



HAL
open science

Strain and retrogression partitioning explain long-term stability of crustal roots in stable continents

Bénédicte Cenki-Tok, P. F. Rey, D. Arcay

► **To cite this version:**

Bénédicte Cenki-Tok, P. F. Rey, D. Arcay. Strain and retrogression partitioning explain long-term stability of crustal roots in stable continents. *Geology*, 2020, 48 (7), pp.658-662. 10.1130/G47301.1 . hal-02547410v2

HAL Id: hal-02547410

<https://hal.umontpellier.fr/hal-02547410v2>

Submitted on 23 Nov 2020

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 ¹GSA Data Repository item 2020xxx, Table DR1 (thermal and mechanical parameters),
2 Figure DR1 (slow modeling results), Figure DR2 (fast modeling results), and the Python
3 input file (Script-G47301_285-Cenki-Tok-et.al.ipynb), is available online at
4 <http://www.geosociety.org/datarepository/2020/>, or on request from
5 editing@geosociety.org.

6

7 **Strain and retrogression partitioning explain long-term**
8 **stability of crustal roots in stable continents**

9 **B. Cenki-Tok^{1,2*}, P.F. Rey² and D. Arcay¹**

10 ¹*Géosciences Montpellier, Université de Montpellier, CNRS, 34095 Montpellier cedex 5,*
11 *France*

12 ²*Earthbyte Research Group, ARC ITRH Basin GENESIS Hub, School of Geosciences,*
13 *University of Sydney, Sydney, New South Wales 2006, Australia*

14 *E-mail: benedicte.cenki-tok@umontpellier.fr

15 **ABSTRACT**

16 Away from tectonically active regions, the continental crust has an average
17 thickness of 40 ± 1 km. Yet, it shows a remarkable variability from 25 to 65 km,
18 comparable to that of the most tectonically active regions. Here, we consider the problem
19 of the formation and preservation of anomalous deep crustal roots in stable
20 intracontinental regions. Using two-dimensional thermomechanical experiments, we
21 show that the interplay between partial melting, the formation of garnet-pyroxene-bearing
22 rocks, and their strain rate-dependent retrogression result in the preservation of thick and
23 strong crustal roots. We argue that it is the partitioning into narrow regions of strain,
24 retrogression, and weakening coupled into a positive feedback loop that explains why

25 strong high-grade crustal roots remain largely immune to gravitational stresses and are
26 able to persist over hundreds of millions of years.

27 INTRODUCTION

28 The crust-mantle transition is generally well-defined on geophysical images,
29 enabling detailed knowledge of crustal thickness at global and regional scales (Prodehl et
30 al., 2013). Discarding tectonically active regions, the thickness of the stable continental
31 crust has a global average of ~40 km (± 1 km error on the calculated average crustal
32 thickness) (Christensen and Mooney, 1995; Fig. 1A). Yet, crustal root anomalies as much
33 as 65 km deep exist in all stable continents (e.g., Szwillus et al., 2019) from cratonic
34 regions such as the Baltic and Canadian Shields (Cook et al. 2010; Artemieva and Thybo,
35 2013; Fig. 1B) to Proterozoic and Paleozoic terranes such as Antarctica and Australia
36 (Salmon et al., 2012; An et al., 2015; Ebbing et al., 2018; Fig. 1C). Some of these crustal
37 roots have been interpreted as inherited regions of thick orogenic crust (e.g., Fischer,
38 2002; Studinger et al., 2004), others as mantle-derived mafic roots accreted below a
39 continental crust of normal thickness (e.g., Thybo and Artemieva, 2013). In both cases,
40 we expect that the enhanced heat flow would thermally weaken the deep crust, enabling
41 efficient viscous flow to relax gradients of crustal thickness and to flatten the Moho on a
42 regional scale (Clark and Royden, 2000; Beaumont et al., 2001; Nábělek et al., 2009; Rey
43 et al., 2010). Hence, the persistence over hundreds of millions of years of thick crustal
44 roots poses an intriguing problem. Although high heat flow produces migmatites and
45 granites that contribute to the transient weakening of the deep continental crust, it also
46 produces drier and stronger garnet-pyroxene rocks such as granulites (e.g., Jackson et al.,
47 2004). Upon cooling, hydration, and deformation, these stronger rocks may be

48 retromorphosed into weaker amphibolitic gneisses. Here, we explore through two-
49 dimensional (2-D) thermomechanical experiments how the interplay between mechanical
50 weakening due to partial melting, strengthening and density increase due to the
51 crystallization of garnet-pyroxene assemblages, and post-orogenic weakening due to
52 retrogression may impact the long-term crustal thickness. Our results suggest that thick
53 crustal root anomalies could be the remnants of dry garnet-pyroxene-bearing rocks that
54 survived post-orogenic extension and retrogression. These garnet-pyroxene-bearing
55 crustal-scale boudins strengthen the lower crust and reduce its capacity to flow. Our
56 experiments are a first step toward explaining why relaxed orogenic crust may maintain
57 heterogeneities in crustal thickness hundreds of millions of years after orogeny has
58 ceased.

59 **NUMERICAL EXPERIMENTS, CODE, AND MODEL SETUP**

60 Our 2-D thermomechanical experiments consider a 360-km-wide orogenic plateau
61 with a 70-km-thick crust (i.e., the thickness of the Tibetan Plateau; Nábělek et al., 2009)
62 above 40 km of mantle (Fig. 2). A layer of air-like material with low viscosity and low
63 density is imposed on top of the crust to accommodate the development of surface
64 topography. The plateau experiences extensional deformation as the crust returns to a
65 normal thermomechanical state. Extensional-velocity boundary conditions are imposed
66 on both vertical walls of the model. We have tested slow (0.18 cm yr^{-1}) and fast (1.8 cm
67 yr^{-1}) velocities, delivering a strain rate averaged over the length of the model of 3×10^{-16}
68 s^{-1} and $3 \times 10^{-15} \text{ s}^{-1}$ respectively. Horizontal boundaries of the model are free slips. The
69 thermal properties of the material combined with constant basal heat flow and constant
70 top temperature deliver an initial steady-state geotherm leading to a Moho temperature of

71 ~900 °C (Fig. 2). We select from the literature plausible visco-plastic parameters (see the
72 GSA Data [Repository](#)¹) so the mechanical behavior of the modeled lithosphere depends
73 on temperature, strain rate, deviatoric stress, and accumulated strain. Details of modeling
74 procedures, rheological and thermal parameters, as well as the input Python script are
75 available in the Data Repository.

