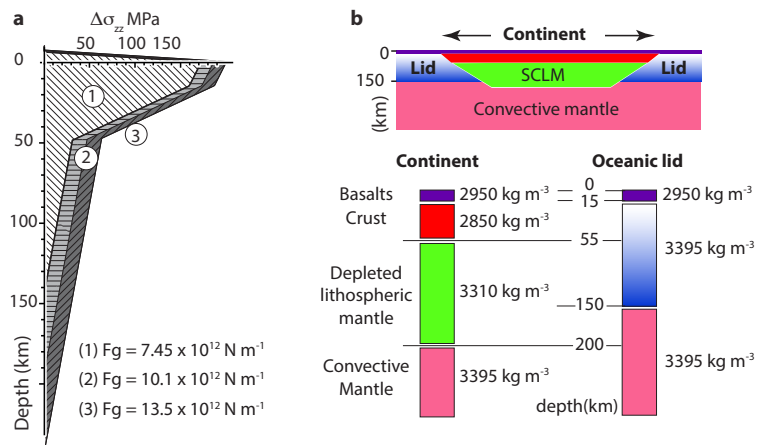
Figure 1 | Numerical solution of an example of continent collapse leading to subduction.



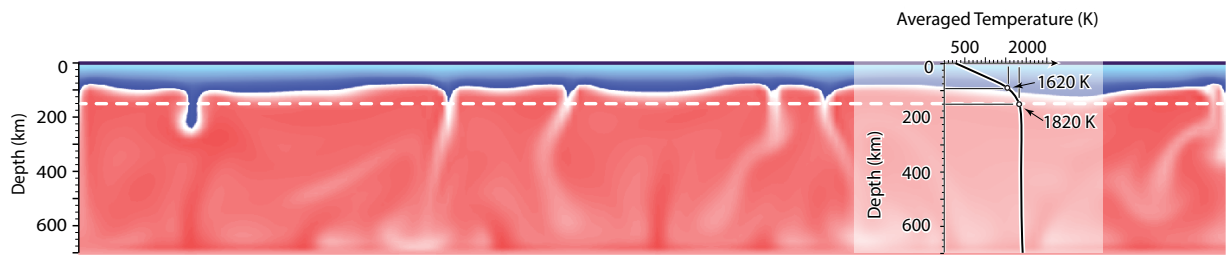
**Figure 2 | Layering of the continental lithosphere through thinning and progressive accretion of moderately depleted mantle.**



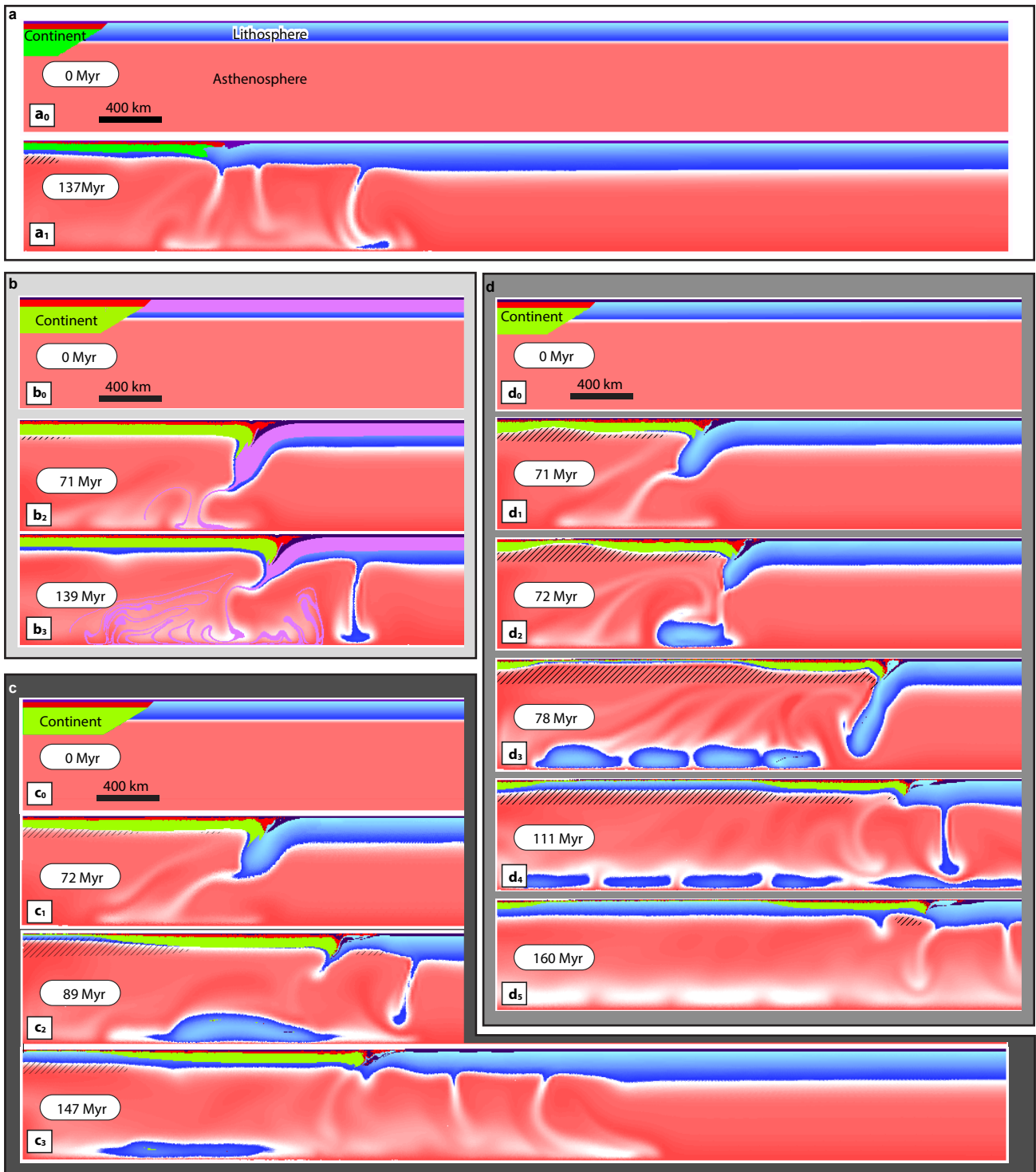
**Figure 3 | Proposed model for the co-evolution of cratonic crust and subcontinental lithospheric mantle.**



