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1 **Quantitative impact of structural inheritance on present-day deformation and seismicity**
2 **concentration in intraplate deformation zones**

3

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7

8 **Highlights**

- 9 - Quantification of strain localization linked with structural inheritance.
10 - Impact of brittle and ductile lithospheric weakening on strain localization.
11 - Parametric study of strain localization in intraplate deformation zones.

12

13 **Abstract**

14 Structural inheritance (i.e. paleo-tectonic) areas, acting as weakened domains, appear to be a
15 key element localizing the seismicity in intraplate deformation zones. However, the impact of
16 structural inheritance on the observed present-day seismicity and strain rate concentration
17 remains to be quantified. In this study, we quantify through 2D numerical modeling the
18 localization and amplification factor of upper crustal strain rates induced by structural
19 inheritance. Our 2D models are constrained by intraplate velocity boundary conditions and
20 include rheology laws that accounts for inherited strain weakening in both the brittle and
21 ductile layers of the lithosphere. The role of structural inheritance is investigated for different
22 localization of the weakened domain in the lithosphere. For an average intraplate geotherm
23 (Moho temperature ca. 500°C), brittle weakening (i.e. inherited faults) alone induces a limited
24 amplification factor of upper crustal strain rates of ca. 4. Ductile weakening can increase the
25 amplification factor to ca. 7 when localized in the lower crust, but has no effect when

26 localized in the lithospheric mantle. Overall, the amplification factors of upper crustal strain
27 rates vary between 1 and 27 depending on the location of the weakened area in the lithosphere
28 and on the different possible net driving forces, crustal strengths, amounts of weakening, and
29 geotherms. These model amplification factors are in reasonable agreement with those derived
30 from GPS and seismicity data over large spatial scale (several hundreds of kilometers) in
31 North America.

32

33 **Keywords**

34 Intraplate deformation zones, Structural inheritance, Rheology weakening, Strain rate
35 amplification factor, Strain concentration

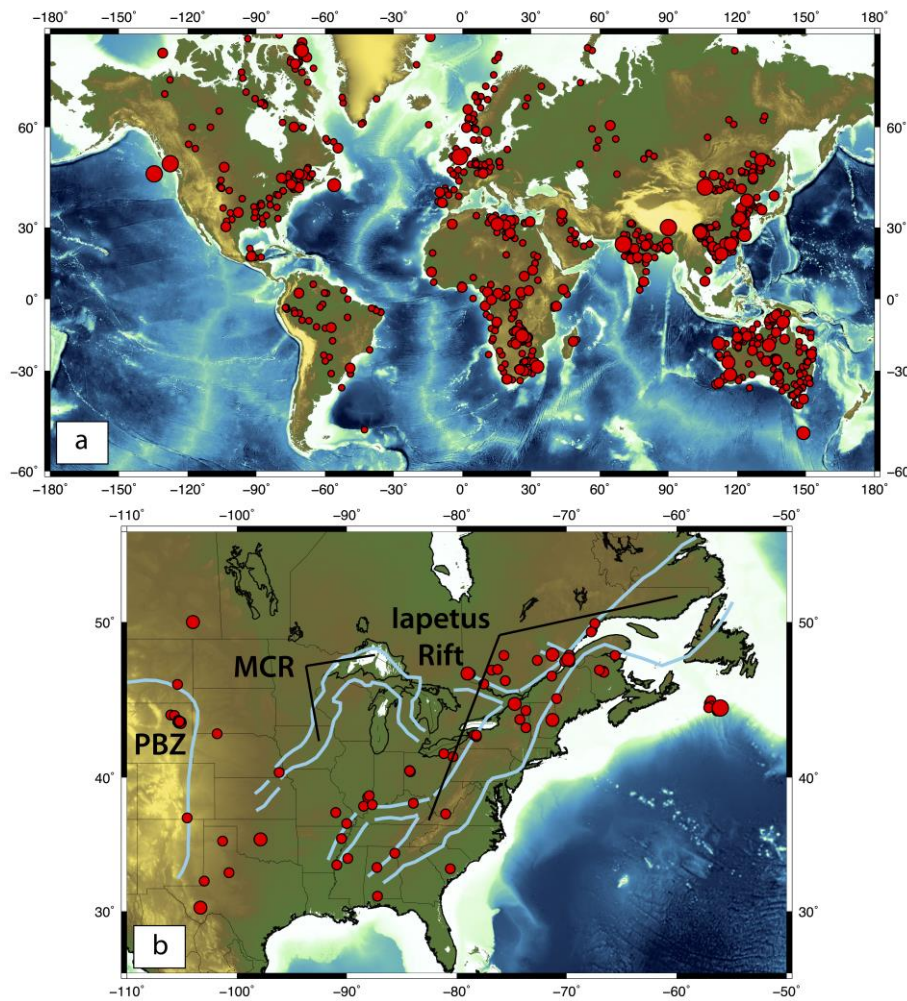
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37

38 **1. Introduction**

39

40 Present-day strain and seismicity in continental intraplate regions are not randomly
41 distributed (Fig. 1a). It is commonly proposed that the localization of intraplate seismicity is
42 related to the presence of structural inheritance zones (Coppersmith et al., 1987; Johnston,
43 1989; Adams and Basham, 1991), which act as weakened domains (Sykes, 1978).
44 Depending on the metrics (number of events, moment budget, etc.), 55–95% of intraplate
45 seismicity is localized in regions of structural inheritance (Johnston, 1989; Schulte and
46 Mooney, 2005). In these studies, structural inheritance is defined as lithospheric-scale tectonic
47 inherited structures (commonly Paleozoic and older). As a consequence, structural inheritance
48 is associated with large domains (tens of kilometers) of significant lithospheric deformation
49 (strain over 100%), mostly related to paleo-rifts or passive margins (Johnston, 1989).



50

51 **Figure 1. Intraplate seismicity.** a/ Global intraplate earthquake catalog (USGS National
 52 Earthquake Information Center). Historical and instrumental earthquakes shown for
 53 magnitudes superior to 4.5 from AD. 495 to 2002. b/ Intraplate seismicity of Central and
 54 Eastern Canada and United States. Blue lines delimit main tectonic features: eastern edge of
 55 North America Plate Boundary Zone (PBZ), Mid-Continent Rift (MCR), Iapetus rifted
 56 margin and grabens.

57

58 Domains presenting structural inheritance are for instance the Iapetus rift in the St
 59 Lawrence Valley, eastern Canada and U.S.A (Kumarapeli, 1966), the Rhine graben in north-
 60 western Europe (Illies, 1972) or the Hercynian system associated with the South American
 61 Shear Zone, western France (Jégouzo, 1980). The observed relation between the presence of

62 structural inheritance and the presence of seismicity is not always verified (Schulte and
63 Mooney, 2005). For example, in the stable continental region of North America (Fig. 1b),
64 seismicity appears to be mostly located along the paleo-rift Iapetus. Conversely, very little
65 seismicity is associated with the Mid Continental Rift (MCR). One of the important
66 consequences of this variability is the integration of structural inheritance in seismic hazard
67 assessment that remains a current challenge (Stein and Mazzotti, 2007).

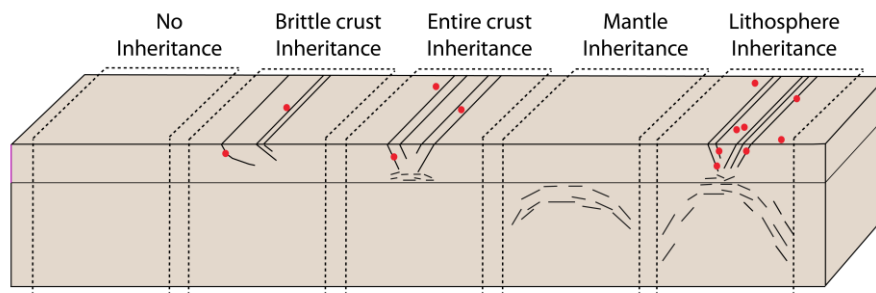
68 Although strain observation in intraplate regions is challenging compared to plate
69 boundary system, seismic and GPS observations can constrain strain rates in term of order of
70 magnitude. For instance, first-order estimations of seismic and GPS strain rates in central and
71 eastern United States or eastern Canada are about 10^{-12} – 10^{-8} yr⁻¹ (Anderson, 1986; Mazzotti
72 and Adams, 2005) and 10^{-10} – 10^{-8} yr⁻¹ (Mazzotti et al., 2005; Tarayoun et al., 2018).

73 Only few studies quantify the impact of structural inheritance on the observed strain
74 and seismicity rates in intraplate domains. In eastern Canada, GPS observations show that
75 structural inheritance amplifies strain rates by a factor of 2-11 (Tarayoun et al., 2018). The
76 impact of a weak zone on surface deformation, which has been studied for various weakening
77 sources, vary between factors of 3-4 (Wu and Mazzotti, 2007) to 100-1000 (Grollmund and
78 Zoback, 2001; Mazzotti and Gueydan, 2017). A significant decrease of viscosity in the lower
79 crust (Kenner and Segall, 2000) or in the lithospheric mantle (Grollmund and Zoback, 2001)
80 can generate strain rate concentrations of 1-3 orders of magnitude in the New Madrid seismic
81 zone, eastern United States. Wu and Mazzotti (2007) investigate the impact of a weak zone in
82 a glacial isostatic adjustment model and show that surface strain rates increase by a factor up
83 to 8 in eastern Canada. Mazzotti and Gueydan (2017) calculate strain rates associated with 1D
84 lithospheric yield stress profiles integrating new rheology laws based on field observations
85 (i.e. mylonite and proto-mylonite) that allow a link between structural inheritance and the
86 reduction of viscosity in both the crust and mantle. They show that to explain observed GPS

87 and seismic strain rates, the crust and lithospheric mantle in intraplate deformation zones must
88 be significantly weakened. The latter study assumes a lithosphere at near-failure equilibrium
89 implying constant strain rates with depth. In low strain regions such as intraplate deformation
90 zones, whole-lithosphere near-failure equilibrium may not be reached due to the existence of
91 an elastic layer (or elastic core, cf. Kuszniir, 1991) between brittle deformation in the upper
92 part of the lithosphere and ductile deformation in the middle or lower parts. The presence of
93 such elastic layers could result in significant effects on strain rate concentration in weak areas.