76 In order to explore the interplay between partial melting, the formation of stronger
77 garnet-pyroxene-bearing rocks, and their retrogression into weaker amphibolite facies
78 rocks, we parameterize three first-order metamorphic phase transitions. The first phase
79 change simulates partial melting and its feedback on density, viscosity, and temperature
80 (Rey et al., 2009; see the Data Repository). A second phase change with feedback on
81 density and viscosity occurs at temperature $T = 777$ °C to simulate prograde amphibolite
82 to garnet-pyroxene rock reaction (Philpotts and Ague, 2009). Finally, a third phase
83 change with feedback on density and viscosity accounts for the retrogression of garnet-
84 pyroxene-bearing rocks back into amphibolite facies rocks. This third phase change
85 occurs at $T = 777$ °C as well and for a strain rate $\geq 10^{-14}$ s⁻¹. Our model implicitly
86 assumes that water is available. Therefore, retrogression is contingent upon strain rate,
87 which simulates the metastability of dry high-grade rocks during exhumation. This strain-
88 rate threshold is in the range of expected strain rates measured in orogenic shear zones
89 (Sassier et al., 2009; Boutonnet et al., 2013; Fagereng and Biggs, 2019). Rock solidus
90 depends on rock fertility and availability of fluid. Hence, we have tested different solidii
91 for the continental crust and the garnet-pyroxene-bearing rocks (Data Repository) in the
92 range commonly accepted for these rock types. For the continental crust, we have tested a
93 solidus representative of fertile metapelites with a melting temperature at room pressure

94 of 650 °C (Figs. 3A and 3B; White et al., 2001), and a solidus representative of less-
95 fertile rocks with a melting temperature at room pressure of 720 °C (Fig. 3C; Rey and
96 Müller, 2010). For the dry garnet-pyroxene-bearing crust, we use a melting temperature
97 at room pressure of 790 °C representative of refractory granulites (Cenki-Tok et al.,
98 2016). We use *Underworld*, a well-tested open-source finite-element code
99 (<https://underworld2.readthedocs.io/>), to solve the equations of conservation of
100 momentum, mass, and energy for an incompressible fluid on a Cartesian Eulerian mesh
101 (Moresi et al., 2007; Beucher et al., 2019).

102 **RESULTS**

103 When a slow divergent velocity is imposed (0.18 cm yr^{-1}), the crust thins
104 homogeneously, the Moho remains flat, and deformation is dominated by pure shear
105 strain whether melt and/or garnet-pyroxene rocks are present or not (Fig. DR1 in the Data
106 Repository). In contrast, under faster extensional velocities (1.8 cm yr^{-1}), the
107 experimental outcome depends on phase changes. When the formation of strong garnet-
108 pyroxene rocks is not allowed, partial melting makes the deep crust hot and mobile,
109 which allows the formation of a migmatitic dome (Fig. 3A). In the partially molten dome,
110 finite strain ellipses are strongly flattened, with a vertical long axis indicating the
111 presence of a vertical high-strain zone separating two sub-domes. This double-dome
112 geometry has been well documented (Rey et al., 2011, 2017; Korchinski et al., 2018).
113 Figures 3B and 3C show a different result when prograde garnet-pyroxene rock formation
114 and retrogression into amphibolite are allowed. In the case where retrogression does not
115 occur (Fig. DR2A), the crust thins homogeneously. As the formation of garnet-pyroxene
116 rocks strengthens the deep crust, its capacity to flow is much reduced and the upper crust

117 remains mechanically coupled to the mantle. Extensional deformation is more distributed
118 and heterogeneous as documented by the crustal-scale pinch-and-swell strain pattern, as
119 well as the finite strain field imaged by the finite strain ellipses (Fig. 3B). As strain rate
120 controls the retrogression of garnet-pyroxene rocks (Figs. DR2B and DR2C), we observe
121 that retrogression is partitioned into the pinch regions where strain rate is higher, whereas
122 garnet-pyroxene pods are preferentially preserved in the swell regions where strain rate is
123 lower and below the threshold required to activate retrogression. Because retrogression
124 leads to weakening, favoring strain localization and therefore higher strain rates, there is
125 a positive feedback loop between strain rate, retrogression, and weakening. When the
126 crustal solidus is that of a fertile metapelite, portions of the lower crust are partially
127 molten and able to flow under gravitational stresses, whereas flow is inhibited in the
128 strong garnet-pyroxene rock pods (Fig. 3D). Raising the solidus temperature of the
129 continental crust by 70 °C results in a similar outcome except that there are no more
130 partially molten domains within the continental crust (Fig. 3C). Because of the formation
131 of garnet-pyroxene rock pods, the Moho presents a winding geometry with crustal
132 thickness variations of as much as 50%, from 35 to 53 km. After 25% of extension and
133 thinning, we have left these experiments to thermally and mechanically relax over 180
134 m.y. under fixed boundary conditions (i.e., setting the kinematic boundary condition to 0
135 cm yr^{-1}). We observe that the heterogeneity of crustal thickness persists throughout this
136 long cooling history.

137 **DISCUSSION**

138 Our numerical experiments suggest that strain rate-dependent retrogression that
139 typically localizes along ductile shear zones cutting through high-grade rocks may

140 explain how remnants of thick and strong orogenic crust can survive orogenic collapse.
141 These regions can be ~50% thicker than the adjacent crust and as narrow as a few tens of
142 kilometers across, and survive for hundreds of millions of years. Anomalous deep crustal
143 roots have been imaged in stable intracontinental regions all around the globe. In the
144 eastern Canadian Shield, for example, the Lithoprobe project
145 (<https://lithoprobe.eos.ubc.ca/>) has documented several crustal roots (Cook et al., 2010).
146 Below the Torngat orogen along the eastern Canadian Shield, a Paleoproterozoic crustal
147 root as much as 50 km deep, 15 km deeper than the average adjacent crust, and ~80 km
148 wide and >200 km long has been imaged on seismic profiles (Fig. 1B; Funck and
149 Louden, 1999). It is interesting to note that this Paleoproterozoic crustal root is bounded
150 to the north and east by major shear zones (Cook et al., 2010). In the Baltic Shield, along
151 an Archean–Paleoproterozoic suture, the Moho reaches a depth of ~60 km over a region
152 centered on southern Finland (Artemieva and Thybo, 2013). In central Australia, even
153 though this continent has been tectonically relatively stable for the past 300 m.y., crustal
154 roots reaching 65 km depth have been imaged as deep regions of diffuse reflectivity over
155 circular domains a few hundred kilometers in diameter (Fig. 1C; Kennett et al., 2011;
156 Salmon et al., 2012). In Antarctica, a series of crustal roots as much as 60 km deep have
157 been documented between Dronning Maud Land and Gamburtsev Subglacial Mountains
158 (An et al., 2015; Ebbing et al., 2018). In peninsular India, made up of Archean to
159 Paleozoic terranes, the Moho depth varies from ~38 km below the southernmost tip of
160 India’s Proterozoic Southern Granulite terrane, to 50 km below the Archean Dharwar
161 craton in semicircular regions ~250 km in diameter (Reddy and Vijaya Rao, 2013; Das et
162 al., 2019). The structure and nature of the lower crust below the Southern Granulite

163 terrane is heterogeneous, but because the middle and upper crust shows a constant
164 thickness of 20–25 km, this variability must be accommodated by variation in thickness
165 of the lower crust (18–32 km; Das et al., 2019).