Extended Data Figure 1| Gravitational force between continent and oceanic lid.

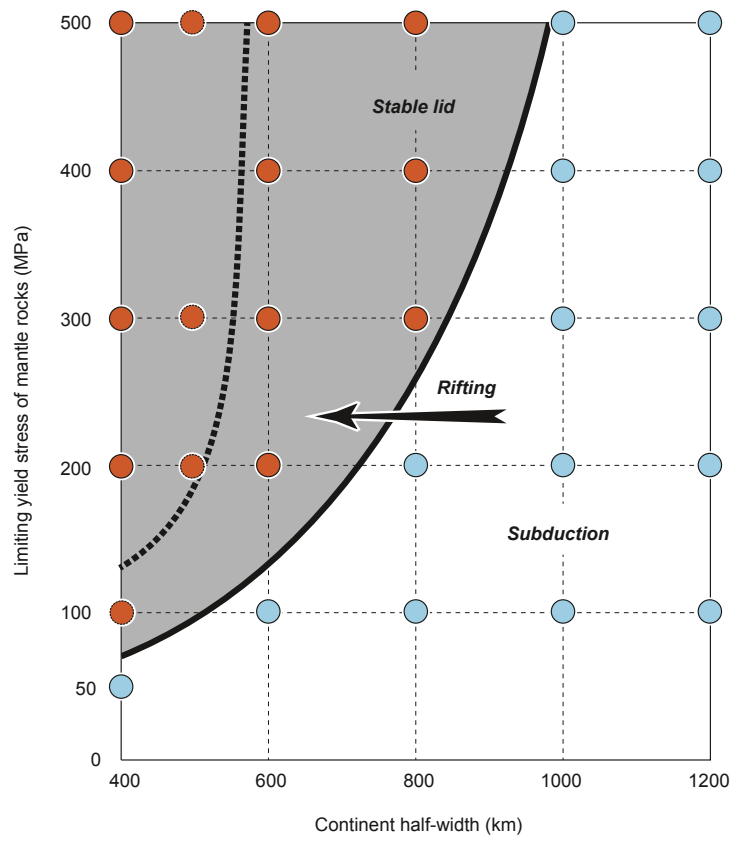


Extended Data Figure 2 | Stagnant lid convection model before lateral averaging and introduction of a continent.



Extended Data Figure 3 | Numerical solutions for various models showing contrasting tectonic evolution.





Extended Data Figure 4 | Tectonic phase diagram: subduction vs stable lid regime as a function of yield stress and continent width.