94 In this paper, we provide first quantitative estimations of the impact of structural
95 inheritance on present-day surface deformation in intraplate deformation zones, i.e.
96 seismically active intraplate regions. We do not attempt to provide quantification of structural
97 inheritance impact in stable continental regions (i.e. cratons) where the active deformation is
98 either not measurable or very poorly constrained. Our study is based on 2D numerical
99 mechanical models. Our models are tuned to intraplate deformation zones conditions
100 (boundary conditions and geotherm) and integrate inherited weakening through a rheology
101 scaling based on field observations (Gueydan et al., 2014). In order for our model to be
102 generic for all intraplate deformation zones, we assume a general weakened domain of several
103 10s km scale (cf. Gorczyk et al., 2012). In other words, we do not investigate the impact of a
104 single fault or a single shear zone but rather, the structural inheritance domain represents the
105 averaged effect of numerous faults and shear zones of any geometry. We focus our analysis
106 on five scenarios testing different inheritance localization (Fig. 2): (1) a non-weakened
107 lithosphere; (2) a weakened domain only in the brittle crust; (3) a weakened domain in the
108 entire crust; (4) a weakened domain only in the lithospheric mantle; (5) a whole weakened
109 lithosphere. The five scenarios are meant to represent all intraplate deformation zones, from a
110 thick-skin thrust system for upper crust inheritance (for instance the Appalachians Province,

111 Eastern Canada; Thomas, 2006) to a rift structure for whole lithosphere inheritance (for
112 instance the paleo Iapetus rift).



113

114 **Figure 2. Conceptual scenarios of possible structural inheritance localization in the**
115 **lithosphere and associated earthquakes** (red dots).

116

117 We quantify the amplification factor of present-day upper crustal strain rates related to
118 each scenarios using velocity boundary conditions coupled with an integrated lithospheric
119 strength control, in order to take into account fully elasto-visco-plastic deformation.

120 Compared to previous modeling studies, the main novelties of our study are, first, the use of
121 2D numerical models that allow taking into account variations of strain rates both laterally
122 and with depth. This point allows robust results of the concentration of upper crustal
123 deformation associated with lower-crust and upper-mantle weakened domains. Second, we
124 quantify the role of inherited weakening within the conditions of a fixed net driving force.

125 This point allows investigating the strength of the lithosphere and, thus, the associated upper
126 crustal strain rates at different mechanical stages before reaching steady-state deformation.

127

128 **2. Numerical model setup**

129

130 **2.1 Rheology**

131

132 **2.1.1 Reference rheologies (non-weakened lithosphere)**

133

134 The mechanical behavior of the lithosphere is defined by brittle and ductile rheology
135 laws, commonly presented as yield stress profiles, which correspond to the minimum between
136 brittle and ductile differential stresses for given depths, temperature profiles and strain rates.
137 Hereafter, we use analytic yield stress profiles to compare with differential stress profiles
138 derived from the numerical models at various mechanical stages (cf. section 3). In analytic
139 profiles, the brittle yield stress is equal to the Mohr-Coulomb stress (Byerlee, 1978):

140

$$141 \quad \sigma_B = p * \sin(\phi_B) + C * \cos(\phi_B) \quad (1)$$

142

143 where p is the lithostatic pressure (density of 2.7 g.m^{-3} and 3.3 g.m^{-3} for the crust and mantle,
144 respectively), C the cohesion (10 MPa) and ϕ_B the internal friction angle (30°) as defined by
145 Byerlee (1978). In the numerical model, the brittle stress is equal to the Drucker-Prager stress
146 (Chéry et al., 2001), which is an approximation of the Mohr-Coulomb failure criterion (Owen
147 and Hinton, 1980):

$$148 \quad J_2 = \frac{-1}{3} J_1 + \frac{c}{\tan(\phi_D)} \quad (2)$$

149

150 where J_1 and J_2 are the first and the second invariants of the stress tensor, respectively. To
151 equalize the internal friction angle between Mohr-Coulomb and Drucker-Prager laws, we set
152 ϕ_D at 15° (Chéry et al., 2001).

153 In both analytic and numerical models, the ductile yield stress σ_D is derived from the
154 dislocation creep law (Weertman, 1978):

155

$$156 \quad \sigma_D = \left(\frac{\dot{\epsilon}}{A}\right)^{-n} * \exp\left(\frac{Q}{nRT}\right) \quad (3)$$

157

158 where $\dot{\epsilon}$ is the strain rate (s^{-1}), T the temperature (K) and R the gas constant
159 ($8.31 \text{ J mol}^{-1} \text{ K}^{-1}$). As a first-order approximation of lithospheric mineral composition, quartz
160 and olivine rheology parameters are used for crust and mantle, respectively ($A = 1.1 \times 10^5$ and
161 $3.9 \times 10^{-10} \text{ Pa}^{-n} \text{ s}^{-1}$, $Q = 135$ and $530 \text{ J} \cdot \text{mol}^{-1}$, and $n = 4$ and 3.5 ; Hirth and Kohlstedt, 2003; Luan
162 and Paterson, 1992). A stronger rheology in the crust will also be considered (section 5.1).

163

164 *2.1.2 Weakened rheologies*

165

166 Several mechanisms inducing inherited strain weakening have been proposed. In the
167 brittle crust, maturation of fault zones is achieved by nucleation of new minerals, such as
168 mica or talc, decreasing the friction coefficient from ca. 0.6 to 0.1 (Holdsworth, 2004). In the
169 ductile crust, intense weakening is related to the progressive development of layering (shear
170 zone or foliation) enriched in mica (Wintsch et al., 1995; Gueydan et al., 2003). Shear heating
171 is also proposed as a weakening process in the deep crust (Regenauer-Lieb and Yuen, 2003;
172 Thielmann and Kaus, 2012). In the lithospheric mantle, two main processes could promote
173 inherited strain weakening: grain size reduction during dynamic recrystallization of olivine
174 (Hirth and Kohlstedt, 2003, Précigout and Gueydan, 2009) and preferred orientation of
175 olivine leading to an inherited anisotropy (Tommasi et al., 2009).

176 Annealing (dynamic or static recrystallization) and reduction or suppression of the
177 inherited weakening is possible if an event involving a significant increase of temperature
178 (e.g., tectonic event or hotspot) occurs after the formation of the structural inheritance (Boneh
179 et al., 2017). In an intraplate deformation zone where no major tectonic event has occurred
180 since Paleozoic, and with generally low geotherms, we can assume that, in most cases, no
181 annealing has taken place and thus the mechanisms of inheritance weakening are maintained
182 through time.

183 In this study, we model inherited weakening using a generic expression that can
184 represent any of the weakening processes listed above. Following Mazzotti and Gueydan
185 (2017), we introduce an inherited strain-weakening rheology law that consists in integrating a
186 finite weakening in the standard brittle and ductile rheology laws (section 2.1.1). The effect
187 on differential stress of this finite weakening, based on field observations, is from the study of
188 Gueydan et al. (2014):

189

$$190 \quad \sigma = \sigma_0 * \left[1 + \alpha * \left(\exp^{\frac{-\varepsilon}{\varepsilon_c}} - 1 \right) \right] \quad (4)$$

191

192 where σ_0 is either the initial brittle stress ($\sigma_B, J_2, Eqs. 1 - 2$) or ductile stress ($\sigma_D, Eq. 3$), α
193 represents the maximum strain weakening factor (reduction of strength or effective viscosity
194 for brittle and ductile behavior, respectively), ε the finite inherited strain of the considered
195 domain and ε_c the characteristic strain over which the deforming rock fabric changes
196 according to layering development (in the crust) or grain size reduction (in the mantle). The
197 maximum strain weakening ($\alpha = 0.9$) and the characteristic strain ($\varepsilon_c = 0.5$) are based on
198 numerical experiments of large deformation (Gueydan et al., 2014).

199 Thus, in our model, the amount of effective weakening is controlled by the finite strain
200 parameter, ε , specific to a given region. As we defined structural inheritance as lithospheric-
201 scale paleo-structures, we assume a large finite strain ($\varepsilon = 2$), equivalent to a stress scaling
202 factor of 0.12. Lower finite strain will be tested (section 5.1). The weakening effect is
203 restricted to small temperature ranges: from 0°C to 500°C and from 600°C to 800°C for the
204 crust and mantle, respectively. Outside those ranges of temperature, the weakening disappears
205 due to the mineral transformations leading to possible hardening in the crust (Gueydan et al.,
206 2014) and to the lower impact of grain size reduction phenomenon in the mantle (Précigout
207 and Gueydan, 2009).