166 The gravimetric and seismic characteristics of these crustal roots suggest the
167 presence of garnet-pyroxene-bearing rocks. For example, in Canada, crustal roots
168 showing P-wave velocities $>7 \text{ km s}^{-1}$ led Cook et al. (2010) to propose that in the
169 absence of later tectonic reworking, the variations in Moho depth originate solely from
170 rheological variations. In southern India, crustal roots display compressional-wave
171 velocities that are systematically $>7 \text{ km s}^{-1}$ (Reddy and Vijaya Rao, 2013), and shear-
172 wave velocities between 4 and 4.2 km s^{-1} (Das et al., 2019). The contrasting density and
173 seismic characteristics between granitic rocks and/or amphibolite facies gneisses (<2700
174 kg m^{-3} and $<6.4 \text{ km s}^{-1}$) and higher-grade garnet-pyroxene-bearing rocks ($>2800 \text{ kg m}^{-3}$
175 and $>6.6 \text{ km s}^{-1}$; Christensen and Mooney, 1995; Artemieva and Thybo, 2013) suggest
176 that deep crustal roots are made of the latter (Williams et al., 2014). This proposition is
177 compatible with the seismically diffuse boundary that is commonly observed between the
178 lower crust and the mantle (O'Reilly and Griffin, 2013). Because the petrophysical
179 properties of garnet-pyroxene-bearing rocks are intermediate between the ones of the
180 crust and the mantle, a garnet-pyroxene-rich lower crust would explain the seismic
181 properties of the transition between the crust and the mantle observed in Peninsular India
182 for example (Reddy and Vijaya Rao, 2013).

183 Importantly, in all of these examples, crustal roots are interpreted as inherited
184 remnants of ancient orogenic crust that have survived gravitational collapse and the
185 flattening of the Moho. We propose that these strong orogenic crustal roots owe their

186 survival to the presence of retrogressed and therefore weaker adjacent crusts in which
187 deformation is strongly partitioned. The positive feedback loop between strain,
188 retrogression, and weakening insures that deformation remains localized into retrogressed
189 domains, isolating and protecting garnet-pyroxene-bearing pods that remain largely
190 immune to deformation.

191 **CONCLUSIONS**

192 In this study, we have explored through 2-D thermomechanical modeling how the
193 interplay between partial melting, the formation of garnet-pyroxene high-grade rocks, and
194 strain rate–dependent retrogression could explain the long-term preservation of deep
195 crustal roots in stable continents. Though 2-D experiments are sufficient to illustrate how
196 strain rate, retrogression, and weakening can explain the preservation of thick roots,
197 future work involving 3-D experiments will allow investigation of triclinic boundary
198 conditions. Our experiments show that following the formation of high-grade rocks in
199 deep orogenic crusts, extension is partitioned into regions where strain, retrogression, and
200 weakening are coupled into a positive feedback loop. This results in the preservation of
201 thick, dense, and strong garnet-pyroxene-rich pods, separated by retrogressed and
202 attenuated pinched regions. The strong high-grade pods form crustal-scale boudins that
203 are able to survive through the orogenic relaxation phase and over a duration of >100
204 m.y. As a result, the equilibrated orogenic crust preserves deep crustal roots similar to
205 those documented in all stable continents. These results are first steps toward
206 understanding of the feedback between metamorphic reactions and deformation. In the
207 future, 3-D models involving porous flow and surface processes will allow a more
208 detailed understanding of these systems.

209 **ACKNOWLEDGMENTS**

210 We acknowledge funding from the European Union's Horizon 2020 research and
211 innovation program under grant agreement 793978. This research was undertaken with
212 the assistance of resources from the National Computational Infrastructure (NCI),
213 through the National Computational Merit Allocation Scheme supported by the
214 Australian Government; the Pawsey Supercomputing Centre (Perth, Australia) with
215 funding from the Australian Government and the Government of Western Australia, and
216 support from the Australian Research Council through the Industrial Transformation
217 Research Hub grant ARC-IH130200012. We thank Julian Giordani and Romain Beucher
218 for their expert support with *Underworld* (<https://underworld2.readthedocs.io/>). We thank
219 Chris Clark for editorial handling, as well as Gregory Dumond and two anonymous
220 reviewers for their constructive reviews.

221 **REFERENCES CITED**

- 222 An, M., Wiens, D.A., Yue, Z., Mei, F., Nyblade, A.A., Kanao, M., Li, Y., Maggi, A., and
223 Lévêque, J.-J., 2015, *S*-velocity model and inferred Moho topography beneath the
224 Antarctic Plate from Rayleigh waves: *Journal of Geophysical Research: Solid Earth*,
225 v. 120, p. 359–383, <https://doi.org/10.1002/2014JB011332>.
- 226 Artemieva, I.M., and Thybo, H., 2013, EUNAsis: A seismic model for Moho and crustal
227 structure in Europe, Greenland, and the North Atlantic region: *Tectonophysics*,
228 v. 609, p. 97–153, <https://doi.org/10.1016/j.tecto.2013.08.004>.
- 229 Beaumont, C., Jamieson, R.A., Nguyen, M.H., and Lee, B., 2001, Himalayan tectonics
230 explained by extrusion of a low-viscosity crustal channel coupled to focused surface
231 denudation: *Nature*, v. 414, p. 738–742, <https://doi.org/10.1038/414738a>.

- 232 Beucher, R., et al., 2019, UWGeodynamics: A teaching and research tool for numerical
233 geodynamic modelling: *Journal of Open Source Software*, v. 4, 1136,
234 <https://doi.org/10.21105/joss.01136>.
- 235 Boutonnet, E., Leloup, P.H., Sassier, C., Gardien, V., and Ricard, Y., 2013, Ductile strain
236 rate measurements document long-term strain localization in the continental crust:
237 *Geology*, v. 41, p. 819–822, <https://doi.org/10.1130/G33723.1>.
- 238 Cenko-Tok, B., Berger, A., and Gueydan, F., 2016, Formation and preservation of biotite-
239 rich microdomains in high-temperature rocks from the Antananarivo Block,
240 Madagascar: *International Journal of Earth Sciences*, v. 105, p. 1471–1483,
241 <https://doi.org/10.1007/s00531-015-1265-0>.
- 242 Christensen, N.I., and Mooney, W.D., 1995, Seismic velocity structure and composition
243 of the continental crust: A global view: *Journal of Geophysical Research*, v. 100,
244 p. 9761–9788, <https://doi.org/10.1029/95JB00259>.
- 245 Clark, M.K., and Royden, L.H., 2000, Topographic ooze: Building the eastern margin of
246 Tibet by lower crustal flow: *Geology*, v. 28, p. 703–706,
247 [https://doi.org/10.1130/0091-7613\(2000\)28<703:TOBTEM>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<703:TOBTEM>2.0.CO;2).
- 248 Cook, F.A., White, D.J., Jones, A.G., Eaton, D.W.S., Hall, J., and Clowes, R.M., 2010,
249 How the crust meets the mantle: Lithoprobe perspectives on the Mohorovic
250 discontinuity and crust–mantle transition: *Canadian Journal of Earth Sciences*, v. 47,
251 p. 315–351, <https://doi.org/10.1139/E09-076>.
- 252 Das, R., Ashish, and Saha, G.K., 2019, Crust and shallow mantle structure of south India
253 by inverting interpolated receiver function with surface wave dispersion: *Journal of*

- 254 Asian Earth Sciences, v. 176, p. 157–167,
255 <https://doi.org/10.1016/j.jseaes.2019.02.011>.
- 256 Ebbing, J., Hass, P., Ferraccioli, F., Pappa, F., Szwillus, W., and Bouman, J., 2018, Earth
257 tectonics as seen by GOCE—Enhanced satellite gravity gradient imaging: Scientific
258 Reports, v. 8, <https://doi.org/10.1038/s41598-018-34733-9>.
- 259 Fagereng, Å., and Biggs, J., 2019, New perspectives on ‘geological strain rates’
260 calculated from both naturally deformed and actively deforming rocks: Journal of
261 Structural Geology, v. 125, p. 100–110, <https://doi.org/10.1016/j.jsg.2018.10.004>.
- 262 Fischer, K.M., 2002, Waning buoyancy in the crustal roots of old mountains: Nature,
263 v. 417, p. 933–936, <https://doi.org/10.1038/nature00855>.
- 264 Funck, T., and Loudon, K.E., 1999, Wide-angle seismic transect across the Torngat
265 Orogen, northern Labrador: Evidence for a Proterozoic crustal root: Journal of
266 Geophysical Research, v. 104, p. 7463–7480, <https://doi.org/10.1029/1999JB900010>.
- 267 Jackson, J.A., Austrheim, H., McKenzie, D., and Priestley, K., 2004, Metastability,
268 mechanical strength, and the support of mountain belts: Geology, v. 32, p. 625–628,
269 <https://doi.org/10.1130/G20397.1>.
- 270 Kennett, B.L.N., Salmon, M., Saygin, E. and the AusMoho Working Group, 2011,
271 AusMoho: The variation of Moho depth in Australia: Geophysical Journal
272 International, v. 187, p. 946–958, <https://doi.org/10.1111/j.1365-246X.2011.05194.x>.
- 273 Korchinski, M., Rey, P.F., Mondy, L., Teyssier, C., and Whitney, D.L., 2018, Numerical
274 investigation of deep-crust behavior under lithospheric extension: Tectonophysics,
275 v. 726, p. 137–146, <https://doi.org/10.1016/j.tecto.2017.12.029>.

- 276 Moresi, L., Quenette, S., Lemiale, V., Meriaux, C., Appelbe, B., and Mühlhaus, H.-B.,
277 2007, Computational approaches to studying non-linear dynamics of the crust and
278 mantle: *Physics of the Earth and Planetary Interiors*, v. 163, p. 69–82,
279 <https://doi.org/10.1016/j.pepi.2007.06.009>.
- 280 Nábělek, J., Hetényi, G., Vergne, J., Sapkota, S., Kafle, B., Jiang, M., Su, H.P., Chen, J.,
281 Huang, B.S., and the Hi-CLIMB Team, 2009, Underplating in the Himalaya-Tibet
282 collision zone revealed by the Hi-CLIMB Experiment: *Science*, v. 325, p. 1371–
283 1374, <https://doi.org/10.1126/science.1167719>.
- 284 O'Reilly, S.Y., and Griffin, W.L., 2013, Moho vs crust-mantle boundary: Evolution of an
285 idea: *Tectonophysics*, v. 609, p. 535–546,
286 <https://doi.org/10.1016/j.tecto.2012.12.031>.
- 287 Philpotts, A.R., and Ague, J.J., 2009, *Principles of Igneous and Metamorphic Petrology*:
288 Cambridge, UK, Cambridge University Press, 667 p.,
289 <https://doi.org/10.1017/CBO9780511813429>.
- 290 Prodehl, C., Kennett, B., Artemieva, I.M., and Thybo, H., 2013, 100 years of seismic
291 research on the Moho: *Tectonophysics*, v. 609, p. 9–44,
292 <https://doi.org/10.1016/j.tecto.2013.05.036>.
- 293 Reddy, P.R., and Vijaya Rao, V., 2013, Seismic images of the continental Moho of the
294 Indian shield: *Tectonophysics*, v. 609, p. 217–233,
295 <https://doi.org/10.1016/j.tecto.2012.11.022>.
- 296 Rey, P.F., and Müller, R.D., 2010, Fragmentation of active continental plate margins
297 owing to the buoyancy of the mantle wedge: *Nature Geoscience*, v. 3, p. 257–261,
298 <https://doi.org/10.1038/ngeo825>.

- 299 Rey, P.F., Teyssier, C., and Whitney, D.L., 2009, Extension rates, crustal melting, and
300 core complex dynamics: *Geology*, v. 37, p. 391–394,
301 <https://doi.org/10.1130/G25460A.1>.
- 302 Rey, P.F., Teyssier, C., and Whitney, D.L., 2010, The limit of channel flow in orogenic
303 plateaux: *Lithosphere*, v. 2, p. 328–332, <https://doi.org/10.1130/L114.1>.
- 304 Rey, P.F., Teyssier, C., Kruckenberg, S.C., and Whitney, D.L., 2011, Viscous collision in
305 channel explains double domes in metamorphic core complexes: *Geology*, v. 39,
306 p. 387–390, <https://doi.org/10.1130/G31587.1>.
- 307 Rey, P.F., Mondy, L., Duclaux, G., Teyssier, C., Whitney, D.L., Bocher, M., and Prigent,
308 C., 2017, The origin of contractional structures in extensional gneiss domes:
309 *Geology*, v. 45, p. 263–266, <https://doi.org/10.1130/G38595.1>.
- 310 Salmon, M., Kennett, B.L.N., Stern, T., and Aitken, A.R.A., 2012, The Moho in Australia
311 and New Zealand: *Tectonophysics*, v. 609, p. 288–298,
312 <https://doi.org/10.1016/j.tecto.2012.07.009>.
- 313 Sassier, C., Leloup, P.H., Rubatto, D., Galland, O., Yue, Y., and Lin, D., 2009, Direct
314 measurement of strain rates in ductile shear zones: A new method based on
315 syntectonic dikes: *Journal of Geophysical Research*, v. 114, B01406,
316 <https://doi.org/10.1029/2008JB005597>.
- 317 Studinger, M., Bell, R.E., Buck, W.G., Karner, G.D., and Blankenship, D.D., 2004, Sub-
318 ice geology inland of the Transantarctic Mountains in light of new aerogeophysical
319 data: *Earth and Planetary Science Letters*, v. 220, p. 391–408,
320 [https://doi.org/10.1016/S0012-821X\(04\)00066-4](https://doi.org/10.1016/S0012-821X(04)00066-4).

- 321 Szwillus, W., Afonso, J.C., Ebbing, J., and Mooney, W.D., 2019, Global crustal thickness
322 and velocity structure from geostatistical analysis of seismic data: Journal of
323 Geophysical Research: Solid Earth, v. 124, p. 1626–1652,
324 <https://doi.org/10.1029/2018JB016593>.
- 325 Thybo, H., and Artemieva, I.M., 2013, Moho and magmatic underplating in continental
326 lithosphere: Tectonophysics, v. 609, p. 605–619,
327 <https://doi.org/10.1016/j.tecto.2013.05.032>.
- 328 White, R.W., Powell, R., and Holland, T.J.B., 2001, Calculation of partial melting
329 equilibria in the system Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O
330 (NCKFMASH): Journal of Metamorphic Geology, v. 19, p. 139–153,
331 <https://doi.org/10.1046/j.0263-4929.2000.00303.x>.
- 332 Williams, M.L., Dumond, G., Mahan, K., Regan, S., and Holland, M., 2014, Garnet-
333 forming reactions in felsic orthogneiss: Implications for densification and
334 strengthening of the lower continental crust: Earth and Planetary Science Letters,
335 v. 405, p. 207–219, <https://doi.org/10.1016/j.epsl.2014.08.030>.