208

209 *2.1.3 Weakening impact on lithospheric yield stress profiles*

210

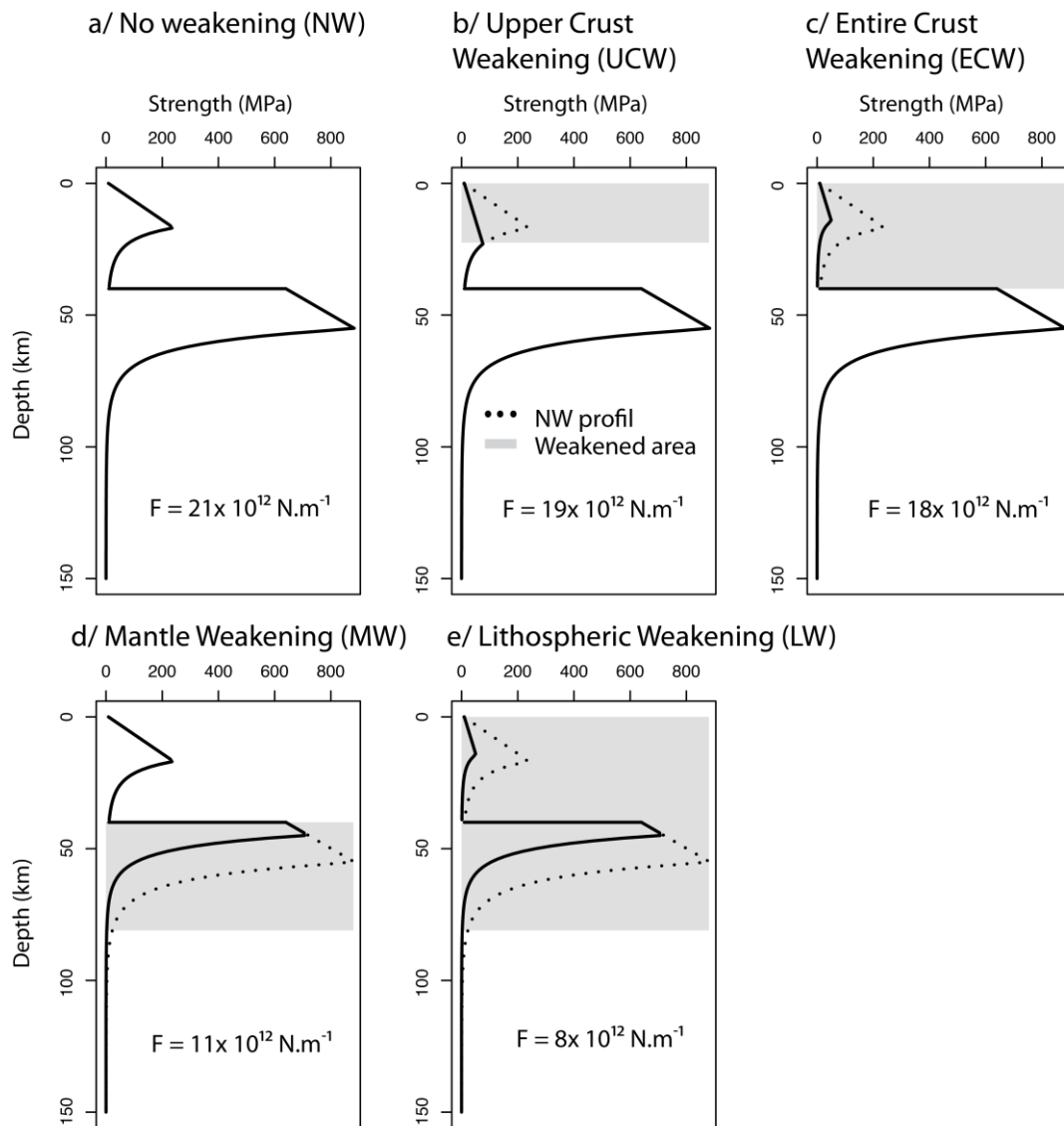
211 In order to illustrate the effect of the weakening laws, we calculate analytic yield stress
212 profiles for the five scenarios presented in Figure 2: non-weakening (NW), Upper Crust
213 Weakening (UCW), Entire Crust Weakening (ECW), Mantle Weakening (MW) and whole
214 Lithosphere Weakening (LW). Results are presented in Figure 3. Yield stress profiles are
215 calculated for a finite inherited strain $\varepsilon = 2$ and with a constant strain rate with depth $\dot{\varepsilon} = 5.6$
216 $\times 10^{-17} \text{ s}^{-1}$, which corresponds to the bulk deformation expected in the numerical modeling.
217 We use the Moho temperature, T_M , as a proxy for the geotherm. The surface temperature is
218 set at 0°C , T_M at 500°C and the base of the lithosphere (150 km-depth) at 1300°C . The crustal
219 thickness is 40 km, which is a reasonable average for continental intraplate regions (Mooney
220 et al., 1998).

221 With non-weakening (Fig. 3a), the maximal yield stresses (at the brittle-ductile
222 transitions) are 235 and 883 MPa for the crust and mantle, respectively. With an upper crust
223 weakening, the maximal yield stress at the crustal brittle-ductile transition drops to 76 MPa
224 (Fig. 3b). Ductile weakening in the lower crust has a low impact on strength reduction (Fig.
225 3c). The major weakening impact is in the lithospheric mantle between 40 and 80 km depth,
226 where the maximal yield stress drops to 706 MPa (Fig. 3d). With a whole lithosphere
227 weakening (Fig. 3e), the maximal yield stresses drop to 50 and 706 MPa for the crust and
228 mantle, respectively.

229 The impact of the weakening is also expressed through the integrated lithospheric
230 strength, which is calculated for each scenario as the depth integral of yield stress down to the
231 lithosphere thickness defined by the 1300°C isotherm. Unsurprisingly, the highest integrated
232 lithospheric strength is reached with a non-weakened lithosphere (Fig. 3a) at $21 \times 10^{12} \text{ N.m}^{-1}$.

233 The lowest integrated lithospheric strength is reached with a whole lithosphere weakening
 234 (Fig. 3e) at $8 \times 10^{12} \text{ N.m}^{-1}$. Intermediate strengths are found with a brittle, entire crust and a
 235 mantle weakening at 19, 18 and $11 \times 10^{12} \text{ N.m}^{-1}$, respectively. The mantle weakening reduces
 236 the integrated lithospheric strength more than 60%, versus ca. 10% with a brittle or entire
 237 crust weakening. This major role of a weakened mantle on analytic yield stress profiles is one
 238 of main result of Mazzotti and Gueydan (2017). The importance of investigating the
 239 integrated lithospheric strength will be presented in section 3.

240



241

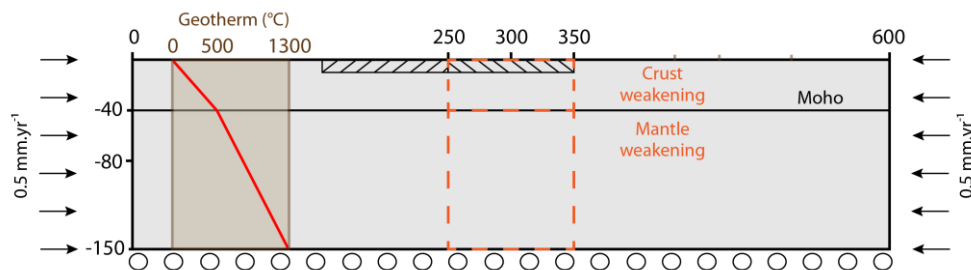
242 **Figure 3. Theoretical yield stress profiles without and with weakening.** Mohr-Coulomb
 243 criterion and dislocation creep law are used for brittle and ductile behavior, respectively. A
 244 weakening coefficient is used for brittle and ductile weakening (cf. text). Yield stress profiles
 245 calculated for various localizations of the weakened domain (grey area), with a uniform strain
 246 rate of $5.6 \times 10^{-17} \text{ s}^{-1}$. F: integrated lithospheric strength assumed at equilibrium with net
 247 driving force.

248

249 2.2 Geometry and boundary conditions

250

251 We use the 2D numerical thermo-mechanical finite-element code ADELI (Hassani et
 252 al., 1997). The model integrates elastic, viscous and plastic behaviors. Our model is tuned to
 253 apply to a generic intraplate deformation zone represented by a lithosphere of 600 km length
 254 and 150 km thickness including a 40-km-thick crust (Fig. 4). It is discretized in 10 000 linear
 255 elements (triangles), with a node interspacing of ca. 4 km. The geotherm is uniform for the



256

257 **Figure 4. Geometry and boundary conditions of the elasto-visco-plastic thermo-**
 258 **mechanical model.** Basal boundary condition: null vertical velocity and a free horizontal
 259 velocity. Lateral boundary condition: null vertical velocity and fixed horizontal velocity (e.g.,
 260 0.5 mm.yr^{-1}). Black numbers are distances (km). Weakened areas are delimited by orange
 261 dashed lines. The two hatched zones between 0 and 10 km depth indicate the two areas where
 262 upper crustal strain rate amplification factor is calculated (weakened over non-weakened
 263 area).

264
265 whole model and defined as linear gradients between the surface (0 °C), Moho, and base
266 (1300 °C) temperatures. In continental intraplate domains, measured surface heat flow and
267 geotherm models correspond to Moho temperatures varying between 400-500 °C in the
268 coldest environments (e.g., Canadian Shield; Mareschal et al., 2000) and ca. 600 °C in milder
269 settings (e.g., central USA; Zoback and Townend, 2001). As our models represent intraplate
270 deformation zones, excluding cratons, we set the reference Moho temperature at 500°C.
271 Higher Moho temperatures will be considered in section 5.1.

272 The models are constrained by a shortening velocity of 1 mm.yr⁻¹ (0.5 mm.yr⁻¹ on each
273 side of the model), constant with depth. The impact of lower velocities (0.05, 0.1 and 0.5
274 mm.yr⁻¹), representative of the range of deformation rates in intraplate deformation zones,
275 will be tested section 5.1. The base of the model is free horizontally and fixed in the vertical
276 component. These boundary conditions define a displacement flow that converges towards the
277 model center, with no strain rate concentration in the upper crust in the central region
278 compared to the peripheries (see section 3). This ensures stable numerical results in the
279 various tests. The kinematic conditions predefine the model bulk strain rate. Thus, in order
280 not to provide strain rate results that are controlled by the boundary conditions, we do not
281 discuss the modeling results in terms of absolute strain rate values but rather as a normalized
282 strain rate (relative to the predefined bulk). Similarly, we express the impact of structural
283 inheritance on upper crustal deformation in terms of an amplification factor, i.e. the ratio of
284 strain rate in the weakened region over that in a non-weakened region (calculated over two
285 conterminous 100 x 10 km zones, cf. Fig. 4).

286 The chosen length of 100 km of the weakened area represents an approximation of the
287 spatial extent of structural inheritance (for example, the paleo Iapetus rift, North America, or
288 the Hercynian domain in western France including the South Armorican Shear Zone). Ductile

289 deformation leading to shear zones in the lower crust and lithospheric mantle in rift zones also
290 appear to be spatially spread over 50-100 km (Gueydan et al., 2008). In our approach, we
291 assumed that the weakening occurs homogeneously over the 100 km length. In other words,
292 the structural inheritance is modeled as a weakened domain representing a distributed fault
293 and shear zone system. Modeling a weakened domain allows representing any intraplate
294 deformation zones, whereas modeling a complex fault system would be representative of one
295 specific area.