336 **FIGURE CAPTIONS**

337 Figure 1. (A) Crustal thickness histogram for world shields extracted from CRUST 5.1
338 model (modified from <https://earthquake.usgs.gov/data/crust/crust.php>). (B) Interpolated
339 compressional-wave velocities across the Torngat orogen, northeastern Canada (modified
340 from Funck and Loudon, 1999). (C) Interpolated Moho surfaces for Australia constructed
341 by interpolating weighted averages for each $0.5^\circ \times 0.5^\circ$ pixel (modified from Kennett et
342 al., 2011).

343

344 Figure 2. Model geometry and initial conditions, as well as geotherm, viscosity, and
345 density profiles. Weak prismatic region dipping 45° simulates detachment fault in upper
346 crust. Circles pattern superimposed on continental crust represents finite-strain ellipses.

347

348 Figure 3. Fast-velocity modeling results (1.8 cm yr^{-1} extension speed) at average strain
349 rate of $3 \times 10^{-15} \text{ s}^{-1}$ and 25% extension. Colors are the same as in Figure 2. (Model A)

350 Only partial melting is allowed (garnet-pyroxene isograd and retrogression into
351 amphibolite are removed). (Model B) Partial melting, crystallization of garnet-pyroxene
352 assemblages, and retrogression are allowed. Temperature for transformation of
353 continental crust into garnet-pyroxene-rich rocks is 777°C (see text for explanation).

354 Reference temperatures for solidus of continental crust and garnet-pyroxene-rich crust are
355 650°C and 790°C , respectively. (Model C) Same as model B but temperature for
356 continental crust solidus is increased to 720°C . (Model D) Zoom on model B illustrating
357 velocity field (black arrows) when boundary condition mimicking extension is removed
358 (after 2 m.y. of gravity forces operating), showing that partially molten crust flows while
359 garnet-pyroxene-rich rocks do not.

360

Figure 1

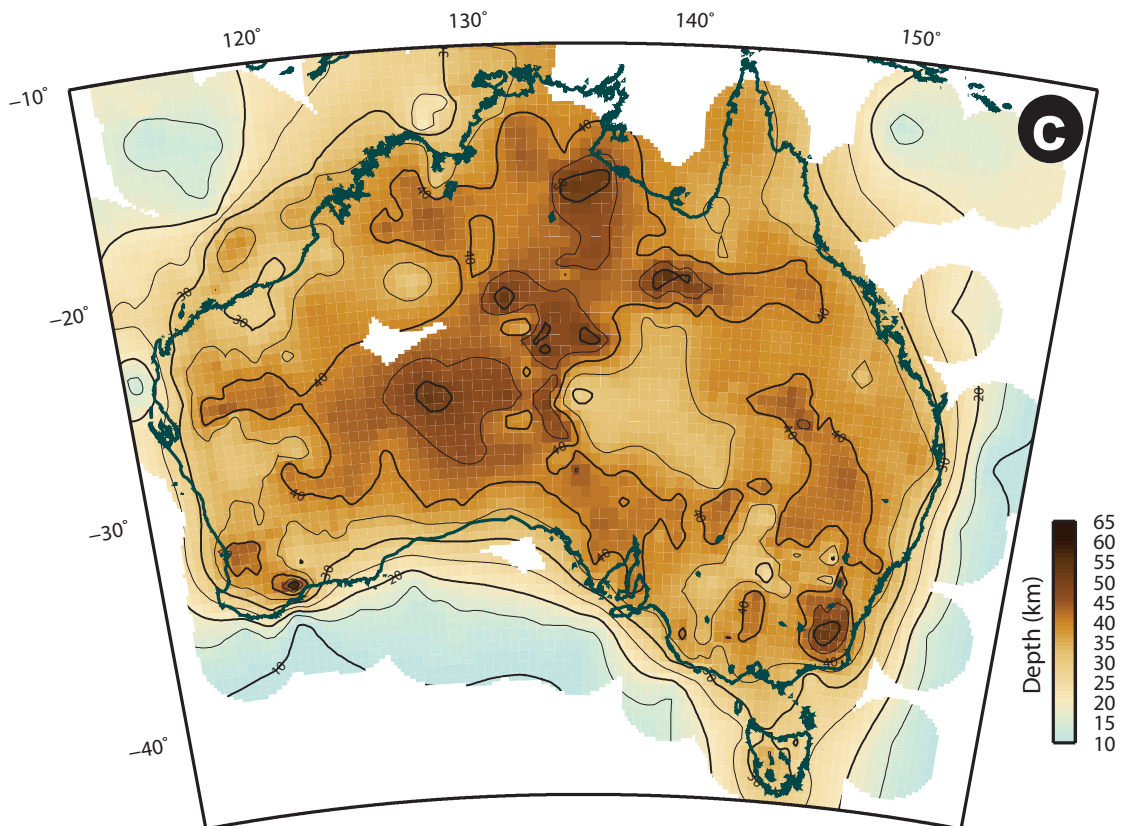
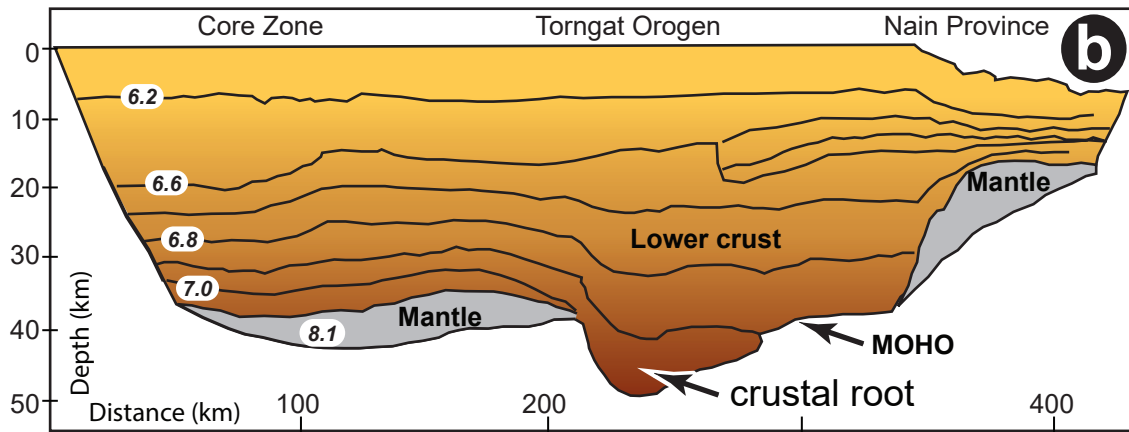
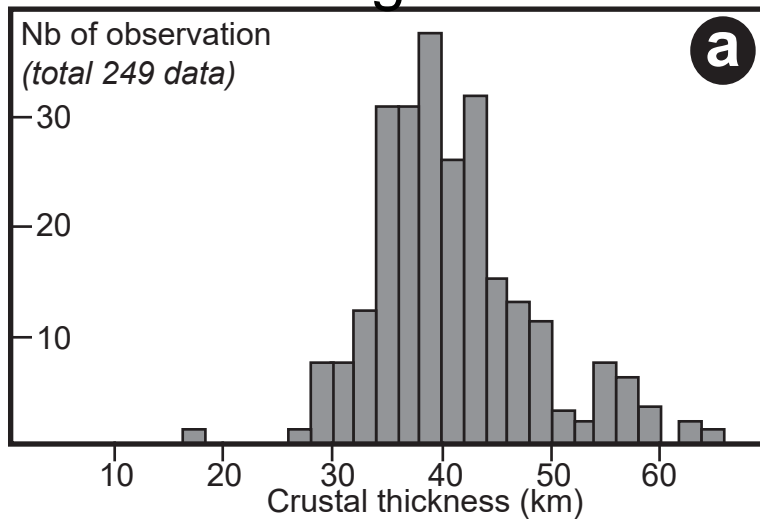