296

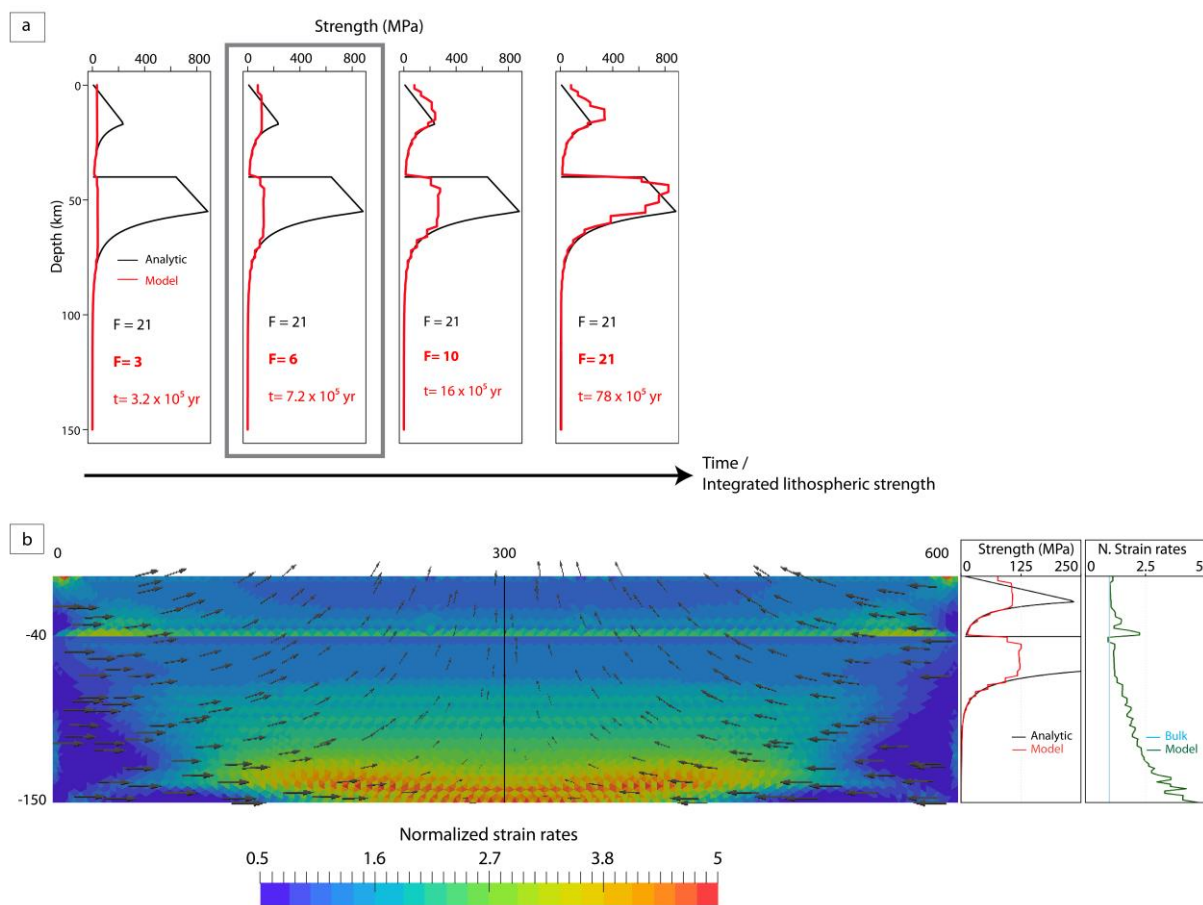
297 **3. Reference non-weakened model**

298

299 Because of the velocity boundary conditions, the strain and stress values in our
300 numerical models change with every time step. Figure 5a shows profiles of differential stress
301 (second invariant J_2) of the model at various run times (0.3 – 7.8 Myr), compared with the
302 steady-state yield stress analytic profile. The model differential stresses increase with each
303 time step, until it becomes similar to the analytic yield stresses at 7.8 Myr. Slight differences
304 exist between the two for the brittle domains that can be attributed to the Drucker-Prager vs.
305 Mohr-Coulomb parameterizations (cf. section 2.1.1). The modeled integrated lithospheric
306 strength follows a similar pattern and reaches the analytic value ($21 \times 10^{12} \text{ N.m}^{-1}$) at 7.8 Myr.

307 Thus, model stress and strain vary with time, depending primarily on the imposed
308 velocity boundary condition. In order not to depend on the imposed velocity, for which we
309 only know the upper bound in intraplate deformation zones, we analyse the model results
310 assuming that the lithosphere is at equilibrium between the integrated strength and a net force
311 that corresponds to the combined effect of tectonic and other transient processes (cf. Zoback
312 and Townend, 2001; Mazzotti and Gueydan, 2017). Estimations of tectonic forces range from
313 1 to $10 \times 10^{12} \text{ N.m}^{-1}$ (e.g., Forsyth and Uyeda, 1975; Copley et al., 2010). Transient processes

314 such as erosion or sedimentation pulses, or glacial isostatic adjustment can produce force
 315 increments about $1-2 \times 10^{12} \text{ N.m}^{-1}$ (Calais et al., 2010; Wu and Johnston, 2000; Wu and
 316 Mazzotti, 2007). Hereafter, we consider models associated with an integrated lithospheric
 317 strength and a net driving force of $6 \times 10^{12} \text{ N.m}^{-1}$, which corresponds to a time run of 0.7 Myr
 318 (Fig. 5a). The impact of lower ($3 \times 10^{12} \text{ N.m}^{-1}$) and higher ($10 \times 10^{12} \text{ N.m}^{-1}$) forces will be
 319 tested section 5.1. As a result, our analysis corresponds to models at intermediate run times
 320 and differential stress profiles and not to model that have reached a steady state.



321
 322 **Figure 5. Reference model with no weakening.** a/ Comparison of analytic steady-state yield
 323 stress (black) and model J2 stress (red). F: integrated lithospheric strength assumed at
 324 equilibrium with net driving force ($\times 10^{12} \text{ N.m}^{-1}$), t: time (yr). b/ Non-weakening model for a
 325 force $F = 6 \times 10^{12} \text{ N.m}^{-1}$. Background colours are normalized strain rates (relative to model
 326 bulk). Black vectors show the displacement field. Black numbers are distances (km). Right
 327 panels: analytic and model stress profiles and normalized strain rate profile located at model

328 center (vertical black line). Bulk line represents boundary condition mean strain rate (e.g.,
329 velocity of 1 mm.yr⁻¹ over 600 km).

330

331 For this net driving force of $6 \times 10^{12} \text{ N.m}^{-1}$, a major feature in our model is the
332 presence of large elastic layers (Fig. 5a) due to the slow stress build up. The presence of
333 elastic layer in the lithosphere for a non-steady state model is well established (e.g. Kuzsnir,
334 1991). For forces of $(3 - 6) \times 10^{12} \text{ N.m}^{-1}$, elastic layers are preserved in the upper-middle crust
335 and in the upper lithospheric mantle. The thickness of the elastic layers decreases with time as
336 differential stress build up to reach brittle and ductile yield stress values. For a force of $10 \times$
337 10^{12} N.m^{-1} , the elastic layer has disappeared in the crust. Whole lithosphere near-failure
338 equilibrium occurs for a force of $21 \times 10^{12} \text{ N.m}^{-1}$.

339 The overall deformation pattern of the non-weakened model is presented in Figure 5b.
340 In order not to depend on the imposed boundary velocity and to help visualisation, we present
341 the strain rates as normalized to the overall model bulk strain rate (boundary velocity divided
342 by the model length). High strain rates are concentrated in two main shear zones: just above
343 the Moho (due to the weak ductile stress of quartz at the lowermost-crust temperature) and at
344 the base of the lithosphere (due to the vertical-fixed base of the model). Conversely, low
345 strain rates occur in domains of high differential stress in the upper crust and upper
346 lithospheric mantle. The two elastic layers seen in Figure 5a are characterized by low strain
347 rate values. The thickness of the elastic layers is ~18 km in the upper-middle crust and ~27
348 km in the upper lithospheric mantle. As discussed section 2.2, the model presents no strain
349 rate concentration in the upper crust in the central region, compared to the peripheries (Fig.
350 5b). This point is important for the interpretations of the following models where the
351 weakened domain is located in the center. The high strain rates located on each upper corners
352 of the model are due to the velocity boundary condition. However, the strain rate

353 amplification factors are calculated near the center of the model (see Figure 4) and are thus
354 not affected by these boundary effects.

355

356 **4. Weakened models for a net driving force of $6 \times 10^{12} \text{ N.m}^{-1}$**

357

358 In the following, we assess the effect on upper crustal strain rate amplification factor
359 of weakened zones in various locations (from upper crust to whole lithosphere) for a given
360 amount of weakening ($\epsilon = 2$), geotherm ($T_M = 500 \text{ }^\circ\text{C}$) and net driving force ($F = 6 \times 10^{12}$
361 N.m^{-1}). The models are shown in Figure 6 and, for each model, the upper crustal strain rate
362 amplification factor is shown Figure 7 with respect to the reference non-weakened model.

363

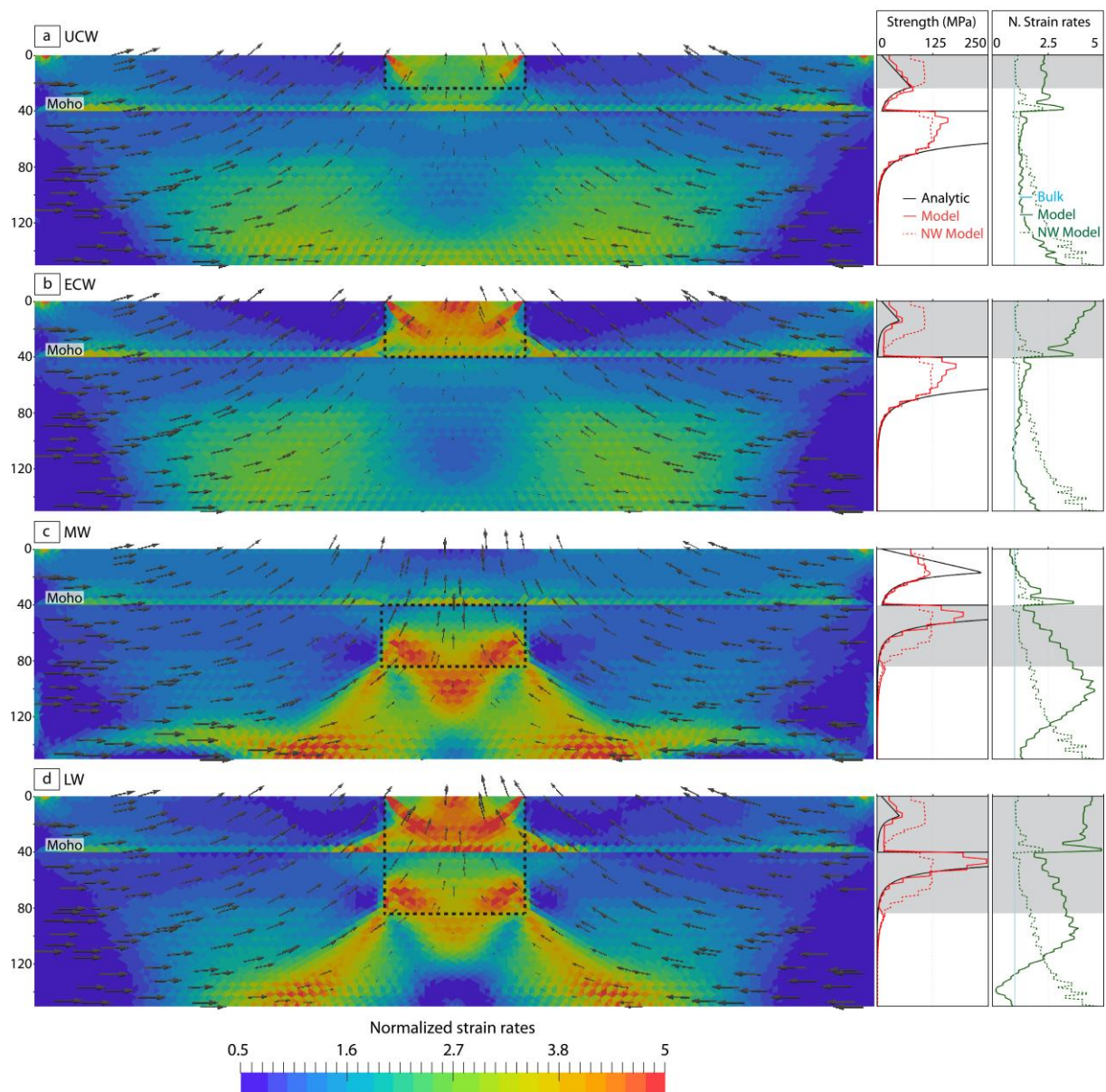
364 ***4.1 Upper Brittle Crust Weakening (UCW)***

365

366 With reduced friction coefficient, the strain rate amplification factor in the uppermost
367 crust (strain rate ratio in weakened over non-weakened area) induced by the weakened upper
368 crust is about a factor of 4 (Fig. 7). The strain rates concentrate in two bands on each side of
369 the weakened area (Fig. 6a), which correspond to first-order to the Coulomb frictional bands
370 that tend to accommodate and localize the shortening across the weakened upper crust. The
371 maximum differential and yield stress in the weakened crust drop from $\sim 125 \text{ MPa}$ to ~ 75
372 MPa . Those two values correspond to the elastic differential stress and to the weakened brittle
373 yield stress, respectively. Brittle weakened crust implies that yield stress is reached in the
374 whole crust leading to the disappearance of the elastic layer in the weakened zone.

375 The strain rate concentration in the upper crust weakening impact the whole
376 lithosphere profile. In the lower crust, strain rates increase compared to the reference non-
377 weakened model due to stress concentration in the weakened upper-middle crust. In contrast,

378 strain rates decrease in the lithospheric mantle. The presence of the weakened upper crust
 379 creates a reorganization of the displacement field, leading to a reorganization of stress and
 380 strain rates. Because the boundary conditions prescribed the overall strain rate in the model, a
 381 local increase of the strain rates has to be balanced by a local decrease elsewhere in the
 382 model. This process explains also the reduced upper crustal strain rates directly outside the
 383 weakened area (ratio ca. 0.5 – 0.6).



384

385 **Figure 6. Models with weakening for a force of $6 \times 10^{12} \text{ N.m}^{-1}$.** Legend as Fig. 5. Dash
386 lines in model and grey shaded areas in profiles show weakened areas. NW profile: reference
387 non-weakening profile (Fig. 5b).

388

389 ***4.2 Entire Brittle and Ductile Crust Weakening (ECW)***

390

391 With an entire crust weakened, the amplification factor of the upper crustal strain rate
392 is about a factor of 6.6 (Fig. 7). The concentrated strain rates are mostly localized in three
393 specific zones (Fig. 6b): (1) the two Coulomb bands on each side of the weakened area, which
394 have propagated in depth and connected to the lower crust; (2) a shear zone at the Moho; and
395 (3) a major brittle zone at the surface and center of the weakened area. The latter is not seen
396 on the UCW model. The localized strain rate zones imply lateral variations of upper crustal
397 strain rates. The major feature of the ECW model is that ductile weakening in the lower-
398 middle crust significantly impacts the upper crustal strain rate concentration and amplification
399 factor. Ductile weakening involves reduced differential stresses and larger strain rates (Eqs. 3-
400 4) in the lower-middle crust. Compared to brittle weakening alone (Fig. 6a), the upper crustal
401 strain rate is amplified by a factor of 2. This strong mechanical coupling between brittle and
402 ductile layers is highlighted in studies investigating the role of each deformation mechanisms
403 (i.e. brittle failure and viscous flow) in localized or distributed fracturing (e.g., Schueller et
404 al., 2005, 2010).

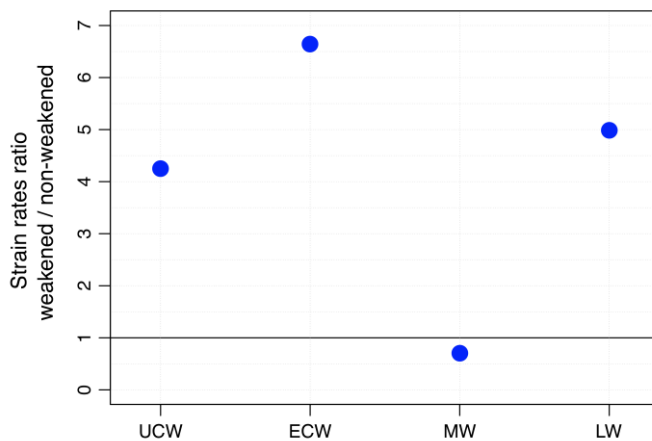
405

406 ***4.3 Ductile Mantle Weakening (MW)***

407

408 As seen section 2.1.2, only the lithospheric mantle with a temperature lower than
409 800°C is weakened, resulting in non-weakened lower lithosphere. Surprisingly, weakening of

410 the lithospheric mantle results in a reduction of upper crustal strain rate by a factor of 0.9
 411 (Fig. 7). The displacement field associated with the enhanced mantle flow towards the weak
 412 domain leads to the development of major shear bands in the lower part of the weakened
 413 mantle and in the non-weakened mantle to accommodate the localized flow. More
 414 specifically, the strain rates are localized on each side of the lower weakened part and in shear
 415 zones from the weakened mantle to the base of the lithosphere (Fig. 6c). The weakened
 416 mantle also creates a minor shear zones at the Moho. Despite the strong weakening in the
 417 upper lithospheric mantle, the elastic layer is still present in the uppermost part of the
 418 weakened mantle (40 – 60 km depth). This elastic layer prevents the stress propagation from
 419 the mantle to the surface, explaining the absence of high strain rates in the crust.



420
 421 **Figure 7. Upper crustal strain rate amplification factor for five weakening scenarios.**
 422 UCW: Upper Crust Weakening, ECW: Entire Crust Weakening, MW: Mantle Weakening and
 423 LW: Lithosphere Weakening. Amplification factor expressed as ratio of average strain rate
 424 inside weakened area over non-weakened area (cf. Fig. 4).

425
 426 **4.4 Brittle and Ductile Lithospheric Weakening (LW)**

427
 428 The model with whole lithosphere weakening combines the high upper crustal strain
 429 rate impact of the entire crust weakening and the low upper crustal strain rate impact of the

430 mantle weakening (Fig. 6d). The weakened lithosphere area induces an amplification factor of
431 the upper crustal strain rate by a factor of 5 (Fig. 7). Higher displacements in the weakened
432 domain lead to a major shear zones in the lower crust. This induces that the strain rate is
433 slightly higher in the lower weakened crust than with the entire crust weakening (fig 6b).

434

435 **5. Parametric study**

436

437 In order for our models to be applied to any intraplate deformation zones, the impact
438 of five major parameters (i.e. the velocity boundary condition, the crustal rheology, the
439 amount of weakening, the net driving force, and the geotherm) will be tested separately. We
440 investigate the influence of each parameter for the five scenarios of weakening localization.
441 We pay specific attention to the lithospheric mechanical behavior related to the mantle elastic
442 layer and to the mechanical coupling between the mantle, the ductile crust and the brittle
443 crust.

444

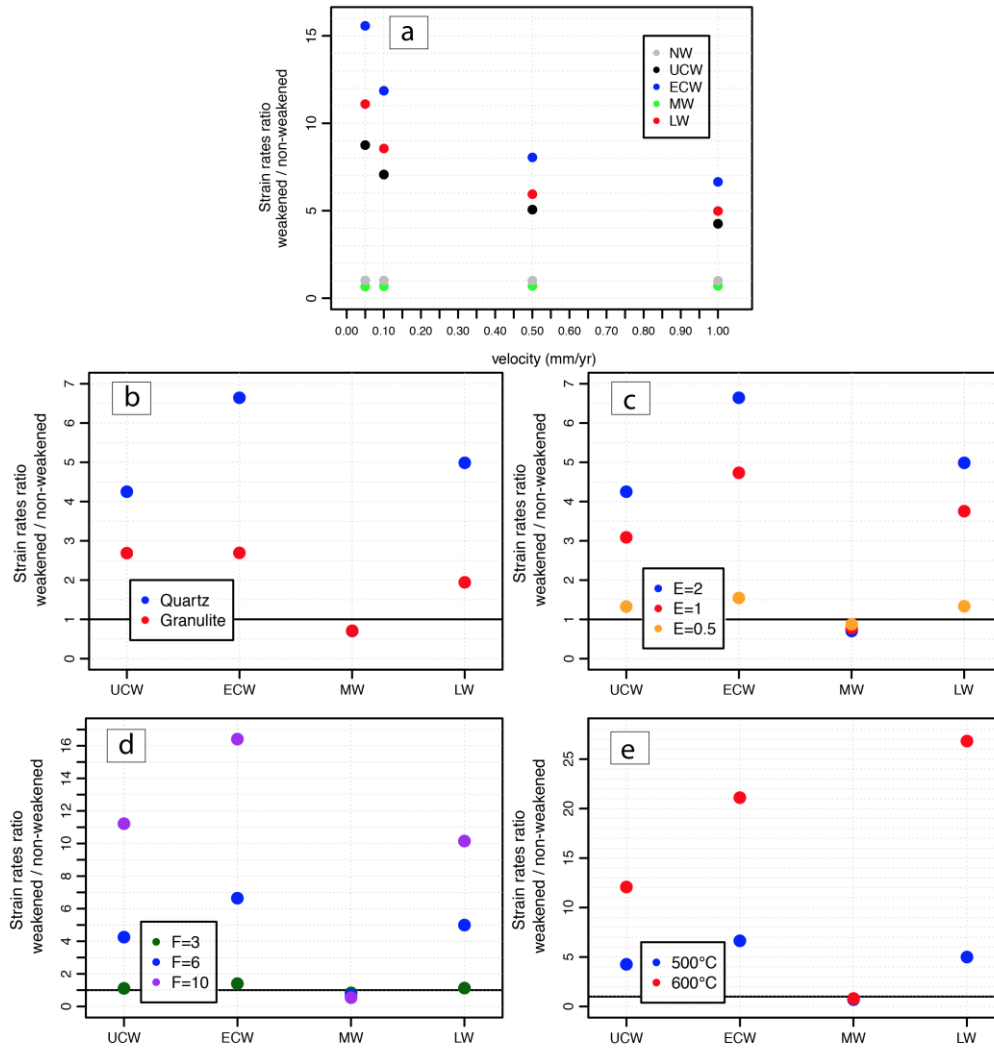
445 ***5.1 Parameters sensitivity***

446

447 We test the impact of the velocity boundary condition using three values (0.05; 0.1
448 and 0.5 mm.yr⁻¹, in addition to 1 mm.yr⁻¹ in the reference models) representative velocities of
449 intraplate deformation zones (Figure 8a). Lower velocities display higher amplification
450 factors, indicating a higher impact of structural inheritance. Changing the velocity boundary
451 condition affects the amplification factors because we consider models for a given net driving
452 force. Thus, lower boundary velocities imply lower average strain rates, promoting viscous
453 versus elastic behaviors. Nevertheless, the amplification factors remain of the same order of
454 magnitude in all experiments, varying from 1 – 7 in the reference models (velocity of 1