Figure 1

Figure 2

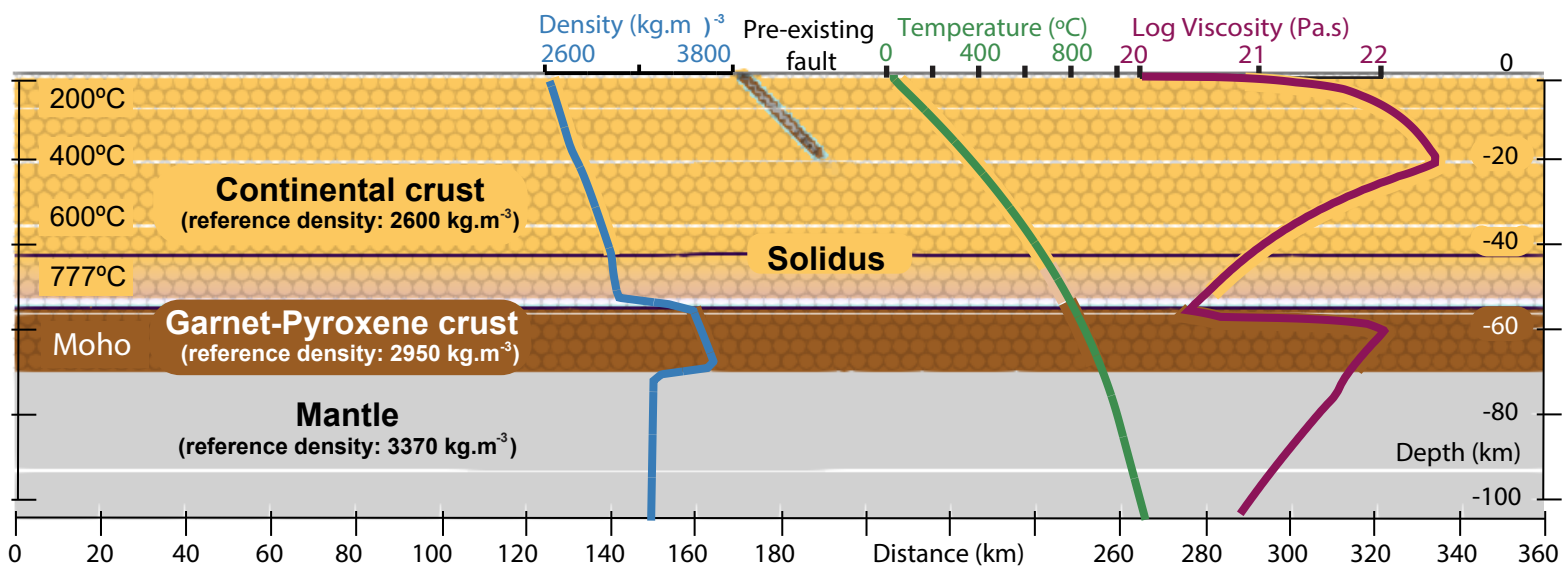


Figure 2

Figure 3

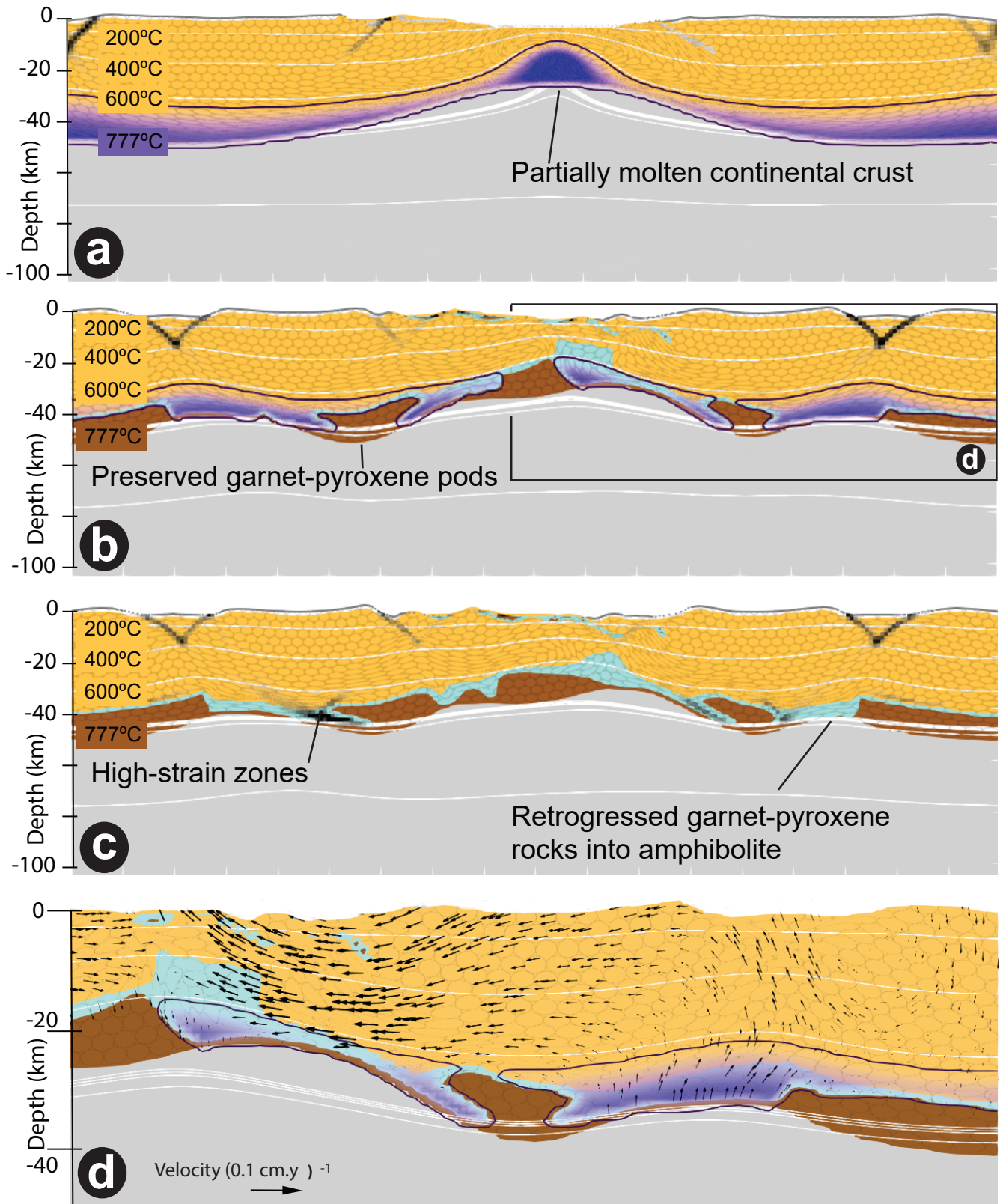


Figure 3

1 Strain and retrogression partitioning explain long-term stability of crustal roots in stable continents

2 Bénédicte Cenki-Tok^{1,2}, P.F. Rey² and D. Arcay¹

3
4 TABLE DR1. THERMAL AND MECHANICAL PARAMETERS

| Parameter | Continental Crust | Retrogressed Amphibolites | Garnet-Pyroxene CC | Upper Mantle | Fault |
|--|----------------------------|----------------------------|-----------------------------------|-------------------------|---------------------------------|
| Reference temperature (K) | 293 | 293 | | | |
| Dislocation creep viscous rheology | Wet quartzite ^a | Wet quartzite ^a | Dry Maryland Diabase ^b | Wet dunite ^c | 0.1 *Wet quartzite ^a |
| Reference density (kg·m ⁻³) | 2600 | 2600 | 2950 | 3370 | 2600 |
| Thermal expansivity (K ⁻¹) | -1.00E-04 | -1.00E-04 | -1.00E-04 | 2.80E-05 | -1.00E-04 |
| Compressibility (Pa ⁻¹) | 8.00E-11 | 8.00E-11 | - | - | 8.00E-11 |
| Heat capacity (J K ⁻¹ kg ⁻¹) | 1000 | 1000 | 1000 | 1000 | 1000 |
| Thermal diffusivity (m ² s ⁻¹) | 9E-07 | 9E-07 | 9E-07 | 9E-07 | 9E-07 |
| Latent heat of fusion (kJ kg ⁻¹ K ⁻¹) | 250 | 250 | 250 | - | 250 |
| Radiogenic heat production (W m ⁻³) ^d | 5.00E-07 | 5.00E-07 | 5.00E-07 | - | 5.00E-07 |
| Melt fraction density change ^e | 0.13 | 0.13 | - | - | 0.13 |
| Solidus term 1 (K) | 923 | 923 | 1063 | - | 923 |
| Solidus term 2 (K Pa ⁻¹) | -1.20E-07 | -1.20E-07 | -1.20E-07 | - | -1.20E-07 |
| Solidus term 3 (K Pa ⁻²) | 1.20E-16 | 1.20E-16 | 1.20E-16 | - | 1.20E-16 |
| Liquidus term 1 (K) | 1423 | 1423 | 1563 | - | 1423 |
| Liquidus term 2 (K Pa ⁻¹) | -1.20E-07 | -1.20E-07 | -1.20E-07 | - | -1.20E-07 |
| Liquidus term 3 (K Pa ⁻²) | 1.60E-16 | 1.60E-16 | 1.60E-16 | - | 1.60E-16 |
| Friction coefficient | 0.44 | 0.44 | 0.44 | 0.44 | 0.44 |
| Softened friction coefficient | 0.088 | 0.088 | 0.088 | 0.088 | 0.088 |
| Cohesion (MPa) | 15 | 15 | 15 | 15 | 1.5 |
| Softened cohesion (MPa) | 3 | 3 | 3 | 3 | 0.3 |
| Pre-exponential factor (MPa ⁻ⁿ s ⁻¹) | 5.00E-06 | 5.00E-06 | 5.05E-22 | 70000 | 5.00E-06 |
| Stress exponent (n) | 3 | 3 | 4.7 | 3 | 3 |
| Activation energy (kJ mol ⁻¹) | 190 | 190 | 485 | 520 | 190 |
| Activation volume (m ³ mol ⁻¹) | 0 | 0 | 0 | 0 | 0 |
| Water fugacity | 0 | 0 | 0 | 0 | 0 |
| Water fugacity exponent ^f | 0 | 0 | 0 | 0 | 0 |
| Melt viscous softening factor | 1.00E-03 | 1.00E-03 | | | 1.00E-03 |
| Softening melt fraction interval | 0.2-0.3 | 0.2-0.3 | | | 0.2-0.3 |

Additional parameters:

Model Size: 360 km length (241 nodes, constant spacing) - 120 km thick (81 nodes, constant spacing) i.e. 15 km air-like material, 70 km crust, 35 km upper mantle. The marker density is uniform (60 per grid cell).

A weak prismatic region dipping 45 °C simulates a detachment in the upper crust

Basal heat flow is set at 0.015 W.m⁻²

Velocities tested: 1.8 cm.y⁻¹ (fast) or 0.18 cm.y⁻¹ (slow) and Isostasy is activated

Prograde amphibolite to garnet-pyroxene rock phase change set at 1050 K

Retrograde garnet-pyroxene rock to amphibolite phase change set at 1050 K and 10⁻¹⁴ s⁻¹ strain rate

Moho temperature at the start of the model is 883 °C

Solidus and liquidus are defined by a polynomial function of pressure (P):

$$T_s = a_0 + a_1 \times P + a_2 \times P^2, T_l = b_0 + b_1 \times P + b_2 \times P^2$$

The density of the continental crust changes according to T and P:

$$\rho = \rho_0 * (1 + (\beta * \Delta P) - (\alpha * \Delta T))$$

Note that the presence of melt has an impact on density.

The maximum melt fraction is 30%.

References:

a Parameters were derived from Brace and Kohlstedt (1980)

b Parameters were derived from Mackwell et al (1998)

c Parameters were derived from Brace and Kohlstedt (1980)

d Parameters were derived from Hasterok and Chapman (2011)

e Melt and other parameters were derived from Rey and Muller (2010)

f A zero value denotes that this effect on the viscous flow law is incorporated into the pre-exponential factor

5 DR Figure1: Slow modeling results (0.18 cm.y^{-1} extension speed) at an average strain rate of
6 $3\text{e-}16 \text{ s}^{-1}$ and 25 % extension. Colors are the same as in Fig. 2. Conditions for a, b and c
7 models are the same as in Fig. 3.

8 DR Figure2: Fast modeling results (1.8 cm.y^{-1} extension speed) at an average strain rate of
9 $3\text{e-}15 \text{ s}^{-1}$ and 25 % extension. Colors are the same as in Fig 3. Variations of model shown on
10 Figure 3b but a. only the prograde phase change is allowed. Retrogression does not occur,
11 and the crust, including the garnet-pyroxene layer, thins homogeneously. b. The strain rate
12 threshold for retrogression has been set to 10^{-13} s^{-1} . Retrogression does not occur, and the
13 crust, including the garnet-pyroxene layer, thins homogeneously. c. The strain rate threshold
14 for retrogression has been set to 10^{-15} s^{-1} , large portions of retrogressed and partially molten
15 crust are exhumed in the center of the model whereas the garnet-pyroxene layer is preserved
16 locally but strongly thinned.