455 mm.yr⁻¹) to 1 –15 in the slowest models (velocity of 0.05 mm.yr⁻¹). The five weakening
456 scenarios maintain the same deformation features with different velocities, e.g. a lower LW
457 model amplification factor compared to the ECW model. This is due to the stability of the
458 lithosphere rheological stratification with different velocity boundary conditions (see
459 Appendix). For instance, the elastic layer is preserved over the same depth range of the upper
460 part of the weakened mantle for all tested velocity boundary conditions, with only limited
461 thinning with low velocity boundary condition. The impact of weakening on strain
462 concentration is then linked to minor changes as the thickness of the elastic layer in the upper
463 part of the weakened mantle decreases. Thus, the velocity boundary conditions do not
464 significantly affect the model results and amplification factors, indicating, to first order, a
465 linear scaling with the velocities.

466 To investigate the impact of crustal rheology, we consider a strong crust composed of
467 granulite (dislocation creep parameters $A = 1.4 \times 10^4 \text{ Pa}^{-n} \text{ s}^{-1}$, $Q = 445 \text{ J.mol}^{-1}$, $n = 4.2$; Wilks
468 and Carter, 1990) instead of the quartz rheology used in the reference models (Fig. 8b). This
469 provides, to first order, upper and lower limits on the rheology impact on upper crustal strain
470 rate concentration. Compared to the quartz models, the amplification factor is significantly



471

472 **Figure 8. Parametric analysis of upper crustal strain rate amplification factor.**

473 Amplification factors are calculated for different (a) velocity boundary conditions, (b) crustal
 474 rheology (granulite vs. quartz), (c) amount of weakening (E: inherited strain), (d) net driving
 475 force and (e) geotherm (given as Moho temperature). Blue circles for (b), (c), (d) and (e) are
 476 reference amplification factors shown Figure 7. Note different representation between (a) and
 477 (b – e).

478

479 smaller for all granulite models (down to factors of 2 – 2.7), except for the MW model (factor
 480 of 0.9). In the weakened area, the strain rate concentration is similar (two bands) for both
 481 rheologies but with lower values for the granulite. The granulite rheology implies a highest

482 yield stress, which results in (1) strain rates that are lower over the lithosphere column and (2)
483 brittle failure occurring in the whole crust for both UCW and ECW models, leading to similar
484 strain rate amplification factors.

485 The inherited finite strain in a structural inheritance domain controls the amount of
486 weakening. In the reference cases, we assumed a high finite strain of 2 as representative, for
487 example of a mature rift with lithospheric-scale paleo-structures (Musacchio et al., 1997).
488 Figure 8c presents the impact of lower finite strains of 1 and 0.5 on the upper crustal strain
489 rate amplification factors. Reducing the finite strain reduces the upper crustal strain rate ratios
490 for all models (except the MW model) down to factors of 3 – 5 for a finite strain of 1 and 1 –
491 1.5 for a finite strain of 0.5. The strain rate concentration in the weakened area is similar but
492 with lower strain rate values as we decrease the finite strain. With a low finite strain, the
493 models for the five scenarios tend toward those of the NW model.

494 Variations of the net driving force change elastic layer thickness and thus the upper
495 crustal strain rate concentration (see section 3). We quantify the impact on the upper crustal
496 strain rate amplification factors of net driving forces of $10 \times 10^{12} \text{ N.m}^{-1}$ and $3 \times 10^{12} \text{ N.m}^{-1}$ (vs.
497 reference value of $6 \times 10^{12} \text{ N.m}^{-1}$, Fig. 8d). For a force of $10 \times 10^{12} \text{ N.m}^{-1}$, the ratios increase
498 for all models (except the MW model) reaching 16.5 for the ECW model. Because of the
499 strain rate adjustment process, the upper crustal strain rates surrounding the weakened area
500 decrease while they increase in the weakened area. This implies a higher weakened over non-
501 weakened ratio for a force of $10 \times 10^{12} \text{ N.m}^{-1}$. For a force of $3 \times 10^{12} \text{ N.m}^{-1}$, the ratio
502 decreases for all models (except the MW model) down to factors of 1 – 1.5. The mantle flow
503 is slower, leading to a lower concentration of strain rates in the weakened area. Thus, for a net
504 driving force of $3 \times 10^{12} \text{ N.m}^{-1}$, weakening has no significant impact on upper crustal strain
505 rate amplification factors.

506 Finally, we test a geotherm defined by $T_M = 600^\circ\text{C}$ (vs. reference value of 500°C) in
507 order to quantify the impact of temperature on upper crustal strain rate amplification factors
508 (Fig. 8e). The upper crustal strain rate ratio increases for all models up to factors of 21 – 27
509 (except the MW model). The highest amplification factor difference is with the LW model.
510 For $T_M = 600^\circ\text{C}$, the differential stresses in the whole weakened lithosphere are significantly
511 lower than with a $T_M = 500^\circ\text{C}$. More particularly, ductile flow occurs in the whole upper
512 lithospheric mantle, suppressing the elastic layer preserved with a $T_M = 500^\circ\text{C}$. As a
513 consequence, the mechanical crust-mantle coupling is stronger, leading to an increase of the
514 upper crustal strain rate amplification factor.

515

516 *5.2 Summary of amplification factor parameters variability*

517

518 The parameter that has the highest impact on upper crustal strain rate amplification
519 factor is the geotherm. A relatively high Moho temperature ($T_M = 600^\circ\text{C}$) leads to maximal
520 amplification factors of 21 – 27. A high net driving force ($10 \times 10^{12} \text{ N}\cdot\text{m}^{-1}$) also promotes
521 high amplification factors of 11 – 17. These two parameters play a major role in upper crustal
522 strain rate concentration only for a high amount of weakening (i.e. a finite strain of 2). The
523 role of the weakening is fundamental to produce high concentration and amplification factors.
524 The parameter that has the lowest impact on the upper crustal amplification factor is the
525 crustal rheology. At first order, the velocity imposed to the model does not influence
526 significantly the amplification factor. Investigating the interactions between these parameters
527 and their combined impact on strain rate concentration and amplification factor will require
528 further dedicated models.

529

530 **6. Discussion**

531

532 ***6.1 Impact of weakened areas on intraplate strain rates and seismicity levels***

533

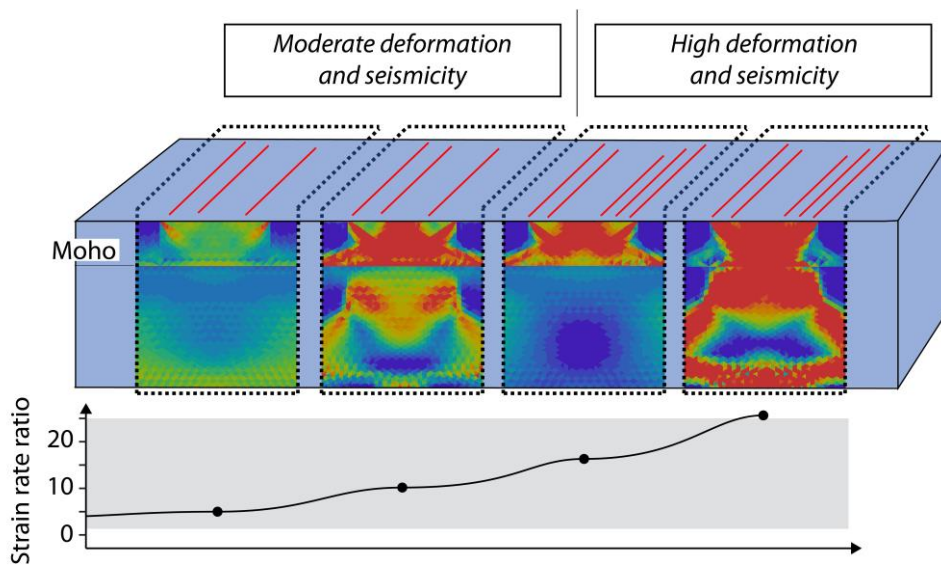
534 On the basis of the numerical models and the parametric tests, we propose a
535 conceptual model that relates, to first order, the structural inheritance with present-day strain
536 rate and seismicity concentration in intraplate deformation zones (Fig. 9). The main objective
537 is to present the possible variations of lithospheric structure linked to high or moderate strain
538 rate concentration. Because earthquakes are not directly modeled in our study, we make the
539 simple assumption that seismicity levels can be directly related to strain rate concentrations.

540 Moderate strain rate amplification factors (ca. 4 – 10) and seismicity levels may be
541 associated with a high inherited weakening ($\varepsilon > 1$) in the crust only or in the whole
542 lithosphere, a moderate or high net driving force ($6 - 10 \times 10^{12} \text{ N.m}^{-1}$), or a medium geotherm
543 ($500^\circ\text{C} \leq T_M < 600^\circ\text{C}$). Examples of this moderate case could be the Appalachian thrust
544 nappes, which may be associated with upper crust weakening but not lithospheric inheritance,
545 or whole lithosphere weakening and a cold geotherm (e.g., parts of the Iapetus rift close to the
546 Canadian Shield). In contrast, the Hercynian domain including the South Armorican Shear
547 Zone (western France) may have preserved lithospheric inheritance and could be explained by
548 moderate net driving force.

549 On the other hand, higher strain rate amplification factors up to 15 – 30, and thus
550 potentially higher seismicity levels, can be reached in domains of crust weakening associated
551 with high net driving force ($10 \times 10^{12} \text{ N.m}^{-1}$) or in domains of whole lithosphere weakening
552 and mild geotherms ($T_M = 600^\circ\text{C}$), as shown Fig. 9. This may be the case of specific regions
553 of the Iapetus Rift (e.g., St Lawrence Valley, New Madrid seismic zone).

554 The structure of the lithosphere (i.e. the presence of structural inheritance and the
555 associated weakened rheology) is representative of a long-term state. The first-order

556 explanation linking the presence of seismicity with models of different lithospheric structures
 557 (Fig. 9) assumes that the seismicity is also representative of a long-term behavior. This raises
 558 the question of seismic concentration as long-term or transient (temporal clusters). Long-term
 559 seismic concentration could be attributed to lithospheric structures, whereas transient seismic
 560 concentration could involve other processes localizing the strain rates. To address this issue,
 561 we compare modeled and observed strain rate amplification factors in the following section.



562
 563 **Figure 9. Conceptual model relating structural inheritance with upper crustal strain rate**
 564 **and seismicity concentration in intraplate deformation zones.** Red lines show schematic fault
 565 traces. Modeling results represent from left to right: UCW model in Fig. 6, model of lithospheric
 566 weakening with $F = 10 \times 10^{12} \text{ N.m}^{-1}$, model of crustal weakening with $F = 10 \times 10^{12} \text{ N.m}^{-1}$ and
 567 model of lithospheric weakening with Moho Temperature of 600°C . Lower curve shows
 568 variations of upper crustal strain rate amplification factor for each model. Amplification factors of
 569 4-10 and 15-30 are representative of moderate and high deformation and seismicity, respectively.
 570 Grey shaded area represents seismic and GPS strain rate amplification factors observed at large
 571 spatial scale (cf. Table 1).

572

573 **6.2 Comparison between modeled and observed strain rate amplification factors**

574

575 Intraplate strain rates, more particularly in non-weakened areas, are challenging to
576 measure because of their low magnitude. Estimations of strain rate amplification factors (i.e.
577 weakened over non-weakened strain rate ratios) can be made in the central and eastern North
578 America using published GPS and seismicity data (assuming that the seismic catalog is
579 representative of a long-term strain rate). Table 1 presents these amplification factors
580 calculated for regions of large (several hundred kilometers) and small (50-100 km) spatial
581 scales. The large spatial scale strain rates are calculated in (i.e. weakened area) and around
582 (i.e. non-weakened area) the Saint Lawrence Valley. The regions IRM, LAB and COC in
583 Table 1 are three subsections of the St Lawrence Valley (see Mazzotti and Adams, 2005).
584 Amplification factors range from 2 to 25 with a good coherence between those calculated by
585 GPS and seismicity. The smaller spatial scale strain rates are calculated in specific seismically
586 active areas: New-Madrid, Charlevoix, Lower St Lawrence (BSL) and Montréal. GPS
587 amplification factors range from 12 to 200. Seismic amplification factors range from 275 to
588 7000.

589 We obtain a reasonable agreement between modeled strain rate amplification factors
590 and large scale observed amplification factors (roughly factors of 5 – 30). If the ergodic
591 hypothesis is verified (i.e. the system has the same behavior averaged over time and averaged
592 over space), strain rates calculated on high spatial scale are representative of a long-term
593 deformation. Modeled strain rate amplification factors should be representative of a long-term
594 deformation and seismicity level. In this framework, the local seismic zones of Charlevoix,
595 Montréal and New Madrid could be associated with temporal clusters of seismicity.

596 However, a direct comparison between the modeled and local (small-scale) observed
597 strain rate amplification factors is not easy to make. A first explanation of the discrepancy is
598 that our models lack the complexity to be compared with natural cases. Secondly, although

599 the differences between large and small-scale amplification factors are significant, significant
600 uncertainties remain. Main uncertainties on observed strain rates are: (1) those specific to the
601 strain rate calculation method (see references in Table 1); (2) the differences between large
602 and small-scale amplification factors are significant for seismic amplification factors but not
603 for GPS amplification factors; (3) the background in Table 1 (i.e. the non-weakened zone) is
604 not the same for all calculated ratios. To address this issue, numerical models representing
605 each specific area are required. Complexity in the lithospheric structure and local processes
606 should be considered.

Region	Seismic strain rate ratio	GPS strain rate ratio
<u>Large scale region:</u>		
SLV / background	25 ^a	2 - 11 ^c
IRM / background	24 ^b	
LAB / background	10 ^b	
COC / background	5 ^b	
<u>Small scale region:</u>		
New Madrid / background	7000 ^d	
Charlevoix / background	6350 ^b	12 ^c - 200 ^e
Montréal / background	277 ^b	13 ^c
BSL / background	275 ^b	

607
608 **Table 1. Strain rate amplification factors from seismicity and GPS observations**
609 (weakened area over non-weakened area strain rate ratios). Large and small scale regions are
610 about 500 – 1000 and 10s – 100s km scale, respectively. SLV: Saint Lawrence Valley, IRM:
611 Iapetus Rift Margin, BSL: Bas Saint Laurent, LAB: southern LABrador and COC:
612 COchrane. All these regions are situated along the St Lawrence Valley (Eastern Canada). a:
613 Mazzotti and Gueydan (2017) and references therein, b: Mazzotti and Adams (2005), d:
614 Anderson (1986), c: Tarayoun et al. (2018) and e: Mazzotti et al. (2005).

615
616 **7. Conclusion**

617

618 The role of structural inheritance, proposed as a strain concentrator, is a key element to
619 understand the current strain rate and seismicity concentration in intraplate deformation
620 zones. In this study, we quantified the impact of the structural inheritance (i.e., presence of
621 large paleo-tectonic structures), through 2D numerical modeling of weakened domains at
622 different locations in the lithosphere. More specifically, we have quantified the amplification
623 factor of upper crustal strain rates associated with structural inheritance. Our analysis yields
624 three main conclusions:

625 (1) Lithospheric structural inheritance has a major impact on the concentration and
626 amplification of upper crustal strain rates. Amplification factors range from 1 to 27,
627 depending on the assumed rheology, geotherm, net driving force, and amount of inherited
628 weakening (Fig. 8). High upper crustal deformation is accentuated with a weak rheology, a
629 high amount of weakening (i.e. a high inherited finite strain), a high net driving force and a
630 mild geotherm.

631 (2) The concentration of upper crustal strain rate varies strongly depending on the
632 location of the weakened area in the lithosphere. Weakened zones with the highest impact are
633 the entire crust and the whole lithosphere for a Moho temperature at 500°C and 600°C,
634 respectively. Lithospheric mantle weakening has no impact for a cold geotherm ($T_M = 500^\circ\text{C}$)
635 and only accentuates the upper crustal deformation very slightly at milder geotherm ($T_M =$
636 600°C).

637 (3) Modeled strain rate amplification factors are in reasonable agreement with those
638 calculated from GPS and seismicity data at large spatial scales (several 100s km), thus
639 potentially representative of a long-term deformation (Table 1).

640 A major feature of our models is the presence of preserved elastic layer in the upper
641 lithospheric mantle for low Moho Temperature (i.e. 500°C). This elastic layer has a strong

642 impact on upper crustal strain rate concentration in a way that it tends to prevent high
643 amplification factors. Our innovative modeling approach, coupling velocity boundary
644 conditions and net force constraints, allows highlighting the presence of this preserved elastic
645 layer, with significant impact on the mechanical behavior of the lithosphere and potentially, in
646 the long term, on seismicity and seismic hazard characterization in intraplate deformation
647 zones.

648

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656

657 **References**

658

- 659 Adams, J., & Basham, P.W. (1991). The seismicity and seismotectonics of eastern Canada. In
660 D. B. Slemmons et al., *Neotectonics of North America, Decade Map* (Vol. 1, pp. 261-
661 276), Geological Society of America Boulder, Colorado.
- 662 Anderson, J. G. (1986). Seismic strain rates in the central and eastern United States, *Bulletin*
663 *of the Seismological Society of America*, 76, 273-290.
- 664 Boneh, Y., Wallis, D., Hansen, L. N., Krawczynski, M. J., & Skemer P. (2017). Oriented
665 grain growth and modification of ‘frozen anisotropy’ in the lithospheric mantle, *Earth*

666 *and Planetary Science Letters*, 474, 368-374.
667 <https://doi.org/10.1016/j.epsl.2017.06.050>

668 Byerlee, J. (1978). Friction of Rocks. In J.D. Byerlee, & M. Wyss, (Eds), *Rock Friction and*
669 *Earthquake Prediction* (pp. 615-626). Contributions to Current Research in
670 Geophysics (CCRG), vol 6. Birkhäuser, Basel. [https://doi.org/10.1007/j978-3-0348-](https://doi.org/10.1007/j978-3-0348-7182-2_4)
671 [7182-2_4](https://doi.org/10.1007/j978-3-0348-7182-2_4)

672 Calais, E., Freed, A. M., Van Arsdale, R., & Stein, S. (2010). Triggering of New Madrid
673 seismicity by late-Pleistocene erosion, *Nature*, 466, 608-611.
674 <https://doi.org/10.1038/nature09258>

675 Chéry, J., Zoback, M. D., & Hassani, R. (2001). An integrated mechanical model of the San
676 Andreas fault in central and northern California, *Journal of Geophysical Research*,
677 106, 22,051-22,066. <https://doi.org/10.1029/2001JB000382>

678 Copley, A., Avouac, J.-P., Royer, J.-Y. (2010). India-Asia collision and the Cenozoic
679 slowdown of the Indian plate: implications for the forces driving plate motions,
680 *Journal of Geophysical Research*, 115, 1-14. <https://doi.org/10.1029/2009JB006634>

681 Coppersmith, K. J., Johnston, A. C., Metzger, A. G., & Arabasz, W. J. (1987). Methods for
682 assessing maximum earthquakes in the central and eastern United States, Electric
683 Power Research Institute, working report EPRI RP2556-12, pp. 312.

684 Forsyth, D., & Uyeda, S. (1975). On the relative importance of the driving forces of plate
685 motion, *Geophysical Journal International*, 43, 163-200.
686 <https://doi.org/10.1111/j.1365-246X.1975.tb00631.x>

687 Gorczyk, W., Hobbs, B., & Gerya T. (2012). Initiation of Rayleigh-Taylor instabilities in
688 intra-cratonic settings, *Tectonophysics*, 514-517.
689 <https://doi.org/10.1016/j.tecto.2011.10.016>.