17

18 REFERENCES CITED:

19 Brace, W.F., and Kohlstedt, D.L., 1980, Limits on lithospheric stress imposed by laboratory
20 experiments: *Journal of Geophysical Research*, v. 85, p. 6248-6252.

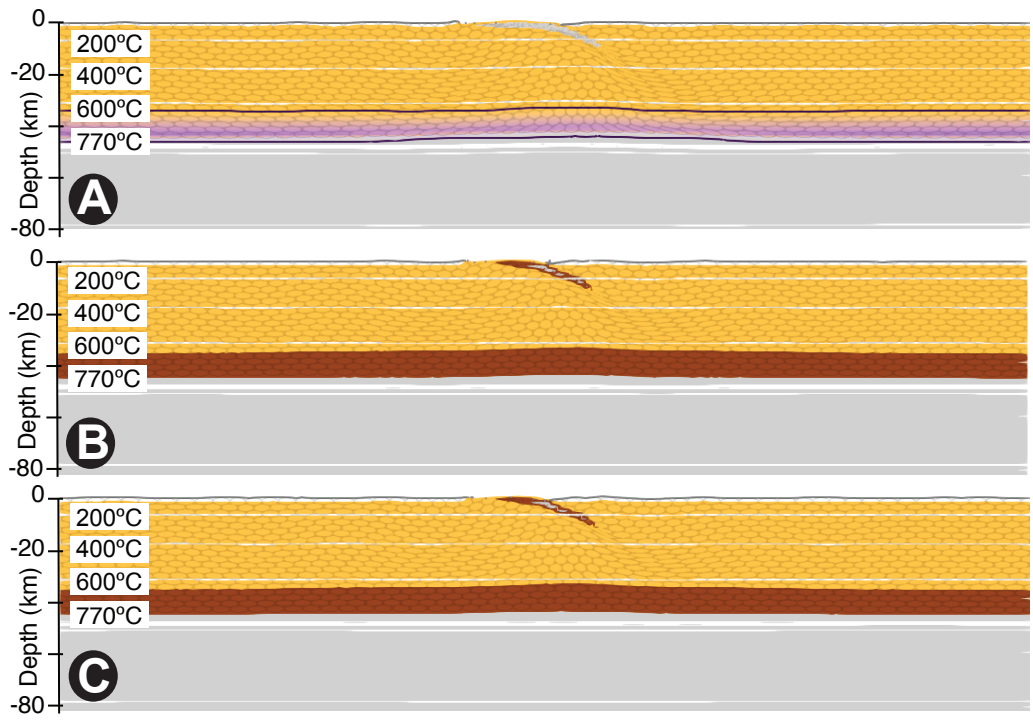
21 Hasterok, D., Chapman, D., 2011. Heat production and geotherms for the continental
22 lithosphere. *Earth Planet. Sci. Lett.* 307 (1), 5970.

23 Mackwell, S. J., Zimmerman, M. E., & Kohlstedt, D. L. (1998). High-temperature
24 deformation of dry diabase with application to tectonics on Venus. *Journal of Geophysical*
25 *Research: Solid Earth*, 103(B1), 975-984.

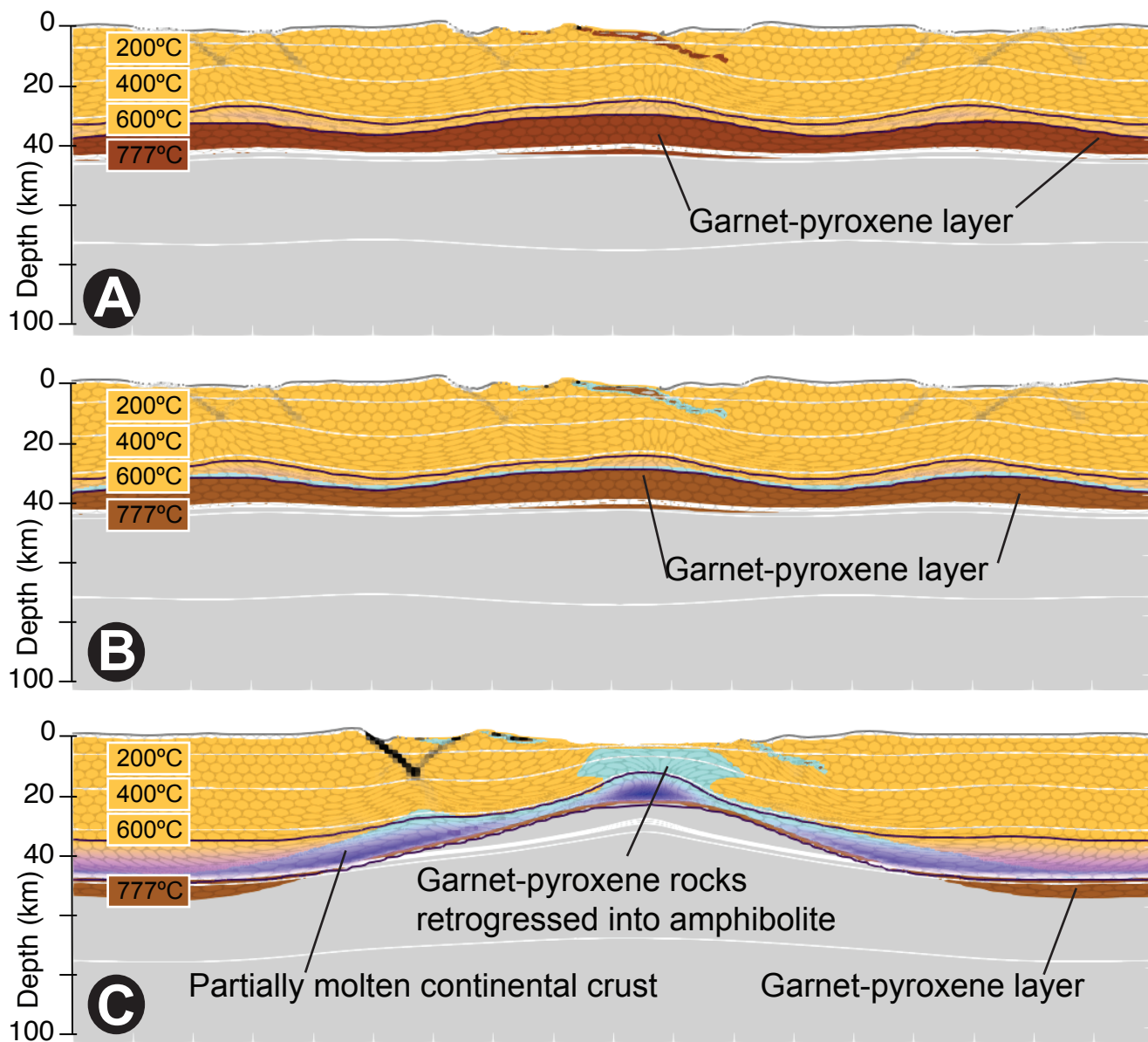
26 Rey, P.F., Muller, R.D., 2010. Fragmentation of active continental plate margins owing to the
27 buoyancy of the mantle wedge. *Nat. Geosci.* 3 (4), 257–261.

28

29



Supplementary Data Figure1



Supplementary Data Figure2