690 Grollmund, B., & Zoback, M. D. (2001). Did deglaciation trigger intraplate seismicity in the
691 New Madrid seismic zone ?, *Geology*, 29, 175-178. [https://doi.org/10.1130/0091-](https://doi.org/10.1130/0091-7613(2001)029<0175:DDTISI>2.0.CO;2)
692 [7613\(2001\)029<0175:DDTISI>2.0.CO;2](https://doi.org/10.1130/0091-7613(2001)029<0175:DDTISI>2.0.CO;2)

693 Gueydan, F., Leroy, Y. M., Jolivet, L., & Agard, P. (2003). Analysis of continental midcrustal
694 strain localization induced by microfracturing and reaction-softening, *Journal of*
695 *Geophysical Research*, 108, 175-178. <https://doi.org/10.1029/2001JB000611>

696 Gueydan, F., Morency, C., & Brun, J-P. (2008). Continental rifting as a function of
697 lithosphere mantle strength, *Tectonophysics*, 460, 83-93.
698 <https://doi.org/10.1016/j.tecto.2008.08.012>

699 Gueydan, F., Précigout, J., & Montési, L. G. J. (2014). Strain weakening enables continental
700 plate tectonics, *Tectonophysics*, 631, 189-196.
701 <https://doi.org/10.1016/j.tecto.2014.02.005>

702 Hassani, R., Jongmans, D., & Chéry, J. (1997). Study of plate deformation and stress in
703 subduction processes using two-dimensional numerical models, *Journal of*
704 *Geophysical Research*, 102, 17951-17965.

705 Hirth, G., & Kohlstedt, D. (2003). Rheology of the uppermantle and the mantle wedge : a
706 view from the experimentalists. In J. Eiler, (Ed), *Inside the Subduction Factory* (138,
707 pp. 83-105).

708 Holdsworth, R. E. (2004). Weak faults – rotten cores, *Science*, 303, 181-182.
709 <https://doi.org/10.1126/science.1092491>

710 Illies, J. H. (1972). The Rhine graben rift system-plate tectonics and transform faulting,
711 *Geophysical Surveys*, 1, 27-60.

712 Jégouzo, P. (1980). The South Armorican Shear Zone, *Journal of Structural Geology*, 2, 39-
713 47.

- 714 Johnston, A. C. (1989). The seismicity of “stable continental interiors”. In S. Gregersen, & P.
715 W. Basham, (Eds), *Earthquakes at North-Atlantic Passive Margins: Neotectonics and*
716 *Postglacial Rebound* (pp.299-327). Kluwer Academic Publishers, Dordrecht, The
717 Netherlands.
- 718 Kenner, S. J., & Segall, P. (2000). A Mechanical Model for Intraplate Earthquakes :
719 Application to the New Madrid Seismic Zone, *Science*, 289, 2329-2332.
720 <https://doi.org/10.1126/science.289.5488.2329>
- 721 Kumarapeli, P. S., & Saull, V. A. (1966). The St. Lawrence Valley system: a North American
722 equivalent of the East African Rift Valley system, *Canadian Journal of Earth*
723 *Sciences*, Vol. 3.
- 724 Kuszniir, N. J. (1991). The distribution of stress with depth in the lithosphere : thermo-
725 rheological and geodynamic constraints. In R. B. Whitmarsh, (Ed), *Tectonic Stress in*
726 *the Lithosphere* (pp. 95-107). Royal Society, London.
- 727 Luan, F. C., & Paterson, M. S. (1992). Preparation and deformation of synthetic aggregates of
728 quartz, *Journal of Geophysical Research*, 97, 301-320.
729 <https://doi.org/10.1029/91JB01748>
- 730 Mareschal, J. C., Jaupart, C., Gariépy, C., Cheng, L. Z., Guillou-Frottier, L., Bienfait, G., &
731 Lapointe, R. (2000). Heat flow and deep thermal structure near the southeastern edge
732 of the Canadian Shield, *Canadian Journal of Earth Sciences*, 37, 399-414.
- 733 Mazzotti, S., & Adams, J. (2005). Rates and uncertainties on seismic moment and
734 deformation in eastern Canada. *Journal of Geophysical Research: Solid Earth*, 110,
735 B09301. <https://doi.org/10.1029/2004JB003510>
- 736 Mazzotti, S., James, T. S., Henton, J., & Adams, J. (2005). GPS crustal strain, postglacial
737 rebound, and seismic hazard in eastern North America: The Saint Lawrence valley

738 example. *Journal of Geophysical Research*, 110, B11301.
739 <https://doi.org/10.1029/2004JB003590>

740 Mazzotti, S., & Gueydan, F. (2017). Control of tectonic inheritance on continental intraplate
741 strain rate and seismicity, *Tectonophysics*. <https://doi.org/10.1016/j.tecto.2017.12.014>

742 Mooney, W. D., Laske, G., & Masters, G. (1998). CRUST 5.1 : A global crustal model at 5° x
743 5°, *Journal of Geophysical Research*, 103, 727-747.

744 Musacchio, G., Mooney, W. D., Luetgert, J. H., & Christensen, N. I. (1997). Composition of
745 the crust in the Grenville and Appalachian Provinces of North America inferred from
746 Vp/Vs ratios, *Journal of Geophysical Research*, 102, 15,225-15,241.
747 <https://doi.org/10.1029/96JB03737>

748 Owen, D. R. J., & Hinton, E. (1980). *Finite Element in Plasticity*. Pineridge Press Limited.

749 Précigout, J., & Gueydan, F. (2009). Mantle weakening and strain localization: Implications
750 for the long-term strength of the continental lithosphere, *Geology*, 37, 147-150.
751 <https://doi.org/10.1130/G25239A.1>

752 Regenauer-Lieb, K., & Yuen, D. A. (2003). Modeling shear zones in geological and planetary
753 sciences : solid- and fluid-thermal-mechanical approaches, *Earth-Science Reviews*, 63,
754 295-349. [https://doi.org/10.1016/S0012-8252\(03\)00038-2](https://doi.org/10.1016/S0012-8252(03)00038-2)

755 Schueller, S., Gueydan, F., & Davy, P. (2005). Brittle-ductile coupling: Role of ductile
756 viscosity on brittle fracturing, *Geophysical Research Letters*, 32.
757 <https://doi.org/10.1029/2004GL022272>

758 Schueller, S., Gueydan, F., & Davy, P. (2010). Mechanics of the transition from localized to
759 distributed fracturing in layered brittle-ductile systems, *Tectonophysics*, 484, 48-59.
760 <https://doi.org/10.1016/j.tecto.2009.09.008>

761 Schulte, S. M., & Mooney, W. D. (2005). An updated global earthquake catalogue for stable
762 continental regions: reassessing the correlation with ancient rifts, *Geophysical Journal*
763 *International*, 161(3), 707–721, <https://doi.org/10.1111/j.1365-246X.2005.02554.x>

764 Stein, S., & Mazzotti, S. (2007). Continental Intraplate Earthquakes: Science, Hazard, and
765 Policy Issues. The Geological Society of America, Special paper 425.

766 Sykes, L. R. (1978). Intraplate seismicity, reactivation of preexisting zones of weakness,
767 alkaline magmatism, and other tectonism postdating continental fragmentation,
768 *Reviews of Geophysics and Space Physics*, 4, 621-688.
769 <https://doi.org/10.1029/RG016i004p00621>

770 Tarayoun, A., Mazzotti, S., Craymer, M., & Henton, J. (2018). Structural inheritance control
771 on intraplate present-day deformation: GPS strain rate variations in the Saint
772 Lawrence Valley, eastern Canada, *Journal of Geophysical Research: Solid Earth*, in
773 press. <https://doi.org/10.1029/2017JB015417>

774 Thielmann, M., & Kaus, B. J. P. (2012). Shear heating induced lithospheric-scale localization:
775 Does it result in subduction?, *Earth and Planetary Science Letters*, 359-360, 1-13.
776 <https://doi.org/10.1016/j.epsl.2012.10.002>

777 Thomas, W. A. (2006). Tectonic inheritance at a continental margin. *GSA Today*, 16, 4-11.
778 [https://doi.org/10.1130/1052-5173\(2006\)016\[4:TIAACM\]2.0.CO;2](https://doi.org/10.1130/1052-5173(2006)016[4:TIAACM]2.0.CO;2)

779 Tommasi, A., Knoll, M., Vauchez, A., Signorelli, J. W., Thoraval, C., & Logé, R. (2009).
780 Structural reactivation in plate tectonics controlled by olivine crystal anisotropy,
781 *Nature Geoscience* 2, 423-429. <https://doi.org/10.1038/ngeo528>

782 Weertman, J. (1978). Creep laws for the mantle of the Earth, *Philosophical Transactions of*
783 *the Royal Society London*, A 288, 9-26

- 784 Wilks, K. R., & Carter, N. L. (1990). Rheology of some continental lower crustal rocks,
785 *Tectonophysics*, 182, 57-77. [https://doi.org/10.1016/0040-1951\(90\)90342-6](https://doi.org/10.1016/0040-1951(90)90342-6)
- 786 Wintsch, R. P., Christoffersen, R., & Kronenberg, A. K. (1995). Fluid-rock reaction
787 weakening of fault zones, *Journal of Geophysical Research*, 100, 13021-13032.
788 <https://doi.org/10.1029/94JB02622>
- 789 Wu, P., & Johnston, P. (2000). Can deglaciation trigger earthquakes in N. America ?,
790 *Geophysical Research Letters*, 27, 1323-1326. <https://doi.org/10.1029/1999GL011070>
- 791 Wu, P., & Mazzotti, S. (2007). Effects of a lithospheric weak zone on postglacial
792 seismotectonics in eastern Canada and the northeastern United States. In S. Stein, & S.
793 Mazzotti, (Eds), *Continental Intraplate Earthquakes: Science, Hazard, and Policy*
794 *Issues. The Geological Society of America*, Special paper 425 (pp. 113-128).
795 [https://doi.org/10.1130/2007.2425\(09\)](https://doi.org/10.1130/2007.2425(09))
- 796 Zoback, M. D., & Townend, J. (2001). Implications of hydrostatic pore pressures and high
797 crustal strength for the deformation of intraplate lithosphere, *Tectonophysics*, 336, 19-
798 30. [https://doi.org/10.1016/S0040-1951\(01\)00091-9](https://doi.org/10.1016/S0040-1951(01)00091-9)